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A Diagnostic Examination of the Eastern Ontario and Western Quebec Wintertime Convection Event of 28 January 2010

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ABSTRACT

The priority of an operational forecast center is to issue watches, warnings, and advisories to notify the public about the inherent risks and dangers of a particular event. Occasionally, events occur that do not meet advisory or warning criteria, but still have a substantial impact on human life and property. Short-lived snow bursts are a prime example of such a phenomenon. While these events are typically characterized by small snow accumulations, they often cause very low visibilities and rapidly deteriorating road conditions, both of which are a major hazard to motorists. On the afternoon of 28 January 2010, two such snow bursts moved through the Ottawa River valley and lower St. Lawrence River valley, and created havoc on area roads, resulting in collisions and injuries. Using the National Centers for Environmental Prediction (NCEP) North American Regional Reanalysis (NARR), these snow bursts were found to be associated with an approaching strong upper-tropospheric trough and the passage of an arctic front. While convection or squall lines are not common in January in Canada, snow bursts are shown to be associated with strong quasigeostrophic forcing for ascent and low-level frontogenesis, in the presence of both convective and conditional symmetric instability. Finally, this paper highlights the need for the development of a standard subadvisory criterion warning of short-lived but high-impact winter weather events, which operational forecasters can issue and quickly disseminate to the general public.

1. Introduction

Occasionally, meteorological events occur that do not meet warning, watch, or advisory criteria, but have a significant impact on life, loss, and property. DeVoir (2004) pointed out that events “characterized by extremely heavy but short-lived snow bursts or squalls” fit into such a category. Snow bursts or squalls have the potential to create very low visibility and dangerous driving conditions, in addition to anxiety and confusion for the general public, owing to the rapid and often unexpectedly changing environment (DeVoir 2004). Additionally, aviation can often be affected, as winter precipitation and low visibility can greatly impact airport operations, albeit for relatively short periods of time during events such as the

one discussed here (i.e., compared to a winter storm or blizzard). This paper focuses on one such event that impacted eastern Ontario and western Quebec in Canada, and northern New York State on 28 January 2010. For example, the Ottawa, Ontario (CYOW), region reported dozens of automobile accidents that resulted in at least one person being critically injured, which is further detailed in section 3. So as not to confuse the reader with “snow squalls” that would typically be associated with lake-effect precipitation regions, we will subsequently refer to these events as “snow bursts,” following Pettegrew et al. (2009).

To our knowledge, there exists little in the way of published literature on the topic of cold-season, high-impact, limited-moisture events (Pettegrew et al. 2009). Pettegrew et al. (2009) is a recent study that included a synoptic–dynamic analysis on an event similar to the one in this paper, which created near-whiteout conditions in Iowa and Illinois in 2003, while only depositing 4 cm of snow. DeVoir (2004) briefly described two such cases in the eastern United States, but did not discuss the

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details of the dynamics and thermodynamics associated with the events. Instead, the focus was more on the human impacts of the snow bursts and the need for the National Weather Service (NWS) to establish a subadvisory-scale warning criterion. In an NWS technical bulletin, Lundstedt (1993) analyzed snow squalls in northern New England not associated with lake-effect snow. Lundstedt (1993) computed a wintertime instability index (WINDEX) to establish the potential for snow squalls typically associated with the passage of an Arctic cold front. WINDEX was based on a combination of low-level lapse rates, boundary layer relative humidity, and the 12-h change in the lifted index. However, to our knowledge, there are no references to Lundstedt (1993) in the refereed literature and it is unclear whether WINDEX was ever implemented operationally.

Three studies have focused on thundersnow events. Market et al. (2002) established a synoptic-scale climatology of such events, Market et al. (2006) performed a composite analysis of proximity soundings during thundersnow occurrences, and Crowe et al. (2006) analyzed the relationship between thundersnow occurrences and deep 24-h snow accumulations. Market et al. (2002) showed that the majority of thundersnow events occur in association with a synoptic-scale cyclone. Market et al. (2002) also showed that thundersnow did occasionally occur in association with other types of events, such as orographically forced, Arctic front passages, and lake-effect snow, but at a much lower frequency than events associated with a synoptic-scale cyclone. Crowe et al. (2006) showed that thundersnow occurrences within an extratropical cyclone tended to indicate a cyclone capable of producing large snowfall accumulations somewhere in its path, although not necessarily collocated with the thunder observations themselves.

Other work has been mostly limited to examining the dynamic and thermodynamic properties of specific cold-season events involving lightning or severe weather (Holle and Watson 1996; Schultz 1999; Hunter et al. 2001; Trapp et al. 2001; van den Broeke et al. 2005; Corfidi et al. 2006). In particular, Schultz (1999) analyzed the difference between lake-effect events with and without lightning for both northern Utah and western New York State, two regions susceptible to frequent lake-effect events (Niziol et al. 1995; Steenburgh et al. 2000; Steenburgh and Onton 2001). Overall, the large majority of case studies regarding occurrences of snow bursts are directly related to either heavy lake-effect snow events (e.g., Niziol et al. 1995; Steenburgh and Onton 2001; Payer et al. 2007), or heavy mesoscale bands within synoptic-scale heavy snowfall events (e.g., Nicosia and Grumm 1999). In fact, events marked by small amounts of snow are rarely analyzed in the literature (Homan and Uccellini 1987);

since, in general, they have a relatively minor impact on human life and property.

Many of the lessons learned from dynamic and thermodynamic analyses of lake-effect snow squalls and heavy mesoscale precipitation bands within larger synoptic-scale heavy snowfall events can be applied to the analysis in this paper. However, this paper differs from all previous work other than Pettegrew et al. (2009) and Lundstedt (1993) in that we are analyzing the dynamics and thermodynamics associated with a low-quantitative precipitation forecast (QPF), high-impact precipitation event in a region not prone to lake-effect snow.

In section 2, an overview of the data used in this study is provided. Section 3 details the surface observations and human impacts of the event on 28 January 2010, while section 4 contains the dynamic and thermodynamic analysis of the snow bursts. Finally, section 5 provides a summary, an outline of future work, and brief recommendations for the operational handling of future similar events.

2. Data

The radar imagery in this paper was obtained using the Environment Canada (EC) historical radar database (available online at http://www.climate.weatheroffice.gc.ca/radar/index_e.html). The radar imagery shown in this paper is entirely from the Franktown (CXFT) radar in eastern Ontario. EC radars are C-band radars with a wavelength of 5 cm and a beamwidth of 0.65°. These radars operate in a continuous scanning mode with the typical volume scan that lasts 5 min. Additionally, EC radars have a Doppler coverage area that is approximately 256 km in diameter. In the winter season, the radar generally operates in snow precipitation mode. (Meteorological Service of Canada 2010b).

The surface observations incorporated into the meteorograms were obtained from Iowa State University's Iowa Environmental Mesonet archive (available online at <http://mesonet.agron.iastate.edu/archive/>). Precipitation data were acquired from the EC historical climate database (located online at http://www.climate.weatheroffice.gc.ca/climateData/canada_e.html). EC forecasts were obtained in real time from the EC Weatheroffice Web site (http://www.weatheroffice.gc.ca/canada_e.html). All dynamic and thermodynamic analyses performed for this manuscript were completed using the National Centers for Environmental Prediction (NCEP) North American Regional Reanalysis (NARR), with a horizontal resolution of 32 km, and a 3-hourly temporal resolution (Mesinger et al. 2006). NARR fields were compared with the 1/2° horizontal resolution NCEP Global Forecast System (GFS) analysis and the NCEP-NCAR Global Reanalysis (Kalnay et al.

FIG. 1. The EC radar imagery from the Franktown (CXFT) for 28 Jan 2010 at (a) 1510, (b) 1610, (c) 1710, (d) 1810, (e) 1910, and (f) 2010 UTC. The approximate location of CYOW (KMSS) is marked in each panel with a black star (red oval). The first snow burst is marked SB1 with a blue arrow and the second snow burst is marked SB2.

1996) and were found to be qualitatively similar in terms of both quasigeostrophic (QG) dynamic and thermodynamic structures. The calculations and analyses in this study are displayed using the General Meteorological Package (GEMPAK) version 5.11.1 [updated from the original package devised by Koch et al. (1983)], a data manipulation and visualization software package.

3. Case overview

This paper focuses on the causes and impacts of the snow burst event of 28 January 2010, which occurred in association with the passage of an Arctic front (detailed in section 4) through parts of eastern Ontario, western Quebec, and northern New York State. For our surface station diagnostics we concentrate on two aviation routine weather report (METAR) stations: 1) CYOW and 2) Massena, New York (KMSS), the latter of which is located 70 km southeast of the former; these stations are marked with a black star and red oval, respectively, in Fig. 1a. On the afternoon of 28 January 2010 two separate snow bursts affected CYOW, KMSS, and their surrounding regions, with one moving through CYOW at approximately 1800 UTC and the other just after 1900 UTC. For the remainder of this manuscript we define $t = 0$ h as the approximate hour that the first snow burst moved through CYOW, at 1800 UTC.

Figure 1 shows a time series of radar imagery from the EC Franktown (Ottawa) radar. Images are shown every hour from 1510 to 2010 UTC as the snow bursts crossed the area. In Fig. 1a, a line of moderate to heavy snow is observed in Ontario along the St. Lawrence River (from east of Kingston northeastward to Cornwall) at $t = -3$ h. This band of snow persisted for approximately two more hours before moving into the Adirondack Mountains of New York State. Although this first snowband did reduce visibility to less than 1 statute mile (sm) at Kingston, Ontario (not shown), and KMSS (Fig. 2b) prior to $t = 0$ h, it is not the focus of the paper, as the two later snow bursts (which appear as convective squall lines in Figs. 1c–e) had a substantially larger impact on the region (Jackson 2010; Spears 2010).

The first convective snow burst (hereafter SB1, and marked in Fig. 1) is first observed on the radar imagery at $t = -3$ h, located approximately 100 km west of CYOW (Fig. 1a). By $t = -2$ h and $t = -1$ h it extends in a north–south-oriented line from just north of the Ottawa River (located on the Ontario–Quebec border) to the St. Lawrence River just west of Brockville, Ontario (Figs. 1b and 1c). Also at $t = -1$ h (Fig. 1c) the second convective snow burst (hereafter SB2 and marked in Fig. 1) is first evident just west of where SB1 originally formed an hour earlier (Fig. 1b). By $t = 0$ h (Fig. 1d) SB1 is moving through CYOW and SB2 is approaching from

FIG. 2. Meteorograms for 28 Jan 2010 at (a) CYOW and (b) KMSS. In the top panels of (a) and (b), temperature (dewpoint) is plotted in red (green) in $^{\circ}\text{C}$. In the bottom panels of (a) and (b), visibility (sm) is on the vertical axis (blue, not shown for values >3 sm), and wind [in knots (kt) where $1 \text{ kt} = 0.514 \text{ m s}^{-1}$] is represented by barbs. Time on the horizontal axis is relative to $t = 0$ h (1800 UTC).

the west at approximately 50 km h^{-1} . SB2 moves through CYOW at $t = +1$ h (Fig. 1e) at approximately the same time that SB1 passes through KMSS. Finally, as SB1 moves eastward into Quebec by $t = +2$ h (Fig. 1f), SB2 approaches KMSS and moves through the station around $t = +3$ h (not shown). The maximum observed reflectivity value for both of the convective lines was between 40 and 45 dBZ, observed in both SB1 and SB2 at $t = 0$ h (Fig. 1d).

Figure 2 presents meteorograms for both CYOW (Fig. 2a) and KMSS (Fig. 2b) for 28 January 2010. Prior to the passage of the snow bursts (and associated Arctic front), the 10-m winds at both stations were observed to be mainly southerly. At CYOW (Fig. 2a), a wind shift (to westerly) is observed at $t = 0$ h in association with the passage of SB1. The visibility also drops to 1.5 sm and snow showers are reported at $t = 0$ h (Fig. 2a).

TABLE 1. Raw METAR reports from CYOW from 1700 ($t = -1$ h) to 2100 UTC 28 Jan 2010 ($t = +3$ h), during the passage of the snow bursts. Special reports issued between hourly observations are included.

Ottawa, ON (CYOW), 1700–2100 UTC 28 Jan 2010
CYOW 281700Z 18007KT 8SM -SN BKN026 OVC050 M03/M07 A2971 RMK SC6SC2 SLP069
CYOW 281726Z 20009KT 1 1/2SM -SN OVC022 M03/RMK SN3SC5
CYOW 281800Z 24009KT 1 1/2SM -SHSN OVC019 M04/M06 A2969 RMK SN2SC6 SLP060
CYOW 281803Z 24010KT 1/2SM R07/2800VP6000FT/D R32/3500VP6000FT/D SHSN VV007 M05/RMK SN8
CYOW 281812Z 27019G28KT 3/8SM R07/2000FT/N R32/1200FT/N SHSN DRSN VV005 M06/RMK SN8
CYOW 281900Z 27012G17KT 15SM SCT018 BKN030 OVC050 M05/M09 A2971 RESN RMK SC3SC2SC3/S01/SLP067
CYOW 281914Z 27012KT 6SM -SN BKN026 OVC050 M05/RMK SC6SC2 VIS N 2 CVCTV CLDS EMBD
CYOW 282000Z 28023G31KT 1/4SM R07/0700V1200FT/N R32/0600V1100FT/N -SN +BLSN VV005 M09/M09 A2976 RESN RMK SN8 SLP084
CYOW 282009Z 29021G31KT 3/8SM R07/2000FT/N R32/1100V1600FT/N -SN BLSN OVC016 M08/ RMK SN6SC2
CYOW 282100Z 27022KT 15SM DRSN FEW020 BKN045 M08/M13 A2980 RMK SC2SC5 OCNL BLSN SLP100

Radar imagery at $t = 0$ h (Fig. 1d) strongly suggests that SB1 moved through CYOW around $t = 0$ h. This is substantiated by a special METAR report (Table 1) issued at 1803 UTC that reported a visibility of $\frac{1}{2}$ sm (with snow showers) and by a second report at 1812 UTC that reported a visibility of $\frac{3}{8}$ sm and a wind that shifted strongly to the west (with reported snow showers). Consequently, the observed visibility during the passage of SB1 at CYOW was actually substantially lower than seen in Fig. 2a, since it occurred in between hourly surface observations (Table 1). Given that SB2 passes through CYOW just after $t = +1$ h (Fig. 1d, Table 1), we believe the report of $\frac{1}{4}$ -sm visibility at $t = +2$ h (Fig. 2a) is mostly representative of blowing snow, in association with the 25-kt 10-m wind (Fig. 2a). This is evidence that visibility was hampered both during and after the passage of SB1 and SB2, which is further supported by the drifting snow observations at CYOW at $t = +3$ and $+4$ h (Fig. 2a). Finally, both the temperature and dewpoint show a marked decrease

shortly after the passage of the convective snow bursts and associated Arctic front, with the temperature dropping more than 5°C between $t = -1$ h and $t = +4$ h (Fig. 2a).

KMSS experienced more continuous light precipitation than CYOW, primarily in association with the aforementioned moderate snowband observed along the axis of the St. Lawrence River well before $t = 0$ h (Fig. 2b, Table 2). This is reflected in the meteorograms, as relatively low visibilities were observed while the wind was still southerly, from $t = -6$ to 0 h (Fig. 2b). SB1 moved through KMSS between $t = +1$ and $+2$ h, during which time a wind shift from southerly to westerly occurs (Fig. 2b). METAR reports issued by the ASOS station between these observation times indicate visibilities of less than 1 sm, including $\frac{1}{4}$ -sm visibility at 1911 UTC (Table 2). The lack of a low-visibility observation in the meteorogram (Fig. 2b) during this time is due to the fact that SB1 passed through KMSS between hourly surface observations. By $t = +2$ h, drifting snow

TABLE 2. Raw METAR reports from KMSS from 1700 ($t = -1$ h) to 2100 UTC 28 Jan 2010 ($t = +3$ h), during the passage of the snow bursts. Special reports issued between hourly observations are included.

Massena, NY (KMSS), 1700–2200 UTC 28 Jan 2010
KMSS 281653Z AUTO 18004KT 1/2SM SN FZFG VV002 M04/M06 A2976
KMSS 281713Z AUTO 19004KT 3/4SM -SN BR VV003 M04/M06 A2975
KMSS 281753Z AUTO 19004KT 3/4SM -SN BR BKN005 OVC013 M04/M06 A2972
KMSS 281824Z AUTO 22011KT 1 1/2SM -SN BKN007 OVC016 M03/M06 A2972
KMSS 281853Z AUTO 22014G21KT 6SM HZ SCT019 SCT030 BKN038 M03/M06 A2972
KMSS 281859Z AUTO 22017G23KT 2 1/2SM HZ FEW007 SCT017 BKN023 M03/M07 A2972
KMSS 281901Z AUTO 22019G25KT 3/4SM -SN FEW007 BKN024 OVC048 M03/M06 A2972
KMSS 281911Z AUTO 25016G31KT 1/4SM -SN FZFG BKN010 BKN017 OVC034 M04/M06 A2974
KMSS 281917Z AUTO 25021G31KT 1SM -SN FEW010 BKN020 OVC034 M03/M07 A2974
KMSS 281919Z AUTO 26018G30KT 1 3/4SM -SN FEW008 BKN020 OVC036 M03/M08 A2974
KMSS 281924Z AUTO 24012G30KT 3SM UP FEW008 BKN023 OVC036 M04/M08 A2975
KMSS 281951Z AUTO 23015G21KT 4SM -SN FEW022 SCT027 BKN042 M04/M09 A2976
KMSS 281953Z AUTO 23012G21KT 5SM BLSN FEW022 BKN042 M04/M09 A2976
KMSS 282022Z AUTO 24018G29KT 1 3/4SM -SN SCT024 M04/M11 A2977
KMSS 282025Z AUTO 23019G29KT 1SM -SN VV020 M05/M10 A2978 RMK AO2 PK WND 22029/2017 SNB20 P0000
KMSS 282053Z AUTO 28019G37KT 1/4SM SN FZFG BKN004 OVC016 M08/M10 A2983
KMSS 282100Z AUTO 28019G26KT 1/2SM SN VV006 M07/M11 A2983

is reported (Fig. 2b) in association with a 20-kt westerly wind. The passage of SB2 through KMSS is evident at $t = +3$ h in Fig. 2b and in Table 2, with observed $\frac{1}{4}$ -sm visibility and moderate snow. The METAR reports between the hourly observations (Table 2) suggest that low visibility at KMSS was a more persistent concern than can be observed in the meteorograms, which only show the hourly observations. Finally, the temperature and dewpoint did exhibit a similar drop (Fig. 2b) to that observed at CYOW (Fig. 2a) following the shift to westerly winds.

It is important to emphasize that while this event had a large impact on human interest, it was a relatively low-precipitation event. The total measured snow accumulation for the day at CYOW was 3.6 cm, very similar to the precipitation accumulation in Illinois in the event studied by Pettegrew et al. (2009). At KMSS, only a trace of liquid equivalent precipitation was recorded by the Automated Surface Observing System (ASOS) on 28 January 2010, although it is likely that this is in part due to the high winds and blowing snow observed at KMSS during the time of heaviest snowfall. Although KMSS is not a manual observation station and snow accumulation reports are not available, it is very possible that this event was associated with high snow to liquid water ratios. To that end, the measured liquid equivalent precipitation at CYOW was 1.6 mm, suggesting a snow to liquid water ratio of approximately 22:1. Precipitation data from a cooperative observation station in Ogdensburg, New York (58 km southwest of KMSS and 80 km south of CYOW), was downloaded from the National Climatic Data Center (NCDC) and showed that 2 in. (5.1 cm) of snow was observed on 28 January 2010. Therefore, it is reasonable to assume that KMSS received a similar total amount of snow compared to CYOW. At the Ogdensburg cooperative observation station, 0.04 in. of liquid equivalent was recorded, suggesting an even higher snow to liquid ratio (approximately 50:1).

The impact of these snow bursts was large, particularly in the CYOW area, where several multicar pileups and dozens of automobile accidents were reported (Jackson 2010). Moreover, several minor injuries were reported and at least one person (a 13-year-old boy) was critically injured when he was struck by a car in Ottawa around 1500 local time, or just after the passage of SB2 (Jackson 2010). In all, 18 vehicle accidents were reported in Ottawa during SB2 alone (Spears 2010), prompting the Ottawa police to

issue a press release for motorists to use extreme vigilance in the area (Spears 2010).

Environment Canada's Ontario Storm Prediction Centre did not issue any warnings for eastern Ontario during this event, as snow bursts did not fit their warning criteria for any type of wintertime event in the province. However, a special weather statement was issued between $t = -1$ and 0 h that called for "brief, but intense bursts of snow over eastern Ontario." Meanwhile, the Quebec Storm Prediction Centre in Montreal issued a "snow squall warning" for regions in western Quebec around $t = 0$ h, while SB1 was passing through CYOW. Warning criteria vary from one province/territory to another, which explains the different handling of the same event by these two storm prediction center; however, neither is ideal in effectively communicating the risks associated with snow bursts. The significant societal impacts that resulted from the 28 January 2010 event clearly highlight the need for the development of a broadly defined advisory or warning that can be applied to events such as this one, as first mentioned by DeVoir (2004). These points are discussed further in section 5, following the dynamic and thermodynamic analysis of the event.

4. Dynamic and thermodynamic analysis

Both Doswell et al. (1996) and Schultz and Schumacher (1999) argue for the use of an ingredients-based methodology in the diagnosis of particular varieties of moist convection (gravitational and slantwise). Moreover, Wetzel and Martin (2001) made a similar argument for the analysis and operational prediction of midlatitude winter precipitation events, highlighting five specific ingredients. Three ingredients are common to all three studies: lift, instability, and moisture. We apply this approach to the analysis in this section, because even though southern Canada in January is not a region in which one would typically consider occurrences of moist convection, this case has all the markings of such an event. Furthermore, while this event clearly did not produce a large amount of precipitation (section 3), it contained enough moisture and high snow to liquid water ratios to create havoc for the general public. Since the largest impacts to the public occurred in the CYOW region, the focus of the following analysis will be on this location.

Two forms of the adiabatic, frictionless QG omega equation are used in this paper and can be shown as [see Eq. 5.6.11 in Bluestein (1992)]

$$\left(\nabla_p^2 + \frac{f_o^2}{\sigma} \frac{\partial^2}{\partial p^2} \right) \omega = - \frac{f_o}{\sigma} \frac{\partial}{\partial p} [- \mathbf{v}_g \cdot \nabla_p (\zeta_g + f)] + \frac{R}{\sigma p} [- \nabla_p^2 (- \mathbf{v}_g \cdot \nabla_p T)] \quad (1)$$

(a)
(b)
(c)

where (a) is the three-dimensional Laplacian of vertical motion (ω), (b) represents differential vorticity advection, and (c) is the horizontal Laplacian of temperature advection. In Eq. (1), f_o is the Coriolis parameter, σ is the static stability parameter, ω is the vertical velocity in pressure coordinates, $\mathbf{v}_g$ is the geostrophic wind vector, $\nabla_p(\zeta_g + f)$ is the gradient of geostrophic absolute vorticity on a constant pressure surface, and R is the gas constant for dry air. The $\mathbf{Q}$ -vector divergence is related to vertical motion by [see 5.7.58 in Bluestein (1992)],

$$\left(\nabla_p^2 + \frac{f_o^2}{\sigma} \frac{\partial^2}{\partial p^2} \right) \omega = -2 \nabla_p \cdot \mathbf{Q}, \quad (2)$$

the $\mathbf{Q}$ -vector form of the inviscid adiabatic QG omega equation in which the sense of vertical motion is related to the divergence of the $\mathbf{Q}$ vector. This is supported by the assertion of Hoskins et al. (1978), which states that “in quasi-geostrophic theory... vertical velocity is forced solely by the divergence of $\mathbf{Q}$.” We define $\mathbf{Q}$ as

$$\mathbf{Q} = -\frac{R}{\sigma p} \left(\frac{\partial \mathbf{v}_g}{\partial x} \frac{\partial}{\partial y} \right) \cdot \nabla_p T, \quad (3)$$

where $\nabla_p T$ is the horizontal temperature gradient and p is pressure.

Figure 3 displays 850–500-hPa $\mathbf{Q}$ -vector divergence, sea level pressure (SLP), and 925-hPa equivalent potential temperature (θ_e). Figures 3a and 3b show a broad area of 850–500-hPa layer-averaged $\mathbf{Q}$ -vector convergence, forcing QG ascent, located west of CYOW and just downstream of a midtropospheric trough (Fig. 4). By $t = -6$ h (Fig. 3c), CYOW is located in the center of the large area of $\mathbf{Q}$ -vector convergence, which has strengthened relative to $t = -18$ h (Fig. 3a). Simultaneously, a large low-level θ_e gradient is present over the region, supporting the earlier assertion of an Arctic cold front approaching and subsequently moving through the CYOW region. By $t = -3$ and 0 h (Figs. 3d and 3e), CYOW remains in the region of ascent forcing, as a weak sea level cyclone develops in this area. In fact, at $t = 0$ h, CYOW is clearly located in an area of cold-air advection, which, in considering Eq. (1), suggests that the main mechanism responsible for QG ascent is cyclonic vorticity advection. Regardless, it is clear that CYOW is located in a large and persistent region of $\mathbf{Q}$ -vector convergence (and thus QG ascent) from $t = -12$ to 0 h, when SB1 passed through the station. By $t = +3$ h (Fig. 3f), most of the $\mathbf{Q}$ -vector convergence moves into Quebec as the snow burst event in eastern Ontario comes to an end. Additionally,

there is a large sea level pressure gradient between a strong high pressure system over the central United States and a weak low pressure system over eastern Ontario and western Quebec; this gradient certainly suggests relatively strong surface winds at the time of the snow bursts, which likely contributed to the low visibilities in the region by causing blowing and drifting snow. Finally, it is clear from Fig. 3 that CYOW and KMSS were located in a region of quasigeostrophic forcing for ascent. Thus, we now examine possible focus mechanisms for ascent below, to help explain why the forcing for ascent is so broad and the response (i.e., the snow bursts) is relatively narrow.

Figure 4 displays NARR potential temperature (θ) on the dynamic tropopause [DT, defined here as the 2-PVU surface, where 1 potential vorticity unit (PVU) = $10^{-6} \text{ m}^2 \text{ s}^{-1} \text{ K kg}^{-1}$], 10-m wind, and the coupling index (CI), a measure of the bulk atmospheric stability, used by Bosart and Lackmann (1995), Roebber and Gyakum (2003), and Galarneau and Bosart (2005) as θ on the DT minus the low-level θ_e . We define the CI as

$$\text{CI} = \theta_{\text{DT}} - \theta_{e850}, \quad (4)$$

where the DT is the 2-PVU surface and the low-level θ_e is taken at 850 hPa (Galarneau and Bosart 2005). Figures 4a and 4b show the upper-level trough (low θ on the DT) upstream of CYOW with low values of the CI located in the center of the trough. By $t = -12$ h (Fig. 4b), a substantial gradient of θ on the DT sets up just west of CYOW, documenting the approach of an Arctic cold front. At $t = -6$ h (Fig. 4c) and $t = -3$ h (Fig. 4d), CYOW is located in the region of the 10-m wind shift, with southerlies (easterlies) to the east at $t = -6$ h ($t = -3$ h) and westerlies to the west. At $t = 0$ h, CYOW is now located on the eastern edge of the coldest θ air (Fig. 4e). Simultaneously, very low values of CI (< 8 K) are present over CYOW, suggesting that the air mass at the initial onset time of the snow bursts was marked by relatively low-tropospheric stability. By $t = +3$ h (Fig. 4f), the coldest θ air and lowest CI values have moved to the east, into eastern Ontario and western Quebec, which is concomitant with the snow bursts moving through KMSS. In conjunction with the results of Fig. 3, we suggest that the narrowness of the snow bursts may be related to the relatively narrow region in which very low values of CI (Figs. 4e and 4f) overlap with large values of $\mathbf{Q}$ -vector convergence (Figs. 3e and 3f). This is discussed further in section 5.

Pettegrew et al. (2009) found substantial low-level (925 hPa) frontogenesis associated with the snow burst in Illinois. Furthermore, frontogenesis has been suggested to play an important role in intense mesoscale bands of snow, as mentioned by Nicosia and Grumm

FIG. 3. Time series of NARR SLP (hPa, solid contours), 925-hPa equivalent potential temperature (K, dashed), and 850–500-hPa layer-averaged $\mathbf{Q}$ -vector divergence $\times 10^{-18} \text{ K m}^{-2} \text{ s}^{-1}$ (shaded cool colors for convergence, warm colors for divergence) on 28 Jan at (a) 0000 UTC ($t = -18$ h), (b) 0600 UTC ($t = -12$ h), (c) 1200 UTC ($t = -6$ h), (d) 1500 UTC ($t = -3$ h), (e) 1800 UTC ($t = 0$ h), and (f) 2100 UTC ($t = +3$ h). The approximate location of CYOW is marked with a black star.

FIG. 4. As in Fig. 3, but for potential temperature (K, shaded) on the dynamic tropopause (2-PVU surface), 10-m wind (barbs), and coupling index (solid, every 4 K from 0 to +16).

(1999) and Hunter et al. (2001). Frontogenesis helps to serve as a focusing mechanism for ascent through a thermally direct circulation, supporting rising motion on the warm side of the cold front (Martin 2006, p. 201). Figure 5 displays NARR 925-hPa total wind frontogenesis, 10-m wind, 1000–500-hPa thickness, and steep values ($\geq 8 \text{ K km}^{-1}$) of the 925–700-hPa lapse rate. At $t = -18 \text{ h}$ (Fig. 5a), two bands of 925-hPa frontogenesis are present: one is located in a line stretching from northeastern Ontario southward through Lake Michigan and into the Mississippi River valley, while the second is located farther to the northwest in northwestern Ontario. Both are indications of a strengthening cold front (Fig. 3) ahead of an Arctic air mass. By $t = -6 \text{ h}$ (Fig. 5c), the first band of frontogenesis has intensified on the leading edge of the frontal zone and is located just west of CYOW. Simultaneously, steep (8 K km^{-1}) low-level lapse rates are present in the area of largest frontogenesis (Fig. 5c). A small area of weak frontogenesis in the St. Lawrence River valley (south of CYOW) is also evident at $t = -6 \text{ h}$ (Fig. 5c). While a full diagnosis is beyond the scope of this paper, we believe this frontogenesis is at least partially related to the local topography, which typically channels surface northeasterly winds down the river valley toward lower pressure, often creating low-level frontogenesis when opposed by geostrophic southerlies. Such a pressure pattern is evident in Fig. 3c and this process might help to explain the pre-SB1 band of snow along the St. Lawrence River that is seen in Figs. 1a–c. By $t = -3$ and 0 h (Figs. 5d and 5e), the frontogenesis throughout Ontario has markedly increased and the 925–700-hPa lapse rates have also steepened ($> 8 \text{ K km}^{-1}$). Both the strongest frontogenesis and steepest lapse rates are located in or near the CYOW area at these times (Figs. 5d and 5e). This suggests that at the time of the snow bursts there was substantial forcing for ascent through frontogenesis in the CYOW area, as well as (at least) a conditionally unstable lower troposphere. These conditions proceeded to move eastward and by $t = +3 \text{ h}$ (Fig. 5f) were firmly entrenched near KMSS and extreme western Quebec, coinciding with the occurrence of the snow bursts in the KMSS and western Quebec region around $t = +3 \text{ h}$. Finally, a band of strong frontogenesis is located in a north–south line in eastern New York State and western New England at $t = -3$ and 0 h (Figs. 5e and 5f). This observation coincides with several lines of intense snow that moved through this area at these times, and affected parts of western and central Massachusetts during rush hour (not shown).

In Figs. 3–5, we have established that SB1 and SB2 in the CYOW and KMSS regions were clearly associated with strong QG forcing for ascent and strong values of low-level frontogenesis. Furthermore, we suggested that these snow bursts occurred in an environment characterized by low

tropospheric stability, which we quantify in Figs. 6–8. Emanuel (1983) pointed out that conditional symmetric (slantwise) instability (CSI) can be largely responsible for some mesoscale precipitation bands within larger-scale storms. However, in cases when convective (potential) instability [$d\theta_e/dz < 0$; Bluestein (1992), p. 222] is collocated with regions of CSI, the convective instability mode tends to dominate over time (Emanuel 1983; Moore and Lambert 1993; Schultz and Schumacher 1999). To assess both convective instability and CSI, a west–east cross-section analysis is presented in Fig. 6, along 45.4°N , roughly the latitude of CYOW; a line representing the cross-section area is included in Fig. 5a. The cross section runs from Petawawa, Ontario (approximately 170 km west of CYOW), to Granby, Quebec (approximately 370 km east of CYOW), allowing for the stability to be assessed across the entire region impacted by the snow bursts.

Figure 6 displays negative (shaded) values of saturated equivalent geostrophic potential vorticity (MPV_g^*), defined as the criterion for CSI by Schultz and Schumacher (1999), and equivalent potential temperature (θ_e). The two quantities plotted serve to diagnose two different types of instability; that is, MPV_g^* is used to diagnose regions of CSI and θ_e is plotted to diagnose regions of convective (potential) instability. Moore and Lambert (1993) and Nicosia and Grumm (1999) mathematically describe MPV_g^* , which is calculated using the built-in GEMPAK potential vorticity function, a layer quantity. The MPV_g^* formula that GEMPAK uses is explicitly defined by Schultz and Schumacher (1999) as

$$\text{MPV}_g^* = g\boldsymbol{\eta}_g \cdot \nabla\theta_e^*, \quad (5)$$

where g is gravity, $\boldsymbol{\eta}_g$ is the three-dimensional geostrophic absolute vorticity vector, ∇ is the gradient operator in x and y , and θ_e^* is the saturated equivalent potential temperature. The cross section is roughly perpendicular to the thermal wind, although Schultz and Schumacher (1999) point out that MPV_g^* is much less sensitive to the orientation of the cross sections than geostrophic absolute momentum (M_g) surfaces, assuming the three-dimensional form of the MPV_g^* equation is used [as in Eq. (5)]. However, recall that $\text{MPV}_g^* < 0$ by itself means relatively little unless convective instability has been assessed first. In other words, CSI should only be assessed from MPV_g^* if the atmosphere is convectively stable (Moore and Lambert 1993). Finally, a plot of M_g (not shown) has affirmed that the atmosphere was inertially stable (Schultz and Schumacher 1999) throughout the area of the cross section.

Figures 6a–d show a layer of $\text{MPV}_g^* < 0$ present below 850 hPa throughout the region prior to the passage of the snow bursts through CYOW (Figs. 6a–c). However,

FIG. 5. As in Fig. 3, but for 925-hPa frontogenesis $\times 10^{-2} \text{ K (100 km)}^{-1} (3 \text{ h})^{-1}$ (shaded), 925–700-hPa lapse rate (K km^{-1} with solid contours starting at -8 with an interval of 0.5), 1000–500-hPa thickness (dam, dashed), and 10-m wind (kt, barbs). A horizontal red rectangle is placed in (a) to represent the cross-section analysis in Fig. 6.

FIG. 6. As in Fig. 3, but west-east cross-section plots from 45.4°N, -77.2°W to 45.4°N, -72.4°W, or approximately 170 km west of CYOW to approximately 150 km east of Montreal's Trudeau International Airport (CYUL). A horizontal red rectangle is in Fig. 5a to represent the cross-section area. Plotted are saturated equivalent geostrophic potential vorticity ($\times 10^{-7} \text{ m}^2 \text{ s}^{-1} \text{ K kg}^{-1}$ shaded for negative values) and equivalent potential temperature (K, solid contours). The location of CYOW is marked on each panel.

this region is predominantly collocated with a layer of convective instability ($d\theta_e/dz < 0$) from $t = -18$ through -6 h (Figs. 6a–c). By $t = -3$ h (Fig. 6d), the layer of convective instability has expanded vertically in the

region of CYOW (and to the west) as the Arctic cold front approached the station. At this time (Fig. 6d), there is now a region of $MPV_g^* < 0$ (CSI) located above the convective instability, albeit one with relatively small

FIG. 7. NARR soundings for CYOW on 28 Jan at (a) 0000 UTC ($t = -18$ h), (b) 0600 UTC ($t = -12$ h), (c) 1200 UTC ($t = -6$ h), (d) 1500 UTC ($t = -3$ h), (e) 1800 UTC ($t = 0$ h), and (f) 2100 UTC ($t = +3$ h). Temperature (dewpoint) is plotted in red (green).

FIG. 8. As in Fig. 7, but for KMSS.

negative values. By $t = 0$ h (Fig. 6e), both the region of convective instability and more negative values of MPV_g^* (CSI) above that are present at CYOW and rapidly moving to the east. This suggests that at the time of the first snow burst, the atmosphere at CYOW was gravitationally and symmetrically unstable, a state known as convective-symmetric instability (Schultz and Schumacher 1999). There is some discrepancy in the literature regarding such a state. Although Emanuel (1983), Moore and Lambert (1993), and Schultz and Schumacher (1999) pointed out that convective instability typically dominates in such a situation, Schultz and Schumacher (1999) state that CSI may actually be present before convective instability, even if convective instability is likely to dominate over time. However, establishing this case's dominant instability is not our primary focus. Instead, we choose to emphasize that regardless of the type of instability or instabilities present, the state of the atmosphere was extremely conducive to a cold-season convective event. It seems reasonable to conclude, however, that convective instability was the dominant mode in this case, although the presence of some CSI should not be ruled out. Finally, at $t = +3$ h (Fig. 6f), the areas of substantial convective instability and CSI (above) have moved eastward and are now present in the KMSS region and southwestern Quebec.

Areas of frontogenesis and CSI are often collocated in regions of intense mesoscale precipitation bands (Schultz and Schumacher 1999; Nicosia and Grumm 1999). Here, we follow the guidelines of Schultz and Schumacher (1999), who state that the separation of forcing (frontogenesis) and response (the release of CSI) is challenging. This point is mitigated slightly by the apparent dominance of the convective instability mode. However, generally speaking, we conclude that both the forcing for ascent (QG forcing and frontogenesis) and instability (convective and conditional symmetric) are present at CYOW and KMSS during the passage of SB1 and SB2.

Bryan and Fritsch (2000) state that, traditionally, five different states of static stability are considered (where γ is the observed environmental lapse rate, Γ_s is the moist-adiabatic lapse rate, and Γ_d is the dry-adiabatic lapse rate):

- 1) absolutely stable, $\gamma = \Gamma_s$;
- 2) saturated neutral, $\gamma = \Gamma_d$;
- 3) conditionally unstable, $\Gamma_s < \gamma < \Gamma_d$;
- 4) dry neutral, $\gamma = \Gamma_d$; and
- 5) dry absolutely unstable, $\gamma = \Gamma_s$.

Bryan and Fritsch (2000) subsequently argue that a sixth type of static stability exists in nature: a "saturated lapse rate (γ_s) that is steeper than the moist-adiabatic lapse rate," defined as

- 6) moist absolutely unstable, $\gamma_s > \Gamma_s$.

Such a state is suggested to be shallow, relatively short lived, and rare, as Bryan and Fritsch (2000) have found that soundings with a moist absolutely unstable layer (MAUL) of 100 hPa or greater composed only 1.1% of over 100 000 examined soundings in their dataset. Additionally, Bryan and Fritsch (2000) found that nearly all deep MAULs occurred in close proximity to moist convection, and that they are likely created by (and are indications of) intense mesoscale vertical motions. Bryan and Fritsch (2000) defined a deep MAUL as a layer of 100 hPa or more with a lapse rate greater than the moist-adiabatic lapse rate and a dewpoint depression $\leq 1^\circ\text{C}$. For the purpose of this study, we have allowed for a higher dewpoint depression given that the temperatures are substantially below zero and the lower troposphere is almost certainly saturated with respect to ice.

In Figs. 7 and 8 we show a time series of NARR soundings for CYOW and KMSS, respectively, on the day of the snow bursts. From $t = -18$ h (Fig. 7a) to $t = -6$ h (Fig. 7c), it is evident that a deep MAUL [according to the definition of Bryan and Fritsch (2000)] exists at CYOW in the near-surface layer up to about 850 hPa. However, by $t = -3$ h (Fig. 7d) and $t = 0$ h (Fig. 7e), this MAUL has grown to encompass approximately 400 hPa, from the surface to around 600 hPa. Figures 7 and 8 also serve to confirm the steep low-level lapse rates seen in Fig. 5. Additionally, the environmental conditions were quite similar at CYOW and KMSS, with a MAUL of approximately the same depth (400 hPa) as analyzed by the NARR at both stations. Finally, while most thundersnow events (Market et al. 2002) have been shown to occur in the presence of elevated instability, the case documented by Pettegrew et al. (2009) showed evidence of being rooted in the boundary layer with the sort of structure one would expect from a warm-season squall line. Figures 7 and 8 show a similar result to Pettegrew et al. (2009) for the cases presented in this paper. It would be interesting to determine which one of these situations is climatologically more common during intense wintertime convection, in association with the passage of an Arctic front, but that is beyond the scope of this manuscript. Finally, we note that while rawinsonde sounding observations took place at times (0000 and 1200 UTC) before and after the snow bursts moved through the area, the 0000 UTC 29 January sounding (not shown) from Maniwaki, Quebec (CWMW), showed structures very similar to the aforementioned NARR soundings.

5. Concluding discussion and future work

On the afternoon of 28 January 2010, two intense snow bursts (SB1 and SB2) moved through eastern Ontario,

western Quebec, and extreme northern New York State, creating low visibilities, falling temperatures, and strong winds, in association with the passage of an Arctic cold front. In following with the ingredients-based methodology for moist convection, proposed by Doswell et al. (1996) and Schultz and Schumacher (1999), we have established that there were large amounts of lift and instability in this case, with just enough moisture (3.6 cm of snow at CYOW) and high snow to liquid water ratios to create havoc on area roads and several injuries throughout eastern Ontario. Specifically, with regard to lift, Ottawa, Ontario (CYOW), and Massena, New York (KMSS), were located in an environment of strong $\mathbf{Q}$ -vector convergence (e.g., QG ascent, see Fig. 3), ahead of a trough with very low- θ air on the dynamic tropopause (DT, Fig. 4). Moreover, Fig. 5 shows a rapidly intensifying band of low-level frontogenesis in the region just before the two snow bursts passed through CYOW and KMSS.

In terms of instability, Fig. 4 shows relatively low values of the coupling index (CI) present in the region, suggesting very low tropospheric stability. Figure 6 displays an area of convective instability in the near-surface layer at least 18 h prior to SB1 moving through CYOW. As the time of snow burst passage ($t = 0$ h) approached, this layer of convective instability expanded and intensified, and was concurrent with a layer of conditional symmetric instability (CSI) above, throughout all of eastern Ontario, western Quebec, and northern New York State. While Fig. 6 suggests that the convective instability mode was dominant, it is possible for convective instability and CSI to both be factors over short periods of time (Schultz and Schumacher 1999). Finally, the soundings in Figs. 7 and 8 corroborate the steep low-level (925–700 hPa) lapse rates in Fig. 5 at the time of the snow bursts, by showing that a moist absolutely unstable layer (MAUL) of approximately 400 hPa was present from the surface to 600 hPa at both CYOW and KMSS. MAULs of such depth are rare, difficult to maintain, and almost always associated with moist convection and very large values of mesoscale ascent (Bryan and Fritsch 2000). Therefore, it is clear from the diagnostics in this paper that this event occurred in a markedly unstable environment, and one that differed from the thundersnow climatology of Market et al. (2002) in which elevated convection was the primary observation, but similar to the case of Pettegrew et al. (2009), in which convection was rooted in the near-surface boundary layer.

We suggest that the snow bursts occurred in the narrow region (including CYOW and KMSS) where the large area of $\mathbf{Q}$ -vector convergence (Figs. 3e and 3f) and low CI values (Figs. 4e and 4f) intersect. This may help to explain why the snow bursts were relatively narrow in scope, despite the large area of synoptic-scale forcing for ascent ($\mathbf{Q}$ -vector convergence). This corresponds to the

area of convective instability observed in Fig. 6, particularly at $t = 0$ h and $t = +3$ h, when the snow bursts crossed the region (Figs. 6e and 6f).

As previously stated, little in the way of published literature exists on this topic and much work remains to be done. Specifically, many more events similar to this one should be examined, as not all Arctic front passages are associated with intense convective snow bursts. For example, a climatology of similar and null events would allow diagnostics to determine what tools an operational forecaster can use to better assess the occurrence potential of such an event, both on the short- and medium-range time scales.

While the forecasting of such an event remains a challenge given the unique conditions required and the infrequency of wintertime convection in Canada, it is not impossible to predict the occurrence of snow bursts. An examination of the NCEP Global Forecast System (GFS) forecasts (not shown) shows that it accurately predicted the dynamic and thermodynamic structures evident in the NARR up to 72 h prior to the occurrence of the snow bursts. Given that Environment Canada uses the Global Environmental Multiscale (GEM) model to produce their forecasts, we acknowledge that using the GFS to analyze operational model performance in this particular case may be a little misleading and that ideally model verification should be performed using the GEM model grids. Regardless, detailed operational model verification is beyond the scope of this paper and should be a topic of future investigation.

The issue, however, is less so that these types of events are predictable and more so that within the wintertime advisories and warnings available, forecast offices cannot adequately express the expected conditions and threats associated with snow bursts. For instance, the Ontario Storm Prediction Center (OSPC) accurately called for 2–4 cm of snow throughout much of eastern Ontario on the day of the snow bursts (3.6 cm was measured at CYOW), but the forecast implied gradual “flurries over the course of the day.” Apart from the high winds also forecast, there was no reason for the public to expect brief whiteout conditions. A special weather statement was issued for CYOW 20 min before SB1 hit the city stating that “narrow but intense bands of snow are moving through southeastern Ontario... Because of the short duration, snowfall accumulations are not expected to be significant; however visibilities may be reduced to a few hundred metres [meters] or less at times.” While the special weather statement accurately expressed the conditions observed in CYOW, these types of statements are not given the same priority as watches and warnings, and presently are not disseminated to the public through the media and Weatheradio, etc. It is

unlikely, especially given the short 20-min window, that this special weather statement available on the Weather-office Web site reached a wide audience prior to SB1 moving through the area. For the same event, shortly after both squalls were evident on radar and moving through eastern Ontario, the Quebec Storm Prediction Centre (QSPC) issued a snow squall warning for western Quebec. A snow squall warning, as defined by EC (Meteorological Service of Canada 2010a), is “when, to the lee of the Great Lakes or other large lakes, snow squalls are expected and 15 cm or more of snow is likely to fall within 12 hours, OR the visibility is likely to be near 0 for 4 or more hours, even without warning level accumulations of snow.” This event was not forecast to meet the warning criteria yet the QSPC issued the warning anyway to ensure the public was informed of the hazardous weather conditions approaching the area. Neither approach adequately warns for, nor expresses, the threats associated with snow bursts.

Our goal is not to criticize the operational forecasters, or to try to differentiate between model forecast and human forecast errors. Instead, we would like to emphasize the fact that, as DeVoir (2004) pointed out for the United States, no warning or advisory exists that is specific to a high-impact, but short-duration and low-precipitation event such as this one. Much like advisories and warnings are issued and understood by the public regarding squalls (either winter or summer), similar messaging is needed for snow bursts. That being said, the issuance of an advisory or warning alone only goes part way to effectively communicating the threats and associated impacts of this type of event to the general public. However, combined with forecaster, public, and media education of these types of events and their impacts, for example through the engagement of the Warning Preparedness Meteorologists (WPM) of Environment Canada with Emergency Measures Organizations (EMOs) and media across Canada, a greater understanding of snow bursts would encourage appropriate actions to be taken by both the media and the public in their respective roles to reduce loss of life and property.

In conclusion, we strongly encourage the creation of a broadly applied warning or advisory (e.g., a “snow burst warning”) that is specific to this type of high-impact but short-duration and low-precipitation event, as proposed by DeVoir (2004), so that the public can be aware that a wintertime convective event is possible on a given day. If such warnings exist for warm-season convective events, we see little reason why they should not be available for cold-season convection. Alternatively, current warnings or advisories such as the snow squall warning could be amended so that a forecaster feels comfortable issuing such a warning or advisory for

an event like the one studied here, without having to “stretch” the language of the written operational definition. Regardless of how such a change is achieved, it is crucial that the operational forecasters have all the necessary tools to alert the general public to forthcoming or imminent hazardous weather conditions.

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