A Mesoscale Modeling Study of the 1996 Saguenay Flood

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ABSTRACT

A mesoscale simulation of the 19–21 July 1996 Saguenay flood cyclone was performed using the Canadian Mesoscale Compressible Community (MC2) model to study the processes leading to the explosive development and the large amount of precipitation. The performance of the simulation is verified by careful comparison with available observations with particular emphasis on the quantitative forecast of precipitation. It was shown that the model accurately simulates the wind, temperature, and humidity fields. Using the Kong and Yau microphysics scheme, the model performs quite well in the threat scores over a broad range of precipitation thresholds. Comparison of model precipitation against an objective analysis from rain gauge measurements and against the time evolution of accumulated precipitation at specific sites indicates generally good agreement except that the magnitude of the maxima is about 10% lower in the simulation.

Potential vorticity (PV) inversion and sensitivity experiments show that the rapid deepening of the cyclone results from a combination of upper-level forcing from two shortwave troughs that partially merge, an upper-level jet streak, latent heat release, and low-level thermal advection. Condensational heating was integral for the establishment of a phase lock between the surface cyclone and a strong, upper-level trough that steers the cyclone. The flow field associated with a weaker trough, located downstream of the stronger trough, acted to retard the progression of the stronger trough, ultimately causing the cyclone to be located in a favorable position to interact with orography. It was shown that in the middle of the explosive deepening period, the contributions to the 900-hPa geopotential height anomaly from the upper-level dry PV anomaly, the low-level moist PV anomaly, and the surface potential temperature anomaly were 47%, 41%, and 12%, respectively.

The contribution to the precipitation from orographic variation is quantified through sensitivity experiments in which aspects of the orography field are altered in the model conditions. It was found that orographic variation contributed to approximately 15% of the 48-h accumulated precipitation in the region of the flooding and up to over 25% in other local areas.

1. Introduction

A severe flood occurred in the Saguenay region1 of Quebec, Canada, on 19–21 July 1996. The maximum accumulated precipitation is found in the Saguenay valley, which is surrounded by mountains on its north, southwest, and southeast sides. There are two man-made reservoirs, Lake Kenogami and Lake Ha!Ha!, with ten dikes and three dams. Two weeks before the flood, the eastern half of North America was dominated by a long-wave trough and the inclement weather resulted in saturation of the soil in the Saguenay region. Rain started to fall around 1200 UTC 19 July. Although the average rainfall rate over the area is only $8 \text{ mm h}^{-1}$, the persistent precipitation led to a large accumulation. One automatic rain gauge near Lake Kenogami recorded 279 mm of accumulated rain, most of which fell in a 36-h period from 1200 UTC 19 July to 0000 UTC 21 July.

On the morning of 20 July, 28 h after rain started to fall at Lake Ha!Ha!, water ruptured an earthen dike and dug a trench through the forest before joining the Ha!Ha! River to swamp the communities downstream. Similarly, water began accumulating in the nearly full Lake Kenogami on the morning of 19 July. By the next morning, water poured out of the drainage basin and flooded properties downstream. The event represents the most catastrophic flood in Canadian history, with 10 deaths and 800 million dollars (Canadian) in property damage (Grescoe 1997). Heavy rainfall warnings were issued by the Quebec Weather Centre. However, the Canadian operational models initialized at 0000 UTC 19 July predicted the bulk of the precipitation too far east. The forecasted location of rainfall was more accurate when the models were initialized 12 h later but the forecast still underpredicted the 48-h accumulated precipitation beginning at 1200 UTC by 50% (Verret et al. 1996).

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1 The four major communities of the region are Alma, Jonquière, Chicoutimi, and La Baie, which are located along the southern bank of the Saguenay River, approximately 150 km north of Quebec City.

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The fact that precipitation occurred on relatively large temporal and spatial scales indicated the feasibility of using high resolution mesoscale modeling initiated by operational analysis to investigate the synoptic and mesoscale processes that produced the persistent precipitation. Because of the large accumulated rainfall, latent heat release is likely to have played a role. It is also desirable to investigate the importance of the mountains surrounding the Saguenay Valley. Thus, it is the goal of this work to understand the basic processes in the Saguenay flood using the Canadian Mesoscale Compressible Community model (MC2; see Benoit et al. 1997). Specifically, we will seek answers to the following questions:

- What physical processes produced the large amount of precipitation?
- Why did the rainfall last for 36 h?
- What is the effect of the latent heat release?

The organization of this paper is as follows. A synoptic overview is presented in section 2. Section 3 describes briefly the model and the experimental setup. The control experiment and its validation form the subject of section 4. Section 5 contains diagnostic studies and sensitivity experiments using the potential vorticity (PV) inversion technique to clarify the questions posed above. Concluding remarks are given in section 6.

2. Synoptic overview

A low pressure system with a central sea level pressure (SLP) of 1002 hPa originated over southern Manitoba and deepened gradually as it moved eastward. By 1200 UTC 19 July (hereafter, denoted as date/time or date/time-model hour), the Canadian Meteorological Centre (CMC) regional analysis depicts a central SLP of 999 hPa and the system reached just north of Lake Ontario (Fig. 1a). The cyclone then deepened 20 hPa in 24 h as it moved north-eastward toward the Gaspe Peninsula. The system reached just north of Lake Ontario (Fig. 1b). The surface low is located directly under one 500-hPa trough and downstream of another. The cyclone entered the middle of its explosive deepening period at 20/0000. The surface low is located at the Maine and Quebec border with the warm front immediately to its north and west (Fig. 1c). Thermal advection increases in the lower troposphere. The 500-hPa troughs had come together and partially merged to form a closed circulation (Fig. 1d). At 20/1200 the cyclone is near Campbellton, New Brunswick. It is at its deepest stage, with a central SLP of 979 hPa. Strong thermal advection is indicated by strong low-level winds (in the form of a low-level jet) crossing the isotherms at a significant angle (Fig. 1e). Similar to the finding of Uccellini (1990), cyclogenesis of the Saguenay storm is marked by a low-level thermal field that evolves into an S-shape pattern during the rapid development stage. Comparison of Figs. 1e and 1f indicates that the system has become equivalent barotropic, with closed circulations as high as 250 hPa (not shown). For the following 24 h, the cyclone gradually filled at a rate of about 3 hPa (6 h)⁻¹ but remained quasi-stationary over the Gaspé Peninsula.

The observed 48-h maximum accumulated precipitation beginning at 19/1200 at Rivière-Aux-Écorces, in the Saguenay region of eastern Quebec, was 274 mm. A secondary precipitation maximum of 189 mm was found near Les Buissons, about 200 km northeast of the Saguenay region. The average 48-h precipitation was about 200 mm over an area of 5000 km². The period of most intense rainfall coincides with the period of explosive deepening of the cyclone (19/1800–20/0600).

3. Model description and design of experiment

The simulations were performed using the limited-area mesoscale model MC2 (Robert et al. 1985; Tanguay et al. 1990; Benoit et al. 1997). It is based on the fully compressible Euler equations, solved on a polar stereographic projection (true at 60°N) using the semi-implicit and semi-Lagrangian algorithm. The reference state used is an isothermal hydrostatic atmosphere at rest. The prognostic variables are the three velocity components \(u, v, w\); logarithm of a dimensionless perturbation pressure \(P\) from the reference state \(\ln(p/p_o)\) with \(p_o = 1000\) hPa; temperature \(T\); and the mixing ratios for water substances. The space derivatives are discretized by finite differences on a grid staggered in three dimensions, with an Arakawa C grid for the horizontal and a Tokioka B grid for the vertical. Orography is introduced by the use of the terrain-following Gal-Chen vertical coordinate, modified to allow for compressing or stretching the coordinate in the vertical.

The model has a comprehensive physics package (Mailhot et al. 1997). It includes planetary boundary layer processes based on turbulent kinetic energy (Benoit et al. 1989), fully implicit vertical diffusion, and a stratified surface layer based on similarity theory. The surface temperature over land is predicted via the force restore method (Dreearoff 1978; Benoit et al. 1989). The diurnal cycle associated with solar and infrared fluxes over ground is modulated by diagnostic clouds. The solar (Fouquart and Bonnel 1980) and infrared (Garand and Mailhot 1990) schemes in the radiation package of the model are fully interactive with the clouds. The total precipitation is the sum of the convective and stratiform precipitation. The former is generated from a cumulus parameterization and a choice could be made between the Kuo (1974) scheme and the Kain and Fritsch (1993) scheme. Stratiform precipitation is produced either by the Sundqvist et al. (1989) scheme, with
Fig. 1. Canadian Meteorological Centre (CMC) regional analyses for (a) and (b) 1200 UTC 19 Jul, (c) and (d) 0000 UTC 20 Jul, and (e) and (f) 1200 UTC 20 Jul. (left panels) SLP (solid contours every 4 hPa) and 850-hPa temperature (dashed contours every 2°C). (right panels) 500-hPa geopotential height (solid contours every 6 hPa) and absolute vorticity (dashed contours every $2 \times 10^{-5}$ s$^{-1}$, shaded above $16 \times 10^{-5}$ s$^{-1}$). Subjectively drawn trough lines are shown in (b).
Infrared radiation scheme Incorporates cloud interaction and radiation run as compared to CONT, the improved precipitation scale features are equally well simulated in the 35-km within the domain at initial time. Although the large-entire surface cyclone and two midlevel troughs are 20-km grid-spacing domain delineated in Fig. 1a. The best 48-h precipitation accumulation. The specific pa-

| Table 1. Summary of MC2 parameters in control simulation. |
|-------------------------------|------------------|
| Projection                    | Polar stereographic, true at 60°N |
| Horizontal grid               | 180 × 120, 20-km resolution |
| Number of vertical levels     | 25 (Gal-Chen levels) |
| Model top                     | 25 000 m          |
| Time step                     | 150 s             |
| Grid-scale condensation       | Explicit (cold) microphysics (Kong and Yau 1997) |
| Convective parameterization   | Kuo (Mailhot 1994) |
| PBL scheme                    | Based on predictive equation for turbulent kinetic energy (Benoit et al. 1989) |
| Solar radiation scheme        | Accounts for H$_2$O, CO$_2$, O$_3$, and cloud effects (Fouquart and Bonnel 1980). |
| Infrared radiation scheme     | Incorporates cloud interaction and radiation effects of H$_2$O, CO$_2$, and O$_3$ (Garrand and Mailhot 1990) |

As mentioned earlier, the CONT experiment is selected from four simulations with different resolved-scale and subgrid-scale precipitation parameterization schemes (Table 2). In terms of PR48, a comparison of Figs. 2 and 4 shows that with the exception of the Kain-
Fig. 2. Objective 48-h accumulated precipitation analysis for the period between 1300 UTC 19 Jul and 1300 UTC 21 Jul (contour and shading interval is 25 mm), central SLP tracks from CONT (solid lines) and CMC regional analyses (dashed lines). Central SLP locations are marked every 6 h and labeled every 12 h. Inset is a plot of the central SLP values vs time for CONT (solid) and analyses (dashed).

Fritsch experiment (KFEXP), the other three runs successfully captured the local maximum in the Saguenay area. However, only the Kuo experiment (KUOEXP) yields an absolute maximum in the Saguenay region in accordance with the analysis depicted in Fig. 2. The maximum of 224 mm in KUOEXP is also closest to the analyzed maximum of 246 mm, which corresponds to the observed maximum of 274 mm. The underprediction in the model is partially related to the fact that the rainfall in a grid box represents a spatially averaged amount over an area of 400 km². Furthermore, a large spatial variability in precipitation occurred in the measurements, with differences as large as 14 mm in PR48 between two rain gauges less than 1 km apart (Yu et al. 1997). The precipitation band that extends from north of Prince Edward Island to south of Nova Scotia lies outside the range of available rain gauge measurements. However, satellite imagery between 20/0000 and 20/1800 (Environment Canada 1997) does indicate the passage of a squall line over that region.

To evaluate the QPF of the runs more objectively, we calculated the threat score (TS) and bias score (BS), defined by Anthes (1983) as

\[
TS = \frac{C}{F + R - C} \quad \text{and} \quad BS = \frac{F}{R},
\]

where \(C\) is the number of grid points correctly forecast to receive a threshold amount of precipitation, \(F\) is the number of grid points forecast to receive this amount, and \(R\) is the number of grid points where the threshold amount is observed. Alternatively, for a particular isohyet, \(F\) is the area enclosed by the forecast isohyet, \(R\) is the area enclosed by the observed isohyet, and \(C\) is the intersection of \(F\) and \(R\). Hence, TS is the ratio of the correctly forecast area to the sum of the area that was incorrectly forecast and the area missed by the forecast. The values for TS range from 0 to 1, with a higher value denoting a better forecast. Here BS is simply the ratio of the area forecast to receive a threshold amount of precipitation to the area that actually observed that amount. The BS can be any positive number, with scores above (below) 1 implying an overpredicted (underpredicted) area. It is possible to have
a perfect BS = 1, implying that the forecast area equals the observed area, but the two areas may not overlap. In such a case, the TS punishes the model forecast for incorrectly forecasting the location of the precipitation.

Since the interpolation scheme produces smoothing in the precipitation analysis, the TSs are computed directly from precipitation observations, rather than from the analysis shown in Fig. 2. As in Hamill (1999), observation points were assigned to their nearest grid box and for grid boxes that contained more than one observation point, the average observation value was taken. Since our interest is in the skill of the model runs in simulating the large precipitation values, we have included in the computation of TS and BS only those observation points whose 48-h accumulated precipitation values were 50 mm or more. Following Hamill (1999), we equalize the BSs to those of KUOEXP before calculating the TSs.

Figure 5 shows the “equalized” TSs for the four experiments over a range of PR48 thresholds as well as the BSs before equalizing them to the bias values of KUOEXP. For threshold values between 50 and 75 mm, KFEXP and EXPC have slightly higher TSs than KUOEXP and KOUSUND. For all threshold values over 100 mm, the TSs are significantly higher in KUOEXP, with the exception of a small interval between 175 and 185 mm where EXPC scored slightly better. As indicated by the bias scores, precipitation areas are underpredicted in all runs for threshold values below 105 mm. Above this threshold, KFEXP and KUOSUND strongly underpredicted the area of precipitation. The overall performance in the bias score is best for KUOEXP.

The KUOEXP run therefore outperforms the other runs in terms of quantitative precipitation simulation as indicated by the higher TSs over a broad range of thresh-
old values and a better overall precipitation pattern when compared to the objective analysis. There is some spurious convective precipitation, south of the main precipitation area, which is associated with the Kuo scheme (present in KUOEXP and KUOSUND but absent in EXPC, KFEXP, and the objective analysis), but it occurs early in the simulation (within the first 12 h) and is confined to an area sufficiently far from the Saguenay region so as not to contaminate the precipitation simulation in the area of interest. We chose the KUOEXP run as the control simulation for our diagnostics studies and hereafter refer to it as CONT.

It is recognized that the Kuo scheme uses moisture convergence as a closure assumption although convection is not caused directly by the convergence of moisture (Raymond and Emanuel 1993). This is especially the case when convection arises from boundary layer heating and moistening. However, in this particular case the use of the Kuo scheme in conjunction with an ex-
explicit microphysics scheme did produce a very good precipitation simulation. To examine why this is the case, we first partitioned the total accumulated precipitation (PR48) from the model into a stratiform component (produced by the explicit scheme) and a convective component (produced by the Kuo scheme). Figure 6 indicates that the explicit scheme and convective scheme contributed approximately equally to the main precipitation maximum in the Saguenay area. A majority of the convective precipitation fell between 19/1800-06 and 20/0000-15 (not shown). The secondary maximum to the north comprises mainly stratiform rain. Figure 7 shows the brightness temperature (from GOES-8) at approximately 20/0000-12 as well as the instantaneous convective and stratiform precipitation rates at the same time. Convection appears to be correctly simulated as the areas of convective precipitation overlap regions with brightness temperatures <235 K (−38°C). To determine if convection is initiated properly, we plotted the model sounding at the location of the black dot at 19/1200-00. It is clear that the atmospheric column is conditionally unstable. The lower inset in Fig. 7 shows a vertical cross section of upward motion (w) along the arrow at 19/1800-06, before the occurrence of convective precipitation in that area. A deep layer of ascent occurs above 900 hPa. This deep column ascent, accompanied by moisture convergence, initiates convection in a conditionally unstable atmosphere and accounts for the reasonable performance of the Kuo scheme in the present case.

To further substantiate the performance of the CONT run in simulating precipitation realistically, we plotted the evolution of the accumulated precipitation at specific locations in three regions depicted in Fig. 2: Saguenay,
Beauce, and Gaspé. It is clear that for each of the two rain gauge stations in each region (Fig. 8), CONT simulates reasonably accurately the precipitation rates at the correct times. The delay at Cap-Étérnité arises from the spinup of the model, as rain was measured at this station before the model initial time. Although the evolution of precipitation near Cap Madeleine is less satisfactory, three grid points (60 km) to the east the modeled rainfall time series is very similar to the observed series at Cap Madeleine, suggesting a slight displacement of the area of precipitation in the model.

5. PV inversion diagnostics and sensitivity tests

Ertel's (1942) PV, defined as

$$ q = \rho^{-1} \mathbf{\eta} \cdot \nabla \theta, $$

with $\rho$, $\mathbf{\eta}$, and $\theta$ being air density, absolute vorticity vector, and potential temperature respectively, is an attractive variable because it is a conserved quantity on an isentropic surface in the absence of diabatic and dissipative processes (Hoskins et al. 1985). Furthermore, the invertibility principle allows the mass and wind fields associated with any particular PV to be determined. Davis and Emanuel (1991) proposed a piecewise PV inversion technique that has been applied by various researchers (Davis 1992a,b; Davis et al. 1993; Balasubramanian and Yau 1994) to quantify the contribution of upper-level and lower-level processes on cyclogenesis. Recently, the technique has also been applied to alter the initial conditions of a simulation to shed light on the effect of including or excluding a certain feature in the initial state (Huo et al. 1999b). Here we applied
the technique to diagnose the effect of different physical processes.

Before performing the inversion, we must define a mean state and partition the total PV anomaly into individual PV anomalies. The mean state is calculated as a 6-day time average covering the period from 17/1200 to 23/1200 centered at the 48 h of the control simulation. The data used are the 6-hourly regional analyses from CMC (Chouinard et al. 1994). The averaging period corresponds approximately to one synoptic-scale wave period. The total PV anomaly is simply computed as the deviation from the mean.

Following Huo et al. (1999a), the total PV anomaly is partitioned to isolate the perturbations associated with the tropopause depression ($Q_d$), latent heat release in the lower troposphere ($Q_m$), and surface baroclinicity ($Q_u$). The perturbation $Q_d$ is defined as all positive PV anomalies in dry air (with a relative humidity less than 30%) between 800 and 200 hPa, inclusive. The perturbation $Q_m$ is the sum of all positive PV anomalies in moist air (with a relative humidity greater than 70%) between 900 and 500 hPa, inclusive. The relative humidity threshold of 70% is chosen to include PV that may have been created by condensational heating in saturated air but is advected out of the cloudy region. The surface potential temperature anomaly $Q_u$ can be treated as a surrogate for the surface PV anomaly associated with low-level baroclinicity. Because the lower-level interior PV is strongly influenced by surface fluxes, we define the effective lower boundary potential temperature anomaly as the sum of the 1000-hPa potential temperature anomaly and the PV anomaly at the first interior level (900 hPa), unless the latter is already included as part of $Q_m$.

Finally, the residual interior PV anomaly ($Q_r$) is defined as the remaining perturbation consisting mainly of negative PV values associated with the mid- and upper-level ridges. The four PV anomalies are summarized in Table 3.

From the piecewise PV inversion, we obtain the wind, temperature, and geopotential height fields associated with the individual PV anomalies. The inversion is performed every 6 h, from 19/1200 to 21/1200.

**Table 3. Partitioning of PV anomalies (RH denotes relative humidity).**

<table>
<thead>
<tr>
<th>Symbol</th>
<th>PV anomaly</th>
<th>Definition</th>
</tr>
</thead>
<tbody>
<tr>
<td>$Q_d$</td>
<td>Upper-level dry PV anomaly</td>
<td>Positive PV anomaly from 800 to 200 hPa; RH &lt; 30%</td>
</tr>
<tr>
<td>$Q_m$</td>
<td>Low-level moist PV anomaly</td>
<td>Positive PV anomaly from 900 to 500 hPa; RH &gt; 70%</td>
</tr>
<tr>
<td>$Q_u$</td>
<td>Bottom potential temperature anomaly</td>
<td>1000-hPa potential temperature anomaly and 900-hPa PV anomaly with RH &lt; 70%</td>
</tr>
<tr>
<td>$Q_r$</td>
<td>Residual PV anomaly</td>
<td>All remaining PV anomaly from 900 to 200 hPa (not part of $Q_d$, $Q_m$, or $Q_u$)</td>
</tr>
</tbody>
</table>

**a. Structure and evolution of PV anomalies in control simulation (CONT)**

Figures 9a and 9c depict, respectively, the 400-hPa total PV at 19/1800-06 and the 800-hPa total PV at 20/0000-12 during the initial rapid deepening phase of the storm. A thick line delineates regions where the relative humidity at 400 hPa exceeds 70% and can be regarded as the boundary for the cloudy region because of the sharp gradient of relative humidity at the cloud edge. The elongated pattern of PV at 400 hPa (Fig. 9a) is associated with two shortwave troughs shown in Fig. 1. The PV tongue is located in the unsaturated region, immediately to the west of the model-predicted cloud.
field enclosed by the 70% relative humidity contour. The vertical section at 19/1800-06 (Fig. 9b) indicates that the PV feature at 400 hPa results from the dry descent or intrusion of stratospheric air. Six hours later (Fig. 9d), the PV at the upper levels has descended further as the midlevel troughs continue to intensify. A patch of low-level PV, generated by the release of latent heat, can be identified in the cloudy region. Its magnitude increases with time as the cyclone deepens. Note that our choice of the relative humidity criterion for partitioning the various PV anomalies serves to separate, at a particular level, PV that originates at lower levels from PV that arises from stratospheric intrusion.

Figure 10 shows the upper-level dry PV anomaly $Q_d$ averaged from 600 to 200 hPa ($Q_d$), together with the 500-hPa geopotential height field. At 19/1200-00 (Fig. 10a), two distinct troughs can be identified. In terms of $Q_d$, the northern trough has values twice as large as the southern trough and it extends over a larger area. As time progresses, the troughs partially merge (Figs. 10b and 10c). The northern trough appears to intensify due to continued dry intrusion of stratospheric PV. On the other hand, the southern trough weakens and the associated $Q_d$ values become smaller than 0.3 PVU at 20/1800-30 (not shown) and its identity is lost at 21/0000-36 (Fig. 10d).

Figure 11 depicts the evolution of the lower-level PV anomaly, averaged from 900 to 500 hPa ($Q_m$), along with the 850-hPa height field. At 19/1200-00 (Fig. 11a), a positive PV anomaly is already present at the lower levels. As the cyclone deepens, $Q_m$ increases as a result of condensational heating (Fig. 11b). It reaches its high-
Fig. 10. Layer-averaged $Q_d$ (from 600 to 200 hPa) and 500-hPa geopotential height from CONT at (a) 19/1200-00, (b) 20/0000-12, (c) 20/1200-24, and (d) 21/0000-36. Contours and shading for $Q_d$ are 0.2, 1.0, and 2.0 PVU and contour interval for geopotential height is 6 dam. Location of cyclone center is indicated by black dot.

est value (1.64 PVU) at 20/0600-18 (not shown), 6 h before the cyclone achieves its lowest central pressure. Thereafter, $\hat{Q}_d$ decreases as the cyclone begins to fill (Figs. 11c and 11d).

b. Trough removal experiment

The Saguenay cyclone is characterized by the partial merging of two midlevel troughs during its period of rapid deepening. To determine the effect of trough merging, we performed a sensitivity experiment by excluding one of the troughs in the initial condition of CONT, following a similar methodology to Huo et al. (1999b). We first isolate the upper-level dry PV anomaly associated with the trough and then perform a PV inversion to obtain its associated wind and mass fields and subtract them from the original initial conditions. The new initial conditions are then used to perform the 48-h trough removal experiment (NOTR). We removed the southern trough because it has a smaller areal extent and associated $Q_d$ relative to the northern trough. The effect of removing the associated wind and mass fields at the lateral boundaries is small. Insets (a) and (b) of Fig. 12 show, respectively, the vertical cross sections of the total PV anomaly in the initial conditions for experiments CONT and NOTR along the arrow in Fig. 12. It is clear that the PV anomaly associated with the southern trough has been nearly completely removed from the initial conditions in NOTR.

The time series for the maximum value of $\hat{Q}_d$ for the northern trough in the two experiments are plotted in Fig. 13a. In general, the magnitude is larger in exper-
imement CONT than in NOTR. During the rapid deepening period before 24 h, the maximum value of $Q_m$ in both runs increases with time. The situation is however different for the low-level PV anomaly $Q_m$. During the initial spinup period (the first 6 h), the maximum value of $Q_m$ decreases in NOTR but increases slightly in CONT (Fig. 13b). After 19/1800-6, the time rate of increase is similar in the two experiments particularly during the period of maximum deepening. The maximum $Q_m$ value continues to increase in NOTR from 20/0600-18 to 20/1200-24 while it starts to decrease in CONT after 20/0600-18. Evidently, the absence of the southern trough decreases the upper-level forcing and delays the initial spinup of the cyclone at the low levels. Subsequently, the generation of low-level PV is delayed and the value of PV is smaller in NOTR.

Figure 14 displays the deepening rates and the tracks of the cyclones. The initial 4 hPa higher SLP in NOTR results from the removal of the southern trough from the initial conditions. The southern trough was initially located immediately upstream of the center of the cyclone (see Fig. 10a) and its removal is equivalent to the addition of mass. Note that the deepening rate in NOTR and CONT is similar, with the exception of the last 12 h when the cyclone fills more slowly in NOTR.

Inspection of the storm track indicates that the NOTR cyclone always moves to the east and north of the storm in CONT and with a faster speed. A similar behavior is also observed in the tracks of the minimum value of the geopotential height at 500 hPa in the two experiments (not shown). Since the storm track is steered by the flow at the upper levels, the displaced track in NOTR...
results directly from a slightly more eastward and faster propagation of the northern trough at the upper levels.

The reason for the faster progression of the upper northern trough in NOTR can be inferred from the 400-hPa wind field obtained by inverting the 400-hPa PV anomaly \( Q_d \) associated with the upper southern trough in CONT (Fig. 12). It is clear that the presence of the southern trough results in a reduction of wind speed in the vicinity of the northern trough, which therefore reduces the advection of positive PV and thus retards the progression of the northern trough. Another way of interpreting the effect of the southern trough on the progression of the northern trough is in terms of vortex–vortex interaction. When two upper-level positive PV anomalies are aligned north-to-south as in Fig. 12, the effect of the southern vortex is to produce cyclonic (anticyclonic) vorticity advection and geopotential height falls (rises) to the west (east) of the northern vortex, thus inducing a westward propagation [see Fig. 7 in Huo et al. (1999a)].

The displacement of the storm track results in a much different precipitation pattern in NOTR compared to CONT (Fig. 15). The secondary PR48 maximum immediately north of the Saguenay region was reduced somewhat in NOTR and shifted approximately 100 km eastward. Another local maximum, located approximately 400 km northeast of the Saguenay region in CONT, remains in the same location in NOTR but with a slightly higher value. The main precipitation maximum in the Saguenay region of CONT was shifted approximately 200 km eastward in NOTR and its value reduced by 15% (from 224 to 190 mm). As will be
shown below, this reduction is largely caused by the displaced storm track, which leads to smaller surface winds and orographic forcing in the Saguenay region in NOTR.

c. Dry-run experiment

Figure 13b indicates that low-level PV ($\overline{Q}_m$), associated with latent heat release, is generated rapidly during the deepening phase of the cyclone. To investigate the role of latent heating, we perform a DRY experiment by suppressing the release of latent heat of condensation in the thermodynamic equation. The evolution of the maximum-layer-averaged $Q_m$ in Fig. 13b shows that initially $\overline{Q}_m$ is nonzero. This is because the cyclone was present in the initial conditions. However, without latent heating, there is no low-level PV generation mechanism and the initial $\overline{Q}_m$ quickly diminishes. The evolution of the pattern of the layer-averaged $Q_m$ in DRY (not shown) is nearly the same as that of CONT (see Fig. 10) and the evolution of the maximum value in DRY is slightly less than in CONT. Figure 13a indicates that there is a slight reduction in the intensity of the northern trough due to reduced interaction between $\overline{Q}_m$ and $\overline{Q}_d$, but upper-level forcing is only slightly affected by the suppression of latent heating.

What is more noteworthy in DRY is the storm track and the deepening rate. As shown in Fig. 14, there is virtually no deepening of the central SLP in DRY and the track is very different, moving east rather than north-
east as in CONT for the first 24 h. To understand why this is so, we compare the SLP for CONT and DRY at various times (not shown). At 20/0000-12, there are two local minima of SLP in DRY. The northern minimum is located approximately in the same location as the central SLP in CONT at that time. Rather than deepening, the northern SLP minimum in DRY fills while the southern minimum deepens slightly and continues to exist throughout the integration. This process of coastal redevelopment does not take place in CONT. Instead, the latent heat release favors the development of the northern SLP minimum. We conclude that without the latent heat release in the first 12–18 h, the phase lock between the portion of $Q_d$ that can be initially identified as the midlevel northern trough and the surface cyclone is not established and the cyclone wanders hundreds of kilometers southwest of the track in CONT.

d. Contributions to cyclogenesis

To quantify the effects of the individual PV anomalies on the rapid deepening of the storm, we plotted in Fig. 16 the contributions of $Q_m$, $Q_d$, and $Q_u$ to the 900-hPa geopotential height fall averaged over a 150 km × 150 km area centered at the center of the cyclone at each time (the contribution from $Q_r$ is not included because it mainly results in a height rise). The results for a 450 km × 450 km area were nearly identical (not shown). The relative contribution is calculated as the ratio of the contribution from one of the anomalies to the sum of
the contributions from all three anomalies. In CONT, the contribution from $Q_d$ is the largest and its significance increases with time until 20/1800-30 due to stratospheric intrusion and the increasing vertical alignment of the upper northern trough and the surface cyclone. The contribution from $Q_m$ also indicates an increasing trend throughout the deepening phase. The effect of $Q_u$ is the smallest and it actually contributes to a height rise after 20/1200-24 as the cyclone center moves from a region of positive to a region of negative $Q_u$ anomaly.

At 20/0600-18, the relative contributions from $Q_d$, $Q_m$, and $Q_u$ to the 900-hPa height fall are 47%, 41%, and 12% (Fig. 16b).

Relative to CONT, the contribution from $Q_d$ is about 3–4 dam smaller in DRY (Fig. 16c). The absence of the upper southern trough slows down the spinup of the cyclone before 6 h. After 20/1800-30, the contribution from $Q_m$ in the two experiments becomes similar, with maximum values of about −19 dam in NOTR and −23 dam in CONT. The contributions from $Q_u$ are similar in the two runs because the removal of the upper southern trough does not affect the baroclinicity near the surface and the cyclone in NOTR moves into a region of negative $Q_u$ anomaly at approximately the same time as the storm in CONT.

Without latent heating, the contribution from $Q_m$ in DRY is much smaller (Fig. 16e) than in CONT (Fig. 16a). The height fall at the cyclone center associated with $Q_d$ is much reduced mainly because the track of the cyclone in DRY lies hundreds of kilometers south of that in CONT and the influence of the forcing from the upper northern trough significantly diminished. Ex-
cept in the initial 12 h, the contribution from $Q_u$ remains small. Most of the relative contribution arises from upper-level forcing (Fig. 16f).

e. Influence of jet streak

Figure 17 shows the wind vectors and isotachs at 300 hPa at 20/0000-12, when the cyclone is deepening explosively. Two jet streaks are present in the vicinity of the cyclone, one upstream and one downstream. (Note that although the two are essentially part of the same baroclinic system, we identify them as distinct jet streaks according to the strict definition of a jet streak as simply a local isotach maximum.) Cross sections of ageostrophic wind (not shown) indicate considerable enhancement of upward motion ($w$) due to a thermally direct circulation in the entrance region of the downstream jet streak. The inset in Fig. 17 depicts a maximum $w$ of 76 cm s$^{-1}$ at around 450 hPa in the right entrance region of the jet streak. The jet enhances cyclogenesis between 19/2100-09 and 20/0003-15, with the strongest influence at approximately 20/0000-12. Cross sections of ageostrophic flow indicate that no significant dynamic forcing for surface cyclogenesis appears to be associated with the upstream jet streak. However, the two jets may be too close to distinctly identify forcing from one particular jet compared to forcing from the other.

f. Effects of orography

Table 4 lists six sensitivity experiments for investigating the effects of orographic forcing. In each experiment, some feature of the orography is artificially removed and the effect on the prediction of precipitation...
TABLE 4. List of experiments. (“Different” refers to a difference in the maximum PR48 in the Saguenay region of at least 5 mm.)

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Orographic feature(s) modified</th>
<th>PR48 different from CONT</th>
</tr>
</thead>
<tbody>
<tr>
<td>NOR1</td>
<td>NOR, SW, SE, and VAL</td>
<td>Yes</td>
</tr>
<tr>
<td>NOR2</td>
<td>NOR</td>
<td>No</td>
</tr>
<tr>
<td>NOR3</td>
<td>SW</td>
<td>No</td>
</tr>
<tr>
<td>NOR4</td>
<td>SE</td>
<td>Slightly</td>
</tr>
<tr>
<td>NOR5</td>
<td>VAL</td>
<td>Yes</td>
</tr>
<tr>
<td>NOR6</td>
<td>SE and VAL</td>
<td>Yes</td>
</tr>
</tbody>
</table>

analyzed. Figure 18 shows the orography along the path of the cyclone. Important features include the Saguenay valley (VAL), and the mountain ranges to the north (NOR), southeast (SE), and southwest of the valley (SW). To modify a certain orographic feature, we simply reduce the height of a mountain to 400 m or fill up the valley to 400 m. Because the initial location of the cyclone and the moist air is downstream of any orographic feature modified and the air mass associated with the storm is advected into the Saguenay region, the initial conditions are not changed significantly.

In all the experiments, the track and the deepening rate of the simulated storm remained relatively unchanged. In the first experiment, NOR1, where all orographic features in the area of the precipitation maxima were modified, there is a significant reduction in PR48 in the Saguenay region. The main maximum of 224 mm in CONT was reduced to 159 mm in NOR1. The other local precipitation maximum to the north was however not changed significantly. In NOR2 and NOR3, the change in PR48 is small. In NOR4 and NOR5, the peak accumulation becomes 218 and 176 mm, respectively, at the location of the main maximum in CONT. The reductions of peak PR48 in the last two experiments motivate NOR6, which modifies both the Saguenay valley and the mountain to the southeast. The precipitation pattern in NOR6 (Fig. 15c) is essentially the same as NOR1 (not shown). The value of PR48 in the location of the Saguenay maximum in CONT is reduced to 165

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Fig. 18. Orography (contour and shading interval is 100 m), storm track (central SLP location indicated every 6-h), SLP (contour interval is 4 hPa), and surface wind vectors at 20/1200-24 from CONT. Location of total 48-h precipitation maximum indicated by X. VAL, NOR, SE, and SW indicate, respectively, the Saguenay valley, the mountains to the north of the valley, the mountain immediately southeast of the valley and the mountain immediately southwest of the valley. Insets are vertical cross sections of upward motion across arrow at 20/1200-24 for (top) CONT and (bottom) NOR6. Contour interval is 5 cm s⁻¹.
mm. Therefore, almost all the difference between NOR1 and CONT is attributable to the removal of the Saguenay valley and the southeast mountain peak.

Figure 18 also depicts the local orography and the surface wind in CONT at 20/1200-24 when the difference in the precipitation rates in the Saguenay region between CONT and NOR6 was the greatest. The insets compare the vertical motion (w) between CONT and NOR6 along the direction of the arrow which crosses the PR48 maximum in CONT. The maximum w-value over the Saguenay valley at 700 hPa is reduced from 0.33 m s\(^{-1}\) (CONT) to 0.22 m s\(^{-1}\) (NOR6), a reduction of 33%. At 400 hPa, the value decreases from 0.41 m s\(^{-1}\) in CONT to 0.36 m s\(^{-1}\) in NOR6, a reduction of slightly over 10%.

The removal of the Saguenay valley and the southeast mountain in NOR6 results in a significant reduction of upward motion and the precipitation rate. In an attempt to determine whether orographically forced vertical motion is directly responsible, we partitioned PR48 into the explicit and the convective components in CONT and NOR6 (Table 5). The contributions from these two components are approximately equal; both explicit and convective precipitation in NOR6 are reduced in areas where upward motion is reduced. It is possible that the increased precipitation in CONT results partly from enhanced upward motion due to upslope flow in that area. However, a comparison with NOR6 indicates that the orographic variation results in a larger column-integrated moisture convergence in CONT in the region between VAL and SE (not shown). Since the Kuo scheme responds directly to moisture convergence, it is also possible that the Kuo scheme was activated more often to enhance the convective precipitation in the presence of orography. The enhanced convection in turn increased the vertical motion leading to stronger explicit precipitation. Further sensitivity tests are required to pinpoint the exact mechanism for the enhancement of precipitation in the presence of orographic variation.

Table 6 lists the differences between PR48 in CONT and NOR6 over certain areas. The *area of PR48 Saguenay maximum* includes those grid points with PR48 > 180 mm in CONT. It covers an area of 5200 km\(^2\) (13 grid points). The 180-mm threshold clearly distinguish-

<table>
<thead>
<tr>
<th>Point</th>
<th>Total (mm)</th>
<th>Explicit (mm)</th>
<th>Convective (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>CONT</td>
<td>NOR6</td>
<td>CONT</td>
</tr>
<tr>
<td>1</td>
<td>224</td>
<td>165 (26%)</td>
<td>115</td>
</tr>
<tr>
<td>2</td>
<td>214</td>
<td>146 (32%)</td>
<td>96</td>
</tr>
<tr>
<td>Ave.</td>
<td>202</td>
<td>156 (25%)</td>
<td>101</td>
</tr>
</tbody>
</table>

The removal of the terrain variation results in a reduction of over 20% in average PR48 over the region (Table 6). The local reductions can exceed 30% (Table 7). Therefore, orographic variation contributes approximately one-fifth of the total PR48 in the Saguenay region and over one-quarter at local points.

We have shown that the changes to the total precipitation in NOR6 (and NOR1) are entirely due to the removal of both the valley (VAL) and the southeast mountain (SE). Since the total effect on the precipitation can be regarded as the sum of not only the individual effects of VAL and SE but also of the nonlinear interactions between them, we applied the factor separation technique of Stein and Alpert (1993) to determine the importance of this interaction. [Since NOR and SW had little effect on the total precipitation (see Table 4), they are not included in the factor separation analysis.] Four experiments are required: NOR6 (both VAL and SE removed), NOR4 (VAL present and SE removed), NOR5 (SE present and VAL removed), and CONT (both

<table>
<thead>
<tr>
<th>Location of PR48 maximum in CONT</th>
<th>Location of maximum difference in PR48 between CONT and NOR6</th>
</tr>
</thead>
<tbody>
<tr>
<td>Difference in PR48</td>
<td>59.5 mm</td>
</tr>
<tr>
<td>PR48 value</td>
<td>224.1 mm</td>
</tr>
<tr>
<td>Percent difference</td>
<td>26.6%</td>
</tr>
<tr>
<td>67.2 mm</td>
<td>213.2 mm</td>
</tr>
<tr>
<td>31.5%</td>
<td></td>
</tr>
</tbody>
</table>

Table 7. Differences in PR48 between runs CONT and NOR6 at the grid point with the highest PR48 value in CONT and the grid point with the highest difference in the PR48 field between runs CONT and NOR6.
VAL and SE present), and the contributions calculated as:

i) contribution from VAL only: \( \text{NOR4} - \text{NOR6} \)

ii) contribution from SE only: \( \text{NOR5} - \text{NOR6} \)

iii) contribution from interaction between VAL and SE:

\[ \text{CONT} - (\text{NOR4} + \text{NOR5}) + \text{NOR6} \]

The area-average contributions for the area of the Saguenay PR48 maximum and the area of main PR48 differences are summarized in Table 8. The nonlinear interaction between VAL and SE clearly played only a minor role. The total effect on PR48 is essentially the linear combination of the contributions of VAL and SE.

6. Conclusions

We have used a state-of-the-art mesoscale model, at a horizontal grid spacing of 20 km, with a sophisticated physics package to perform a 48-h simulation of the Saguenay cyclone. Comparison with observations and analyses demonstrated that the simulation performed well in terms of the mass and wind fields, humidity distribution, and the quantitative forecasting of precipitation both in space and time.

By applying PV diagnostics and sensitivity experiments, we quantified the relative importance of various forcing mechanisms and their interactions in rapid cyclogenesis and the production of a large amount of precipitation and are in a position to answer the questions posed in the introduction. The explosive deepening of the storm resulted from the combined effects of upper-level divergence caused by an upper-level jet streak and two short-wave troughs, latent heat release from moist air originating from the Gulf of Mexico, and warm advection from strong low-level winds in a low-level baroclinic zone. The northern trough was the main source of upper-level forcing. The flow field associated with the southern trough was important in retarding the progression of the northern trough, which essentially steered the cyclone, ultimately locating it in a favorable position for the surface winds to flow upslope from the Saguenay valley to the mountain located immediately to the southeast resulting in enhanced upward motion and precipitation. Latent heat release was essential during the initial part of the rapid deepening period to establish a phase lock between the surface cyclone and the northern trough. Sensitivity experiments showed that without latent heating, upper-level forcing would still have been strong but the cyclone would have wandered to the south, most likely being steered initially by the southern trough, and virtually no deepening occurred. Potential vorticity diagnostics indicated that in the middle of the explosive deepening period in CONT, the upper-level dry PV anomaly, mainly associated with the northern trough, contributed the most (47%) to the 900-hPa geopotential height anomaly while the relative contribution to the deepening at that point by the low-level moist PV anomaly, which was associated with latent heat release, was 41%. The PV anomaly associated with surface baroclinicity contributed the least (12%) to the cyclone deepening during the explosive deepening phase, but its associated low-level circulation was important for the initial spinup of the cyclone at the beginning of the deepening period.

Orographic variation between the Saguenay valley and the mountain immediately southeast of the valley significantly enhanced the precipitation rate, most notably in the second half of the integration of CONT, contributing to approximately 20% of the 48-h total accumulated precipitation in the Saguenay region with local effects contributing to over 30% in certain areas. Neither the orography to the north of the Saguenay valley nor the mountain peak immediately southwest of the valley played significant roles in affecting the precipitation for this case.

Another mechanism that made an important contribution to the heavy precipitation was the ridge forming to the east of the cyclone, which blocked the cyclone movement after 1200 UTC 20 July. The cyclone thus remained quasi-stationary for approximately 24 h allowing continued precipitation over the Saguenay region for a long duration. Without this blocking, the cyclone would have moved more quickly to the northeast, reducing the effect of the upslope flow, and ultimately resulting in less total precipitation in the Saguenay region. A complete discussion of the mechanisms that contributed to the heavy precipitation after the cyclone reached its minimum central pressure should include an analysis of the intensification of the ridge that blocked the movement of the storm system. Further investigation into this case should therefore include a closer study of the development and interactions of the residual PV anomaly, which consisted mainly of negative PV associated with the ridge.

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REFERENCES


