Ensemble Sensitivity of Precipitation Type to Initial Conditions for a Major Freezing Rain Event in Montreal

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ABSTRACT: The predictability of precipitation type in a January 2017 winter storm over the northeastern United States and southeastern Canada is examined using a convective-scale initial-condition ensemble with the Weather Research and Forecasting (WRF) Model. Real-time forecasts of the event by Environment and Climate Change Canada predicted 15–25 cm of snow accumulation in Montreal, Quebec, Canada. However, the initial 4 h of the event had 5–8 mm of freezing rain instead, followed by 7 cm of snow. While the total liquid-equivalent precipitation was consistent with the forecast, the unexpected freezing rain caused significant disruption in the Montreal region. The fraction of freezing precipitation (freezing rain and/or ice pellets) over the initial 4 h in Montreal varied greatly across the ensemble, with some members producing nearly all snow and others producing nearly all freezing precipitation. In members with larger fractions of freezing precipitation (as opposed to snow), the cyclone’s midlevel trough was displaced slightly to the northwest, and its downstream (eastern) edge was narrower, the latter of which was traced back to model initialization. These differences increased the midlevel southerly flow into southern Quebec, which both enhanced the horizontal warm advection and decreased the vertical cold advection leading up to the event. The consequent midlevel warming over Montreal in these members produced an above-zero layer that melted falling precipitation, leading to freezing upon contact with the ground. This case study highlights the value of convective-scale ensembles for identifying mechanisms by which initial synoptic-scale uncertainties lead to high-impact localized errors in precipitation type.

KEYWORDS: North America; Extratropical cyclones; Precipitation; Freezing precipitation; Ensembles

1. Introduction

Freezing rain is a hazardous winter phenomenon capable of large socioeconomic impacts. Ice formed when supercooled raindrops freeze on contact with subfreezing surfaces can accumulate onto roads and sidewalks, creating slippery conditions that disrupt transport. Ice accretion on powerlines and tree branches can overload these structures with weight and consequently damage infrastructure and communication networks. Changnon (2003) found that economic costs associated with freezing rain events in the United States between 1949 and 2000 exceeded 16 billion U.S. dollars (USD 16B; M is similarly used below for million), in year-2000 dollar amount, in insured property damage alone, summed over 87 events. Although winter storms as a whole account for less than 4% of the inflation-adjusted total direct damages where costs exceeded USD 1B (Smith and Katz 2013), Changnon (2003) estimated that 60% of “catastrophic” winter storms (costing at least USD 1.6M, in inflation-adjusted terms) between 1988 and 1995 were associated with freezing rain.

North America exhibits spatially variable distributions of freezing rain occurrence, with hotspots over Newfoundland, Canada, and the St. Lawrence River Valley (SLRV) region of southern Quebec and southeastern Ontario, Canada (Cortinas et al. 2004; McCray et al. 2019). The latter region is particularly vulnerable given its large population, concentrated in Montreal, Quebec (metropolitan population of 4.5M) and Ottawa, Ontario (metropolitan population of 1.5M) (Statistics-Canada 2021). In 1998, for example, a particularly severe ice storm (now known as “The Great Ice Storm”) caused widespread damage within the SLRV and the northeastern United States, with 44 fatalities and damages approaching USD 5B (Lott et al. 1998). Major disruption was also experienced by a multitude of industries, including the dairy, logging, and maple syrup industries (DeGaetano 2000), as well as power and communication networks (Jones and Mulherin 1998).

The microscale mechanisms involved in the formation of freezing rain are well understood (e.g., Brooks 1920; Stewart 1985; Czys et al. 1996; Thériault et al. 2010; Stewart et al. 2015). Freezing rain is most commonly produced when snow, generated aloft, falls through a sufficiently deep layer of above-freezing air to completely melt the hydrometeors. The liquid precipitation then continues to fall through a surface-based subfreezing layer that supercools the hydrometeors, which then freeze upon contact with a subfreezing surface. The presence and strength of the midlevel melting layer is critical to the formation of freezing rain; a shallower and/or cooler melting layer may lead to only partial melting of the hydrometeors, with subsequent refreezing in the surface-based subfreezing layer resulting in ice pellets. The temperature and depth of the surface-based subfreezing layer is also important in determining the phase of the precipitation, with a colder and deeper layer increasing the potential for refreezing of (partially) melted hydrometeors, favoring ice pellets over freezing rain.

Freezing rain has been described as a self-limiting process (Stewart 1985) due to latent heating feedbacks that can render the vertical temperature profile less supportive of freezing rain. The melting of hydrometeors in the midlevel warm layer...
absorbs latent heat from the surroundings, which cools the layer back toward 0°C, while refreezing of the supercooled droplets at the surface releases latent heat, which warms the air back toward 0°C. Erosion of the melting layer may cause a transition from freezing rain to ice pellets and/or snow (e.g., Kain et al. 2000), while warming of the surface-based cold layer may induce a transition to rain (e.g., Lackmann et al. 2002). Both of these feedbacks increase with increasing precipitation rate, making heavier freezing rain more difficult to sustain. Maintenance of the necessary environmental conditions therefore requires other processes to offset these diabatic effects.

The synoptic environments of freezing rain tend to be very specific, mainly because of the need for an elevated inversion and melting layer aloft. Rauber et al. (2001) studied more than 400 cases of freezing rain over a 25-yr period across the central and eastern United States to determine specific weather patterns associated with freezing rain. The most common synoptic context was ahead of a warm front and/or within the occluded sector of a surface cyclone. In this situation, advancing warm air overrides a cold air mass, creating the midlevel melting layer required for freezing rain. As the warm front approaches, sustained warm advection aloft can maintain the midlevel warm layer even in the face of latent-heating feedbacks. However, attendant warm advection in the surface-based subfreezing layer can cause a transition to rain.

Mesoscale orography can also play an important role in prolonging freezing rain, by trapping or channeling pools of subfreezing air to establish or maintain favorable environmental conditions for freezing rain. Twenty-two percent of the freezing rain events in Rauber et al. (2001) featured cold air “damming,” primarily on the eastern side of the Appalachian Mountains. In this process, cold, east-northeasterly near-surface flow to the south of an anticyclone is blocked by the Appalachian terrain and forced south-southeastward along the Appalachian axis, leading to anomalously cold near-surface flow in the southeastern United States. When a cyclone simultaneously approaches bearing midlevel warmth and precipitation, the conditions can become locally favorable for freezing rain.

Freezing rain is also influenced by the southwest–northeast-oriented orography surrounding the SLRV (Fig. 1). Strong along-valley pressure gradients occur when a cyclone approaches from the southwest and/or a surface anticyclone lies to the northeast. In response, ageostrophic northeasterly flow may develop along the valley, advecting cold air through the SLRV to maintain subfreezing conditions (Razy et al. 2012). This terrain-modified ageostrophic flow can thus prolong freezing rain events and enhance their destructiveness. Ressler et al. (2012) identified three distinct synoptic patterns associated with long-duration freezing rain events in Montreal, all of which develop an ageostrophic northeasterly surface flow causing low-level cold advection within the SLRV.

In addition to prolonging freezing rain events, the ageostrophic northeasterly flow channeled through the SLRV can locally enhance precipitation through frontogenesis. These winds may collide with southerly winds ahead of the cyclone, the latter often directed through the adjoining south–north-oriented Champlain Valley (Fig. 1). Roebber and Gyakum (2003) estimated the combined effects of this northeasterly valley-channeled flow in the SLRV to have enhanced the freezing precipitation during the Great Ice Storm of 1998 by more than 12%.

Operational forecasts of precipitation type are challenged by the sensitivity of precipitation processes to the errors inherent to numerical weather prediction (NWP) models. In many winter storms, midlevel temperature variations as small as 0.5–1°C can dictate the type of precipitation reaching the ground (Thériault et al. 2010). Thus, small-amplitude uncertainties in NWP models that are otherwise tolerable can result in large errors in simulated precipitation types. Sources of model error and uncertainty include the imperfect representation of key subgrid physical processes (including cloud microphysics, boundary layer turbulence, and unresolved orographic effects) and the chaotic growth of initial-condition uncertainties.

To add to the forecasting challenge of freezing rain, even the very diagnosis of simulated precipitation type is uncertain. Bulk microphysics parameterizations used in NWP models often lack the sophistication to accurately track the full diversity and distributions of hydrometeor states needed for explicit precipitation-type diagnosis. This shortcoming necessitates the development of implicit postprocessing algorithms to diagnose precipitation type from vertical thermodynamic profiles (e.g., Ramer 1993; Baldwin et al. 1994; Bourgouin 2000). These algorithms typically derive empirical criteria based on observations of winter precipitation events and apply these criteria to the simulated soundings. Reeves et al. (2014) tested five commonly used algorithms on observed soundings of four winter precipitation types (snow, ice pellets, freezing rain, and rain) and found that none of them could correctly diagnose the precipitation type with an accuracy of >50% for both freezing rain and ice pellets. Moreover, the ability to correctly identify either of these types was at least 15% smaller than for rain or snow for almost all algorithms. Wandishin et al. (2005) and Cortinas et al. (2002) also found that no algorithm
was universally superior at diagnosing freezing rain and ice pellets.

Advances in microphysics parameterizations to include multistate hydrometeors (e.g., Thompson et al. 2008) have allowed for some NWP models to explicitly diagnose precipitation type based on hydrometeor mixing ratios, fall speeds and surface temperatures (e.g., Benjamin et al. 2016; Milbrandt et al. 2016; Ikeda et al. 2017). While generally more accurate than implicit algorithms, these methods are still susceptible to model biases, especially in the surface and/or midlevel temperatures (Ikeda et al. 2017). Additionally, microphysics parameterizations often contain questionable assumptions (e.g., instantaneous melting of frozen hydrometeors in above-frozen temperatures; Iversen et al. 2021), adding to the uncertainty associated with explicit precipitation-type algorithms.

Although improvements to NWP models and precipitation-type diagnoses can reduce forecast errors, there is an intrinsic limit on the predictability of any forecast due to the nonlinear growth of initial-condition uncertainties (Lorenz 1963). Errors arising from these uncertainties can be compounded by the aforementioned errors from physical parameterizations and precipitation-type diagnoses. For example, implicit precipitation-type algorithms can be heavily influenced by small shifts in the atmospheric thermal profile induced by initial-condition uncertainty (Wandishin et al. 2005; Reeves et al. 2014). Given the potentially strong sensitivity of winter precipitation type to initial-condition uncertainties, it is also valuable to understand the physical mechanisms by which these uncertainties manifest into precipitation-type errors in areas of interest. Benefits of such analysis include identification of sensitive regions in the upstream flow where assimilation of additional observations may help to reduce precipitation-type errors.

Operational centers routinely produce forecast ensembles to account for the nonlinear growth of errors. These account for initial-condition uncertainty by sampling from a plausible range of perturbed initial states and may also address model-physics uncertainty by using different formulations of physical parameterizations. Along with benefitting forecasting by producing an envelope of forecast realizations, ensembles are useful for gaining physical insight into the sensitivity of meteorological processes to initial-condition and/or model-physics uncertainties. Studies have interpreted ensemble sensitivities of various processes including winter snowfall (e.g., Grubišić et al. 2005; Novak and Colle 2012; Greybush et al. 2017; Saslo and Greybush 2017), cyclone track errors (e.g., Torn and Cook 2013; Torn et al. 2015), and convective storms (e.g., Hanley et al. 2013; Schumacher et al. 2013; Bednarczyk and Ancell 2015; Trier et al. 2015), among others. However, to the authors’ knowledge, no such analysis has been performed with regard to precipitation type, particularly in the geographic regions of most frequent freezing rain occurrence identified by Cortinas et al. (2004) and McCray et al. (2019).

In this study, we use high-resolution ensembles to interpret the mechanisms by which synoptic-scale uncertainty in the initial conditions can lead to local variations in precipitation type in an east coast winter storm. This effort is motivated in part by a misforecast freezing rain event over Montreal on 24 January 2017. Real-time operational forecasts of the event by Environment and Climate Change Canada (ECCC) predicted between 15 and 25 cm of snow over southern Quebec, triggering a “snowfall warning” (>15 cm of snow) alert for the region. Instead, Montreal International Airport (CYUL in Fig. 1) saw predominantly freezing rain in the first 4 h of the event [0700 EST (1200 UTC) to 1100 EST (1600 UTC)], which transitioned to snow during the afternoon. A total of 12.4 mm of liquid-equivalent precipitation was measured at CYUL with 7.0 cm of snow, implying a reasonable quantitative precipitation forecast. The unforecast freezing rain, however, led to major disruptions in the morning commute and numerous school closures. The large precipitation-type error in this forecast suggests a low degree of predictability, making this event an attractive subject for ensemble sensitivity analysis.

The primary objective of this study is not to resolve why the ECCC operational forecast was incorrect, but rather to investigate the mechanisms by which initial-condition uncertainties on the synoptic scale can limit the predictability of local precipitation type. While accurate weather predictions are always desirable, the importance of forecast skill is amplified when their errors result in large societal impacts. Despite the SLRV being a continental freezing rain hotspot, few previous studies have focused on understanding past freezing rain events in the SLRV region, with the exception of the Great Ice Storm (e.g., Gyakum and Roebber 2001; Roebber and Gyakum 2003) and more general climatologies (e.g., Stuart and Isaac 1999) and synoptic environments (e.g., Ressler et al. 2012). Ensemble analyses of such events, as is done here, are also important to identify processes by which model uncertainties may undermine winter storm forecasts over this region.

The numerical setup and method are detailed in section 2 and the event of interest is further described in section 3, along with a brief verification of the ensemble. The physical processes behind the precipitation-type uncertainty are studied in section 4, with particular focus on processes dictating the formation of freezing rain versus snow over Montreal. Section 5 presents the conclusions.

2. Ensemble simulations

a. Numerical model

The fully compressible, nonhydrostatic Advanced Research Weather and Research Forecasting (WRF-ARW, or WRF) Model, version 4.0, is used to simulate this event. A configuration consisting of three two-way nested grid domains, with horizontal grid spacings of 27 km (D01), 9 km (D02), and 3 km (D03), is used, with the finest domain (D03) positioned to contain most of the surface low pressure system as it approaches Montreal (Fig. 2). The simulations are run with 100 terrain-following vertical levels and are initialized at 1200 UTC on 23 January, 24 h prior to observed precipitation onset in Montreal.

Subgrid physical parameterizations include the Thompson microphysics scheme (Thompson et al. 2008), the Tiedtke cumulus convection scheme (Tiedtke 1989; Zhang et al. 2011), the Mellor–Yamada–Janjic eta turbulent kinetic energy (TKE)
boundary later scheme (Janjić 1994), the unified Noah land surface model (Chen and Dudhia 2001), and the Rapid Response Transfer Model for GCMs (RRTMG) radiation scheme (Iacono et al. 2008). Horizontal diffusion is handled using a Smagorinsky first-order closure scheme evaluated on model surfaces. Because the convective-scale grid spacing of the innermost domain allows for the explicit representation of deep convection, moist convection is only parameterized in the two outermost domains (D01 and D02).

Twenty-four ensemble members are integrated with perturbed initial conditions (ICs) and lateral-boundary conditions (LBCs). Although the ensemble size is modest, ensembles of similar or smaller size have been successfully applied to investigate IC/LBC uncertainties (e.g., Argence et al. 2008; Hanley et al. 2013). To focus on IC/LBC uncertainties herein, the sub-grid parameterizations are held fixed across the ensemble and no data assimilation is performed on the WRF grids. Twenty-one of the members are driven by the National Centers for Environmental Prediction (NCEP) Global Ensemble Forecasting System (GEFS) (Zhou et al. 2017) with 1° (approximately 70 km) horizontal grid spacing. It features a single control member and 20 perturbed members with initial states chosen from the 80-member ensemble Kalman filter produced by the Global Data Assimilation System (GDAS). The three additional ensemble members are driven by the North American Regional Reanalysis (NARR) (28-km horizontal grid spacing), the Global Forecasting System (GFS) analysis (0.5° ≈ 28-km horizontal grid spacing) and the ECMWF’s ERA5 reanalysis (28-km horizontal grid spacing). LBCs are updated every 6 h, but, given the large size of D01 and the short-range nature of the simulations, the ICs dominate the synoptic-scale uncertainty.

b. Precipitation-type diagnosis

To diagnose the precipitation type at a given location, we use the microphysics-dependent semi-explicit algorithm based upon the decision tree described by Benjamin et al. (2016), which is used in NCEP’s North American High Resolution Rapid Refresh regional weather prediction model. The output of this algorithm is grouped into three categories: snow (SN), freezing (FZ) (freezing rain and/or ice pellets), and rain (RN), with multiple categories permitted simultaneously. Despite the tendency for ice pellets to have lesser societal impacts than freezing rain, the decision to place both types in the same category is to focus on the presence or absence of a midlevel melting layer, which is required for both precipitation types.

3. The freezing rain event

a. Precipitation in Montreal

Precipitation was first reported at CYUL at 1200 UTC (0700 EST) and primarily fell in the form of freezing rain, occasionally mixed with ice pellets and snow grains, until about 1530 UTC. The freezing rain was accompanied by a slowly descending radar brightband at approximately 1.5 km above ground level, as indicated by an X-band vertically pointing radar at McGill University (Fig. 3a). This band indicates the melting of hydrometeors, as large snowflakes become covered in a layer of higher reflectivity liquid. Full melting of the frozen hydrometeor causes a reduction in size, and consequently reflectivity, accompanied by an increase in fall velocity (Fig. 3b).

After 1530 UTC, the melting layer gradually extended downward to the surface and then dissipated as the fall speed gradually decreased, indicating a transition from freezing rain to snow. A total of 11 mm of liquid-equivalent precipitation was observed at CYUL between 1200 and 1800 UTC.

With the nearest operational radiosonde site nearly 300 km to its northeast (Maniwaki, Quebec; Fig. 1), radiosondes are generally not available over Montreal. Therefore, we use the NARR, a model reanalysis product, as a proxy of the local environment. We choose NARR over other available model analysis/reanalysis products due to its focus on North America. Although comparison of NARR soundings with operational soundings from Albany (New York) and Maniwaki (see Fig. 1 for locations) indicates strong agreement (not shown), this agreement may stem from the NARR’s assimilation of these same soundings. Thus, the NARR soundings over Montreal are themselves uncertain and not necessarily more accurate than the WRF simulations. Nevertheless, comparison of the WRF results with an independent and widely used dataset helps to demonstrate the plausibility of the simulations.

At precipitation onset, the sounding at the NARR grid point nearest CYUL exhibits a thin layer of above-freezing air centered around 825 hPa (Fig. 4a). Despite veering winds and warm advection, this layer actually cools over the next 6 h to below zero (Figs. 4b,c), which may have resulted in part from hydrometeor melting—the self-limiting mechanism of Stewart (1985). While freezing rain was observed at CYUL at both 1200 UTC and 1500 UTC, the NARR soundings at these times exhibit very thin and/or nonexistent melting layers (Figs. 4a,b), more conducive to ice pellets or snow. Thus, the NARR likely underestimates midlevel temperatures over Montreal. Nevertheless, its midlevel cooling
signature is consistent with the observed transition from freezing rain to snow.

b. Synoptic overview

Despite their uncertainties, the NARR data are also used to characterize the synoptic evolution of the midlatitude cyclone. After forming in the lee of the Rocky Mountains (not shown), the surface cyclone tracked eastward over the south-central United States on 23 January, before moving east and north over the following 36 h along the east coast (Figs. 5a,c,e). At precipitation onset in Montreal, the cyclone was located just off the coast of New Jersey (Fig. 5c). It underwent weak intensification between model initialization and precipitation onset, with the minimum mean sea level pressure (MSLP) decreasing from 993.8 hPa at 1200 UTC 23 January to 992.4 hPa 24 h later.

ECCC’s Canadian Precipitation Analysis (CaPA) is used as an estimate of the 6-h precipitation accumulation (Figs. 5a,c,e). CaPA estimates precipitation using an amalgamation of weather radar data, rain gauges, and forecasts produced by ECCC’s Regional Deterministic Prediction System (RDPS), a limited area version of its Global Environmental multiscale (GEM) model. Estimates are available every 6 h on a grid with 10-km spacing covering most of Canada and the United States.

Fig. 3. (a) Reflectivity and (b) Doppler velocity measurements from the vertical pointing X-band radar located at McGill University during the freezing rain event on 24 Jan 2017. Precipitation was first recorded at 1200 UTC (0700 EST), and freezing rain was last reported between 1500 and 1600 UTC (1000–1100 EST).

Fig. 4. Skew T–logp profiles at grid points nearest CYUL for the NARR (black) at (a) 1200 UTC 24 Jan 2017, the time of precipitation onset at Montreal, and later at (b) 1500 UTC and (c) 1800 UTC of the same day. Overlaid are corresponding profiles for the WRF ensemble mean (purple) on D03, where each member is aligned such that +0 h corresponds to precipitation onset time at Montreal. Solid lines are temperature, and dashed lines are dewpoint. The 0° isotherm is highlighted in light blue. Each half-barb is 5 kt (1 kt ≈ 0.5 m s⁻¹).
Over the 12 h surrounding precipitation onset at Montreal, a band of heavy precipitation was located in the northern quadrant of the cyclone, migrating northward through the mid-Atlantic region, southern New England, and into southern Quebec. The CaPA 6-h precipitation at the grid point nearest CYUL (6.7 mm) is less than that observed at CYUL (11 mm) but similar to that from the McTavish automated recording station a few kilometers away (CWTA; 6.9 mm). Such point comparisons between CaPA and surface observations are compromised by the coarse grid cells of CaPA, which represent a spatial average over a much larger region.

At 500 hPa, a negatively tilted trough was located almost immediately above the surface low 6 h before precipitation onset in Montreal (Fig. 6a). Over the following 12 h, the trough advanced slowly to the northeast, causing the absolute vorticity over Montreal to increase (Figs. 6c,e). The cyclonic circulation carried warm, moist air northward downstream of the trough, aiding the formation of a midlevel melting layer over the SLRV, while the northeasterly surface flow down the SLRV helped to maintain a subzero surface layer. These two factors provided a suitable environment for sustained freezing rain in the Montreal region.

Ressler et al. (2012) identified common synoptic contexts for long-duration (>6 h) freezing rain events in the SLRV, classifying them according to the tilt and position of the 500-hPa trough and the presence and position of surface cyclones and anticyclones. Although the freezing rain event in this study did not qualify as “long duration” by their metric, the position and tilt of the 500-hPa trough bear a strong resemblance to the ring-like setup of a so-called “East” event [cf. Fig. 6c herein and Fig. 5c of Ressler et al. (2012)]. Such events are characterized by a neutral or negatively tilted 500-hPa trough and a surface cyclone, with freezing rain occurring in the classic warm front/occlusion region. Both Fig. 5c and Fig. 6c are consistent with

(Fortin et al. 2015).
this characterization. According to Ressler et al. (2012), East events also tend to be shorter-lived than other identified synoptic types due to the encroaching warm front inducing a transition to rain.

c. Ensemble verification

The verification of the ensemble is challenged by substantial ensemble variability in the time of simulated precipitation onset in Montreal, specifically at the model grid point closest to CYUL. Although the onset times are clustered around 1200 UTC (as observed), a maximal spread of 5 h is found across the ensemble. While daytime warming might be expected to result in more freezing precipitation in ensemble members with a later onset time, this is not the case. Only a weakly negative (statistically insignificant) correlation is found between the onset time and the percentage of freezing precipitation (not shown). Although variations in onset timing across the ensemble can be interesting, we perform all subsequent analysis relative to precipitation onset time to maintain a focus on ensemble variability of precipitation type. The timing of all members is thus aligned with precipitation onset for both the calculation of ensemble means in this section and the composite analysis in the next section.

The ensemble-mean cyclone evolution in WRF is similar to that in the NARR over the 12-h surrounding precipitation onset in Montreal (Fig. 5). For this comparison, we show WRF results on domain D02 and smooth both datasets to approximately 100-km resolution using a three-point bilinear multi-pass filter to focus on coherent synoptic-scale to mesoscale differences between the two realizations. At precipitation onset in Montreal, the cyclone center is located off the coast of New Jersey and is slightly deeper in WRF than in NARR, with minimum mean sea level pressures (MSLPs) of 989.7 and 992.4 hPa, respectively. This difference in strength is consistent throughout the simulation (and across the three different WRF domains) and can be traced back to model initialization, where the WRF cyclone had a minimum MSLP of 991.3 hPa, versus 993.8 hPa in the NARR (Fig. 7a). Thus, the simulated cyclones did not become stronger over the course of the model integration; they were simply stronger throughout. Consistently, at
500 hPa, the WRF and NARR troughs are similarly oriented but the former is slightly stronger (cf. Figs. 6b,d,f and 6a,c,e), a difference that also exists throughout the simulation and that traces back to model initialization (Fig. 7b).

The ensemble also produces a generally similar precipitation distribution to CaPA (Fig. 5). In both, the band of heaviest precipitation along the Atlantic coast migrates northward to reach southern Quebec by the onset time, with wraparound precipitation to the northwest of the low pressure center over Montreal over the first 6 h of the event. However, as compared with the single precipitation maximum along the coastline in NARR, the WRF ensemble mean tends to exhibit dual maxima, one offshore of that in NARR and the other inland in the wraparound region. The WRF precipitation in the latter region tends to exceed that in the NARR, including at Montreal over the first 6 h of the event. Thermodynamic implications of this apparent model precipitation overestimation over southern Quebec will be discussed shortly.

WRF ensemble-mean soundings at the grid point nearest CYUL (and aligned with precipitation onset) broadly agree with the corresponding NARR soundings, with some notable exceptions (Fig. 4). This analysis is performed on D03 for maximum local resolution. The WRF ensemble produces a shallower and colder surface-based mixed layer and a warmer overlying layer, both of which may stem from a less diffusive boundary layer parameterization and/or the neglect of latent-heat release owing to hydrometeor freezing at the surface (e.g., Lackmann et al. 2002). The nose of the simulated warm-frontal temperature inversion reaches 0°C at 850 hPa at precipitation onset, topped by a shallow, nearly isothermal layer (Fig. 4a). The maximum simulated temperature in this layer increases slightly and descends to around 875 hPa by +3 h before cooling to below 0°C by +6 h (Figs. 4b,c). In contrast, the top of the NARR frontal temperature inversion is already sub-freezing by +3 h (Fig. 4b), which, as mentioned in section 3a, is inconsistent with the observed freezing rain at that time. Thus, the midlevel temperatures may in fact be more accurate in the WRF ensemble than in the NARR.

Figure 8 shows the fractions of the cumulative precipitation amount at Montreal that fell as rain, ice, and snow in each ensemble member over the first 4 h of the event (again on D03). Averaged over the ensemble, 53% of the precipitation was freezing (freezing rain or ice pellets) and the rest was snow. To verify this result, we use surface measurements from CYUL, where manual precipitation-type observations were reported hourly but precipitation observations were only reported every 6 h. To estimate the freezing-precipitation fraction at CYUL, we assume a constant precipitation rate over the 6-h interval and an equal accumulation of all precipitation
types reported during each hour. The resulting freezing-precipitation fraction at CYUL (81%; Fig. 8) is larger than the corresponding ensemble-mean value. Although the observational estimate is uncertain, its higher freezing fraction could stem in part from a smaller observed precipitation accumulation (8.9 mm, the mean of CYUL and CWT, vs 12.5 mm in the WRF ensemble mean at Montreal) over the first 6 h of the event. The correspondingly weaker midlevel melting and diabatic cooling may have delayed the transition from ice to snow.

4. Physical mechanisms underlying ensemble variability

To estimate the spatial distribution of ensemble precipitation-type variability (between freezing types and snow) at the onset of precipitation in Montreal, we evaluate the number of ensemble members with mean 900–800-hPa wet-bulb temperatures $T_w$ above and below 0°C for all grid points with surface temperatures below 0°C. The 900–800-hPa layer is selected because it broadly outlines the melting layer in the Montreal soundings (Fig. 4). To highlight the locations of larger expected precipitation-type variability, we determine the fractions of ensemble members at each grid point on D03 with $T_w$ above and below 0°C. A value of 50% indicates maximal variability (one-half of the members have a midlevel melting layer and one-half do not). The results are shown in Fig. 9. Along with pockets of strong variability scattered across New England (mainly in coastal Maine and mountain valleys), a broad region of strong variability prevails in the lower SLRV, including Montreal. Given our specific interest in precipitation type in Montreal, we focus subsequent analysis on factors regulating midlevel temperatures over the latter region.

a. Composite analysis

For the 4-h period encompassing the freezing rain event, we rank the onset-time-aligned ensemble members according to their percentage of freezing precipitation (ice pellets and/or freezing rain) over Montreal. Two composites are formed from this metric, each consisting of four “extreme” members (beyond one standard deviation from the ensemble mean): members 4, 10, 14, and 19 form the freezing-precipitation composite (FZC) (freezing percentage: 86%–99%) and members 9, 18 and 22 form the snow composite (SNC) (freezing percentage: 0%–26%). The following analysis compares composite-mean fields to gain insight into the processes regulating precipitation type in the FZC and SNC. To focus on coherent and larger-scale differences over more ephemeral mesoscale features, these fields are horizontally smoothed to a resolution of approximately 100 km using the same procedure described in section 3c. Smoothing to finer resolutions was also explored, but large variability at smaller scales undermined efforts at meaningful interpretation.

MIDLEVEL TEMPERATURE VARIATIONS

To explain the evolution of precipitation type during the event, we compare FZC and SNC composite soundings over the first 6 h of precipitation at Montreal (Fig. 10). At precipitation onset, a 900–800-hPa temperature difference is apparent between the FZC and SNC, with the former being 1°C–2°C warmer (Fig. 10). These two temperature profiles straddle the 0°C isotherm over 875–775 hPa, with the FZC above it and the SNC below it. Over the subsequent 6 h, the FZC melting layer erodes until the midlevel temperatures fall below 0°C in both composites. The Spearman rank correlation between the percentage of freezing precipitation during the event period and the air temperature at pressure levels between 900 and 750 hPa, evaluated over the full ensemble, indicates the strongest correlation at the 850-hPa level (correlation coefficient $R = 0.89$). This pressure level is thus selected for further analysis. All subsequent maps show D02 (rather than D03) to provide larger-scale context for interpreting local changes near Montreal.

At precipitation onset, differences between the FZC and SNC for the 850-hPa potential temperature $\theta$ are not limited to Montreal or the SLRV but rather extend over a broad area to the south and southeast of Montreal (Fig. 11). These differences in prefrontal midlevel temperatures were not apparent at model initialization (not shown), implying that they formed during the model integration. The development of the midlevel warm anomaly in the FZC over the 6 h prior to precipitation onset is shown by the FZC–SNC composite $\theta$ differences in Figs. 11d–f. Over this time, the prefrontal warm anomaly in the FZC grows from a small and broken region to the south of Montreal, to a much broader, more coherent, and stronger anomaly that covers Montreal (Figs. 11e,f).

Similarly, vertical cross sections across the warm front (see locations in Figs. 11d–f) show a clear warm anomaly in the FZC developing mainly in the 3 h before precipitation onset at Montreal. At 6 h before onset, the $\theta$ surfaces generally tilt
upward from south to north, with the strongest surface baroclinicity at around 41°N (Fig. 12a), and FZC–SNC composite \( \theta \) differences are limited to scattered prefrontal warm pockets in the FZC. As the front advances slowly northward over the next 6 h, the FZC develops a stronger, larger, and more coherent prefrontal warm anomaly at midlevels, particularly between 42° and 46°N and between 800 and 900 hPa at onset (Figs. 12b,c). This difference coincides with broadly strengthening cross-frontal (southerly) flow and weaker midlevel ascent, the latter particularly

![Figure 10](image-url)  
**Fig. 10.** Comparison of simulated FZC (red) and SNC (blue) mean temperature (solid) and dewpoint (dashed) profiles, at Montreal on D03 (a) at precipitation onset, (b) 3 h after onset, and (c) 6 h after onset.

![Figure 11](image-url)  
**Fig. 11.** Ensemble-mean \( \theta \) (shading), \( Z \) (dam; black contours), and wind vectors at 850 hPa at (a) 6, (b) 3, and (c) 0 h prior to precipitation onset in Montreal on D02. (d)–(f) Composite differences (FZC − SNC) in \( \theta \) (shaded) and winds at 850 hPa, along with ensemble-mean \( \theta \) at 850 hPa (black contours), at the same three times. For the winds, each half-barb is 5 kt. In (a)–(c), the approximate position of the warm front at 850 hPa is marked by the thick dashed line, and in (d)–(f) the dashed black line shows the position of the vertical cross section shown later in Fig. 12.
The differences in the vertical advection are dominated by differences in vertical motion, as seen by the similar southerly winds in the FZC at 850 hPa (Figs. 11e,f and 12b,c). Given the ongoing warm advection evident in the ensemble mean (Figs. 11), and that this southerly flow has a component along the midlevel $\theta$ gradient, this wind difference serves to enhance the 850-hPa horizontal warm advection in the hours prior to precipitation onset (Figs. 11a–c).

As might be expected for a quasigeostrophic atmosphere, the above differences in midlevel winds relate to differences in geopotential height $Z$. The FZC exhibits a dipole of higher MSLP and $Z$ just to the southeast, and lower MSLP and $Z$ to the northwest, of the surface cyclone and 850-hPa trough, respectively (Figs. 15b,c). Although this pattern suggests a northward shift in system position, it is shallow and does not reach 500 hPa (Fig. 15a). However, the area of larger MSLP and $Z$ on the system’s downstream (eastern) flank is not confined to the southeastern dipole—it stretches northward to cover most of the eastern flank. This latter feature, which extends upward to 500 hPa and beyond, indicates a narrowing of the downstream flank of the system in the FZC. The correspondingly stronger zonal pressure gradients align with the stronger southerly winds into Montreal (cf. Figs. 15b and 11f). Another notable feature is an area of lower MSLP and $Z$ over the U.S. Great Plains well upstream of the system in the FZC (Figs. 15a–c). Because this distant feature is unlikely to directly impact precipitation type over Montreal, we do not explore it in detail.

Differences in the structure of the downstream flank of the 500-hPa trough can be traced back to model initialization, where the trough is weaker and narrower along its eastern edge in the FZC (Fig. 15d). Both the 850-hPa trough and surface cyclone are also generally weaker in the FZC at initialization, with slightly stronger gradients to their northeast (Figs. 15e,f). Although differences in the depth of the trough vanish over the course of the simulation (Fig. 7b), the stronger gradients along the eastern flank, and the corresponding enhancements in southerly geostrophic winds, persist throughout.

The differences in the vertical advection are dominated by differences in vertical motion, as seen by the similar

![Figure 12](image-url)

**Fig. 12.** Ensemble-mean $\theta$ (black contours), FZC – SNC $\theta$ differences (shading), and FZC – SNC plane-parallel wind differences (barbs) in the cross-frontal direction (see Figs. 11d–f for locations) at three different times: (a) 6 h before precipitation onset, (b) 3 h before precipitation onset, and (c) precipitation onset, on D02. Wind barbs show the magnitude of wind differences (each half-barb is 0.5 kt) and their direction. For the barb direction, the vertical wind is amplified by a factor of 100 to clearly show the sense of both components. The location of Montreal is indicated by the dashed green line.

evident to the north of 44°N (including Montreal) and extending over 550–900 hPa.

Based on the thermodynamic energy equation,

$$\frac{\partial \theta}{\partial t} = -\left( \mathbf{v}_h \cdot \nabla \theta + \mathbf{w} \frac{\partial \theta}{\partial Z} \right) + J,$$

(1)
two processes are responsible for the local evolution of $\theta$: 3D advection and diabatic processes (including radiation, cloud latent heating/cooling, and boundary layer heating/cooling). In the preceding, $\mathbf{v}_h$ is the horizontal wind vector, $w$ is the vertical velocity, and $J$ is the sum of the $\theta$ tendencies of all diabatic processes. Evaluation of (1) in height coordinates (and interpolating to pressure surfaces) indicate that the differences in the advection term (Figs. 13a,c,e) dominate those in the diabatic term (Figs. 13b,d,f) in the area of developing midlevel $\theta$ differences (Figs. 11d,e,f).

As discussed in section 3c, precipitation rate can impact precipitation type by regulating midlevel melting and diabatic cooling. However, this mechanism does not appear to drive the composite $\theta$ differences seen in Fig. 11. The diabatic differences between the two composites in the upstream flow southeast of Montreal change sign over time and cannot explain the size and magnitude of the midlevel warm anomaly downstream (Figs. 13b,d,f). Although the precipitation rate correlates negatively with freezing-precipitation fraction over the ensemble ($R = -0.37$), this correlation is insignificant at the 95% confidence level, suggesting only a weak sensitivity. Moreover, the difference in Montreal precipitation between the composites is minimal over the first 4 h of the event (around 10%) and thus unlikely to explain their substantial differences in midlevel $\theta$.

Both the horizontal and vertical components of the advection term contribute to the FZC warm anomaly at 850 hPa around Montreal (Figs. 14c–f), with the former providing modest warming in a narrow meridional corridor extending through Montreal and the latter providing higher-amplitude warming over a broader region. This increased horizontal $\theta$ advection coincides with a meridional band of stronger
difference patterns of the two fields (but with opposite polarity; Fig. 16). Of particular interest is the wraparound region of weaker vertical advection in the FZC at precipitation onset over Montreal and neighboring regions (Fig. 16a). Within this region, vertical and horizontal advection sum constructively to produce a positive \( u \) anomaly in the FZC (Figs. 14e,f). This region also partially coincides with a broad area of large precipitation-type variability (Fig. 9). Because this difference only appears immediately before precipitation onset in Montreal, we focus on this time to understand its origin.

Given that the 850-hPa flow near Montreal is absolutely stable (Figs. 4a and 10a), ascent produces cold advection that exceeds the diabatic heating owing to latent heat release. Thus, areas of stronger midlevel ascent in the SNC (blue regions in Figs. 16b) lead to greater cooling, and consequently a relative decrease in midlevel \( \theta \). To explain the locally stronger ascent in the SNC over Montreal, we use the traditional form of the quasigeostrophic (QG) omega equation (e.g., Holton 2004, 139–181):

\[
\omega = \left( \nabla^2 + f_0 \frac{\sigma^2}{\sigma p} \right) \omega = \frac{f_0}{\sigma} \frac{\partial}{\partial \phi} \left[ \theta - \frac{f_0}{\sigma} \nabla^2 \phi + f \right] + \frac{\nabla^2}{\sigma} \left[ \theta - \frac{f_0}{\sigma} \nabla^2 \phi \right] - \frac{\kappa}{\sigma p} \nabla^2 f
\]

where \( \omega \) is the pressure velocity, \( \sigma \) is the static stability parameter, \( \phi \) is the geopotential, \( f \) is the Coriolis parameter with \( f_0 = 10^{-2} \text{ s}^{-1} \) being the Coriolis parameter at 45\(^\circ\)N, and \( \kappa = 0.285 \) is the Poisson constant for dry air. The three contributions to vertical motion in (2) are the differential advection of absolute geostrophic vorticity (DVA), the Laplacian of diabatic heating, and the Laplacian of the Coriolis parameter. The right side of Fig. 13 demonstrates this physics, with contour plots covering the period from 6 h to 0 h before precipitation onset in Montreal.
horizontal temperature advection (LTA), and diabatic heating. Given the 3D Laplacian-like operator on the left-hand side, the sum on the right-hand side (RHS) tends to correlate negatively with $\omega$. However, the precise relation between the RHS and $\omega$ can only be determined by solving (2) (e.g., Kirshbaum et al. 2018). For the present purposes, we find it sufficient to qualitatively inspect the RHS terms, which are calculated by interpolating the WRF data to pressure levels and computing derivatives using second-order centered differencing.

At precipitation onset, a positive LTA difference over Montreal (Fig. 17c) prevails due to the aforementioned stronger warm advection in the FZC (Fig. 14e). While this forcing in isolation would tend to produce stronger ascent over Montreal, it is offset by the negative DVA term in the SLRV region (Fig. 17b). In particular, the DVA term shows a similar structure as the wraparound region of weaker ascent in the FZC, extending westward from the frontal baroclinic zone toward the SLRV (Fig. 16b). In this region, both the SNC and FZC exhibit negative DVA (not shown), but the FZC has a larger magnitude. This locally more negative DVA term appears largely responsible for the weaker SLRV ascent in the FZC. The diabatic term is relatively weak in the Montreal area and negative to the west, consistent with the weaker ascent in that region (Fig. 17d).

The sum of the RHS difference terms is slightly negative over Montreal (Fig. 17a), consistent with the locally weaker ascent in the FZC. Thus, despite the limitations of the QG approximation and our decision not to solve (2), the QG theory still provides useful qualitative guidance on the larger-scale processes regulating ascent in the SLRV. Namely, despite the stronger warm advection over Montreal in the FZC, the vertical motion is locally weaker due to the larger offsetting DVA term.

To explain the more negative DVA term in the FZC composite in the Montreal region, we first expand it in (2) via the chain rule of differentiation:
with term A being the advection of absolute vorticity by the thermal wind and term B being the advection of the differential absolute vorticity. At precipitation onset, the composite differences within the SLRV are stronger in term B than in term A in the Montreal region, with term B explaining the more negative DVA in the FZC (cf. Figs. 18a,b). Thus, term B is largely responsible for the negative differences in the DVA term in (2) in the SLRV (Fig. 17a).

Further decomposition of term B into its zonal and meridional components,

\[
\frac{v_x}{\sigma} \cdot \nabla \left( \frac{\partial \psi^2 \phi + f}{\partial p} \right) + \frac{v_y}{\sigma} \cdot \nabla \left( \frac{\partial \psi^2 \phi}{\partial p} \right),
\]

indicates that, at precipitation onset, the two components generally exhibit a large degree of cancellation throughout the domain (Figs. 18c,d). However, at CYUL and its immediate

![Composite differences](image-url)

**FIG. 15.** Composite differences (FZC − SNC) in (a) 500-hPa \(Z\), (b) 850-hPa \(Z\), and (c) MSLP at precipitation onset and (d) 500-hPa \(Z\), (e) 850-hPa \(Z\), and (f) MSLP at initialization on D02. Black contours show ensemble-mean \(Z\) at appropriate pressure levels in (a),(b),(d),and (e) and ensemble-mean MSLP in (c) and (f). Composite differences in 850-hPa geostrophic winds in (b) are indicated by black wind barbs, with each half barb representing 5 kt.

![Composite differences](image-url)

**FIG. 16.** Composite differences (FZC − SNC) in (a) vertical \(\theta\) advection (as in Fig. 14f) and (b) vertical velocity at 850 hPa at precipitation onset in Montreal on D02. Black contours show ensemble-mean \(\theta\) values, and black barbs show difference winds, with each half barb representing 5 kt, at 850 hPa.
surroundings, the negative difference in the meridional component of term B is 2–3-fold larger than the corresponding differences in the zonal component. According to (4), the negative composite differences in the meridional term must be associated with a more positive meridional geostrophic wind or a more negative meridional vorticity gradient in the FZC.

Although not shown, the composite differences in the meridional component of the differential vorticity gradient in the Montreal region are small (only 6% of the background value) and more positive (i.e., smaller magnitude) in the FZC. In contrast, the variation in $v_g$ is approximately 30% of the background value, which is of the same order as the difference in the meridional component of term B, and of greater magnitude (i.e., more positive) in the FZC than in the SNC. Thus, the differences in $v_g$ must be largely responsible for the negative differences in term B of the DVA, and of greater magnitude (i.e., more negative) in the FZC than in the SNC. Consequently, these 850-hPa $v_g$ differences, which were found to stem from a narrower eastern flank of the upper-level trough, were also responsible for variations in the horizontal $\theta$ advection ($\approx$LTA).

b. Ensemble sensitivity analysis

The composite sensitivity analysis of section 4a suggested a causal mechanism linking the variations in ensemble precipitation type to IC/LBC uncertainties: a narrower eastern flank of the mid-to-upper trough in the FZC is associated with stronger southerly geostrophic winds, higher midlevel temperatures, and thus a larger fraction of freezing precipitation at Montreal. The generalization of this sensitivity to the entire ensemble is probed using ensemble sensitivity analysis (ESA) (Ancell and Hakim 2007), a linear regression technique that determines the change in a chosen metric of interest to perturbations in a selected field (e.g., $Z$ on a given pressure surface) over the ensemble. ESA has previously been used to study sensitivities of precipitation metrics such as timing, location, and accumulation to perturbations in the initial conditions (e.g., Hanley et al. 2013; Bednarczyk and Ancell 2015; Greybush et al. 2017), and here we apply the technique to ensemble precipitation-type variation.

We implement ESA following the approach of Garcies and Homar (2009). At every grid point, the raw sensitivity $S$ of a chosen metric $J$ to perturbations in a field $x$ is given by $S = \partial J/\partial x$. The raw sensitivity is then multiplied by a correction factor $R$, which is included to reduce the signal of grid points for which the assumption of linear regression breaks down. The value of $R$ is calculated at each grid point as $R = r^2/c^2$ for $r^2 < c^2$ and $R = 1$ otherwise, where $r$ is the correlation coefficient between $J$ and $x$ at that grid point and $c$ is a constant chosen to represent the minimum correlation coefficient for which sensitivities remain unchanged. Following Garcies and Homar (2009), we calculate $c = 0.47$ as the 88th percentile of the covariance distribution for all variables and levels. As a final step, we multiply by the standard deviation of the field at each grid point $\sigma$. The resultant sensitivity, $S = R\sigma \partial J/\partial x$, is the sensitivity of the metric to a perturbation of typical magnitude in the field, in the same units as the metric itself.
Because the surface precipitation type over the event is largely controlled by midlevel temperature, the 850-hPa \(u\) at precipitation onset, averaged over the area enclosed by the green-outlined box in Fig. 19a, is selected as the metric. This area encompasses the prominent differences in \(u\) between the FZC and the SNC (Fig. 11f) as well as the region of large precipitation-type variability (Fig. 9). The use of 850-hPa \(u\) as the metric, rather than some estimate of freezing-precipitation fraction, avoids uncertainties associated with precipitation-type diagnoses.

The sensitivity of \(J\) to various relevant fields was evaluated, and the strongest coherent sensitivities were tied to the mid-to-upper-level \(Z\) and MSLP fields. The importance of the midlevel trough position and structure at precipitation onset is reinforced by the 850-hPa \(Z\) ensemble sensitivity field (Fig. 19a), which shows positive values along the eastern flank of the trough and negative values along its western flank. This dipole pattern suggests that westward shifts in the 850-hPa \(Z\) position, which give higher \(Z\) to the east of the trough and lower \(Z\) to the west, are associated with warmer midlevel flow over Montreal. Importantly, however, this dipole is zonally asymmetric, with larger values along the eastern flank of the trough than on the western flank. This stronger sensitivity on the eastern flank suggests that an additional factor—the strength of \(Z\) gradients along that flank—is also critically important.

The latter effect can be traced back to model initialization (Figs. 19b,c,d), suggesting that the stronger \(Z\) and pressure gradients along the eastern flank of the upper-level trough and cyclone, respectively, persist throughout the FZC simulations. Through its enhancement of the midlevel southerly geostrophic winds, this difference gives rise to higher midlevel \(\theta\) and, hence, a greater fraction of freezing precipitation, around Montreal. Although the composite analysis also suggested a sensitivity of precipitation type to the initial depth of the upper-level trough (recall Fig. 15), the absence of ensemble sensitivity near the core of the trough in Figs. 19b and 19c does not support that contention. Thus, we conclude that sharper initial \(Z\) gradients along the eastern flank of the upper-level trough was the key factor regulating the ensemble variability of the precipitation type.

Although our composite analysis of section 4a did not distinguish between ice pellets and freezing rain in the FZC, differences in freezing rain and ice-pellet fractions across the ensemble likely relate to the same robust dynamical factors underlying the differences in midlevel \(\theta\) explained above. Given that the type of freezing precipitation is mainly determined by the strength and depth of the midlevel melting layer (e.g., Zerr 1997), ice pellets can be considered an intermediate precipitation type between freezing rain (warmest midlevel flow) and snow (coldest midlevel flow). Thus, the same dynamical processes that distinguish freezing precipitation from snow likely also distinguish freezing rain from ice pellets. In the event studied herein, larger freezing rain fractions likely correspond to the strongest southerly midlevel flows over Montreal, with progressively weaker flows for ice pellets and then snow.

![FIG. 18. Composite differences (FZC – SNC) between (a) term A and (b) term B of the DVA term of (2), and (c) the zonal component and (d) meridional component of term B, at 850 hPa during precipitation onset in Montreal on D02. Black contours show corresponding ensemble-mean \(\theta\), and wind barbs show difference winds, with each half barb representing 5 kt, at 850 hPa.](image-url)
5. Conclusions

The ensemble sensitivities of precipitation type to initial-condition uncertainty, and the physical mechanisms behind these sensitivities, have been investigated using a 24-member convective-scale ensemble of a recent freezing rain event in Montreal. This event occurred in the context of a midlatitude cyclone traversing the east coast of North America on 24 January 2017, and it was notable for large operational-forecast errors in precipitation type. The operational forecast from Environment and Climate Change Canada predicted 15–25 cm of snow and, while 12.4 mm of liquid-equivalent precipitation fell at Montreal International Airport, only 7 cm of snow was reported. The majority of the precipitation during the early stages of the event instead fell unexpectedly as freezing rain.

The event was simulated using the WRF-ARW Model with three nested domains, including a convection-permitting inner domain with 3-km grid spacing that covered Montreal and the core of the midlatitude cyclone during the event. The majority of the simulated precipitation over Montreal in the first 4 h of the event was diagnosed as either freezing rain or ice pellets, in approximate agreement with observations. However, the precipitation type varied considerably among the ensemble members, with some members producing exclusively freezing rain and ice pellets over this period and some producing primarily snow. The processes responsible for this variation were studied by creating two composites: a freezing composite comprising the three members with the highest percentage of freezing rain and ice pellets (FZC) and a snow composite comprising the three members with the highest percentage of snow (SNC).

The composites differed in midlevel (850 hPa) temperatures over Montreal, with the FZC approximately 1°C warmer at the nose of a warm-frontal inversion at around 850 hPa. The midlevel temperatures in the two composites straddled the 0°C isotherm, with the FZC above it and the SNC below it. Thus, precipitation in the FZC was more prone to midlevel melting and, ultimately, refreezing on its descent toward (or upon contact with) the surface.

The composite temperature differences were controlled by differences in horizontal and vertical θ advection at midlevels, with the FZC exhibiting both stronger horizontal warm advection and weaker vertical cold advection, each contributing to larger θ ahead of the warm front. These two effects ultimately stemmed from the same factor: enhanced midlevel southerly flow across the frontal baroclinic zone in the FZC. This factor was responsible not only for increasing horizontal warm advection along and ahead of the warm front in the FZC, but also for enhancing negative differential-vorticity advection ahead of the front. From a quasigeostrophic perspective, the latter effect countered the tendency for larger horizontal warm advection to produce stronger ascent, leading to slightly weaker ascent and less vertical cold advection in the FZC.

The enhanced midlevel southerly flow, and hence the warmer midlevel air, in the FZC was associated with a narrowing of the downstream (eastern) flank of the upper trough and cyclone, which promoted stronger geostrophic southerly
winds into Montreal. This difference was a coherent and synoptic-scale feature that, at precipitation onset, extended from the southeastern United States into southern Quebec. Using ensemble compositing in conjunction with ensemble-sensitivity analysis, the stronger gradients along the downstream flank of the trough were traced back to the model initialization. Although the FZC was also found to exhibit a weaker trough at model initialization, this difference vanished during the model integration, leaving the width of the trough leading edge as the dominant source of uncertainty in precipitation type over Montreal and nearby regions.

The findings herein underscore the importance of modest initial-condition uncertainties in dictating local precipitation types. Although small-amplitude errors in the initial strength and position of upper-level features are unavoidable, their potentially large effects on mesoscale and convective-scale precipitation processes can only be fully exposed through the use of high-resolution ensembles. Because the results herein point to just one of potentially many multiscale mechanisms causing local variability in precipitation type, further study of a wider variety of cases is needed. The analyzed event bears strong resemblance to an “East” event [according to the synoptic subtyping of Ressler et al. (2012)], which was hypothesized to have lower predictability than slower-moving “West” events. Thus, studies covering both types of events would provide a more thorough characterization of the precipitation-type predictability for common synoptic disturbances affecting this region. Likewise, while we have focused on understanding the mesoscale variability arising from synoptic-scale initial-condition uncertainties, the sensitivity of winter-precipitation type to model physics and grid resolution constitutes an important avenue for future work.

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Data availability statement. The raw WRF output generated herein, as well as the codes used to generate the figures, are available from the authors upon request (daniel.tootill@mail.mcgill.ca). Observational data from surface stations were freely obtained from MesoWest (https://mesowest.utah.edu), the radar was freely obtained from the Marshall Radar Observatory (http://www.radar.mcgill.ca), and NARR data were freely obtained from NOAA’s Physical Sciences Laboratory (https://psl.noaa.gov/data/gridded/data.narr.html).

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