Topographic Impacts on the Spatial Distribution of Deep Convection over Southern Quebec

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ABSTRACT

Observations and numerical simulations reveal pronounced mesoscale variability in deep-convection occurrence over southern Quebec, Canada. A 22-yr climatology from the McGill radar just west of Montreal shows that deep-convection maxima exist (i) within the St. Lawrence valley surrounding Ottawa; (ii) within the Champlain valley of upstate New York, extending north to just east of Montreal; and (iii) in the lee of the Laurentian Mountains northeast of Trois-Rivières. These features are sensitive to the background low- to midlevel geostrophic wind direction, shifting northward as the southerly wind component increases. A meridional axis of suppressed convection also extends from Lake Ontario and the Adirondacks of New York north through Montreal and into the Laurentians. To physically interpret these features, a suite of quasi-idealized convection-permitting simulations is conducted. Analysis of the simulations, which broadly reproduce the observed extrema in convection occurrence, reveals that the maxima develop within pockets of moisture and mass convergence at the junctions of major river valleys and in the lee of prominent mountain ridges. In these locations, enhanced boundary layer humidity and convective available potential energy (CAPE) coincides with minimal convective inhibition (CIN). The minima occur over and downwind of water bodies, where limited surface heat fluxes reduce CAPE and increase CIN, and over the higher terrain, where reduced low-level moisture limits storm intensity.

1. Introduction

Convective-storm occurrence may vary profoundly in areas of complex topography. Over mountains, for example, forced ascent of impinging flow and/or elevated heating drive vertical ascent, which, along with elevated flow destabilization, may initiate convection more readily than over the flatter surroundings (e.g., Banta 1990; Weckwerth et al. 2011; Soderholm et al. 2014). In a summer radar-echo climatology of the Black Hills range of South Dakota and Wyoming (United States), Kuo and Orville (1973) found a climatological maximum in echo frequency over the lee slopes, relative to the mean wind direction above 5 kft (~1.5 km). In southwesterly (northwesterly) flow, the maximum was located on the northeastern (southeastern) side of the mountain ridge. Wasula et al. (2002) also found severe weather and lightning occurrence in New York State and western New England to depend on the terrain and the midlevel wind direction. Storm density was generally maximized within the north–south-oriented Hudson Valley near Albany, New York, with the precise location and structure of the relative maximum varying with the 700-hPa winds. By contrast, deep convection was minimized over the Catskill and Adirondack Mountains.

Land-use variations, including land–water contrasts, natural variations in vegetation cover, agriculture, and urbanization, also strongly influence deep convection (e.g., Sills et al. 2011; Robinson et al. 2011; Kirshbaum et al. 2016). These variations give rise to spatial differences in surface friction, which create variations in horizontal convergence and, hence, low-level vertical motion. They also modulate the surface energy balance, causing horizontal gradients in static stability and buoyancy, the latter driving thermal circulations with potentially strong updrafts. In general, soils that are darker, drier, and/or lower in specific heat capacity produce larger sensible heat fluxes, deeper convective boundary layers, and thermally direct ascent, all of which facilitate convection initiation (e.g., Souza et al. 2000; Roy and Avissar 2002; Garcia-Carreras and Parker 2011). Over wetter and/or...
densely vegetated surfaces, an increased fraction of solar radiation is used for evaporation, which increases the latent heat flux at the expense of the sensible heat flux. While slowing the removal of convective inhibition, this effect increases the storage of moist instability, which can ultimately give rise to more explosive convection (e.g., Barthlott and Kalthoff 2011).

Numerous previous studies have examined the impacts of urbanization on deep convection, which largely arises from the urban heat island (UHI) effect (e.g., Changnon 1981; Rozoff et al. 2003; Niyogi et al. 2011; Haberlie et al. 2015). In general, convection initiation is enhanced downwind of cities, particularly over otherwise homogeneous terrain where competing topographic impacts are minimized. During the daytime, the UHI largely arises from locally enhanced sensible heat fluxes over impermeable paved surfaces. The resulting warming causes a local doming of the convective boundary layer, destabilizing the flow over the city and downwind. Such thermodynamic effects are complemented by the dynamical tendency for horizontal convergence and ascent to develop downwind of localized heat sources (e.g., Han and Baik 2008; Kirshbaum 2013). While the UHI strengthens at night because of the larger heat capacity of urban surfaces, the UHI-induced ascent is still maximized during the daytime (Bornstein and Lin 2000), where reduced boundary layer stability promotes deeper and stronger thermal circulations.

Southern Quebec—namely, the region surrounding Montreal—in Canada exhibits large mesoscale variability in both terrain height and land use (Fig. 1). Located at the center of the St. Lawrence River valley (SLRV), Montreal is surrounded by the Laurentian Mountains to the northwest, the Adirondacks of New York to the southwest, and the Green Mountains to the east (Fig. 1d). Although these mountains are relatively modest, with peaks in the 0.5–2-km range, they may still profoundly influence the regional distribution of convective storms. According to the U.S. Geological Survey (USGS) land-use database, central Montreal is classified as urban land use, and its immediate surroundings are a combination of irrigated and dryland cropland and pasture (Fig. 1a). To the north the land use is predominantly mixed (deciduous broadleaf and needleleaf) forest, and to the south it is deciduous broadleaf forest with embedded irrigated cropland and pasture. These contrasting surfaces differ in their climatological albedo (Fig. 1b) and soil-moisture content (Fig. 1c), causing significant mesoscale variation in the surface energy balance.

Bellon and Zawadzki (2003, hereinafter BZ03) performed a 9-yr summer climatology (1993–2001) of mesocyclones and severe-convection frequency over southern Quebec, using Doppler radar data from McGill University’s Marshall Radar Observatory. Their Fig. 8 suggests that, over the period of record, severe convection was the most frequent in three regions: (i) a meridional axis extending from the Champlain valley north into the SLRV to points east of Montreal, (ii) a broad area within the western SLRV surrounding Ottawa, and (iii) along the northwestern edge of the SLRV at the foot of the Laurentians, north of Trois-Rivières. By contrast, a broad half-hourglass-shaped meridional minimum extended from the Adirondacks of New York State north through Montreal and into the Laurentians.

Firm conclusions cannot be drawn from BZ03’s climatology because its statistical sampling was limited by its short duration and its focus on severe convective storms, which occur relatively infrequently in southern Quebec. Moreover, BZ03 did not propose any physical hypotheses for the observed patterns of deep-convection occurrence. To translate such findings into improved operational forecasting, and to generalize the results to other regions, a longer observational climatology—and a deeper physical understanding of it—are needed. To this end, we conduct an observational and numerical study of deep convection over southern Quebec. We focus specifically on the impacts of the regional topography (terrain and land-use/land-cover variations), which we hypothesize to strongly influence the regional convection signature. The observational part (section 2) extends BZ03’s climatology to 2014, more than doubling its length. The numerical part (section 3) uses quasi-idealized simulations to gain insight into the topographic processes that influence regional convection. Section 4 presents a discussion of the results, and section 5 contains the conclusions.

2. Observations

The principal instrument used for the observational analysis is the McGill radar, located at the Marshall Radar Observatory (MRO) in Ste. Anne de Bellevue, Quebec (30 km west of downtown Montreal). It is the largest S-band (10-cm wavelength) Doppler and dual-polarization weather radar in Canada and is one node of the Meteorological Service of Canada’s operational precipitation monitoring network. The scanning strategy consists of 24 elevation angles (0.5°–34.4°) separated into two volumetric scans of 12 odd and even angles, each lasting 2.5 min for a total of 5 min, the length of one cycle. The radar range used herein is 240 km with a resolution of 1 km up to 120 km and 2 km between 120 and 240 km.

Objective identification of deep convection by scanning radar generally requires the use of a prescribed threshold to distinguish convective from stratiform precipitation. This threshold is often obtained from radar reflectivity Z, using data from the lowest elevation

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angle or the maximum column reflectivity (e.g., Johnson et al. 1998; Soderholm et al. 2014). However, such analysis is prone to error because of anomalously high $Z$ within ground clutter and the radar bright band, and anomalously low $Z$ in areas affected by beam blocking and attenuation. To minimize such adverse effects, we compute the vertically integrated liquid water content (VIL). Assuming pure liquid precipitation with an exponential raindrop size distribution, the rainwater content $M$ is proportional to $Z^{4/7}$ (Greene and Clark 1972). Integrating $M$ over a vertical layer (from $h_{\text{bot}}$ to $h_{\text{top}}$), we obtain the VIL:

$$VIL = \int_{h_{\text{bot}}}^{h_{\text{top}}} M \, dz = 3.44 \times 10^{-6} \int_{h_{\text{bot}}}^{h_{\text{top}}} Z^{4/7} \, dz. \quad (1)$$

Although the above assumptions behind (1) are questionable, the VIL still represents a useful first-order proxy for storm intensity. As convection intensifies, its updraft strength, depth, and ability to suspend hydrometeors increases, which increases VIL regardless of the phase of the water substance. Previous studies have used VIL for various purposes including storm detection and tracking (Johnson et al. 1998) and hail detection (Amburn and Wolf 1997). Note that a blind spot is found

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**Fig. 1.** USGS topography of southern Quebec and northern New England at 1-km grid resolution during the summer. (a) Land-use index, with land-use types described in the legend; (b) surface albedo; (c) volumetric soil moisture, with Canadian rivers and lakes overlaid; and (d) terrain height. The cities of Montreal, Ottawa, Trois-Rivières, and Burlington, Vermont, are shown by asterisks in all four panels and are labeled in (b).
over a ~20-km radius surrounding the radar site because the shallow elevation angles of the radar volume scans do not reach the cloud top. This blind spot is henceforth masked to avoid misinterpretation.

**Table 1.** Periods of major radar downtime (exceeding 48 h) over the 1993–2014 McGill radar climatology. The start and end dates indicate the full down period, and the “hours lost” column includes just the downtime on summer afternoons (1700–0000 UTC June–August).

<table>
<thead>
<tr>
<th>Start date</th>
<th>End date</th>
<th>Hours lost</th>
</tr>
</thead>
<tbody>
<tr>
<td>28 Apr 2003</td>
<td>15 Aug 2003</td>
<td>532</td>
</tr>
<tr>
<td>12 Jul 2007</td>
<td>17 Jul 2007</td>
<td>42</td>
</tr>
<tr>
<td>3 Apr 2008</td>
<td>12 Sep 2008</td>
<td>644</td>
</tr>
<tr>
<td>1 Aug 2010</td>
<td>14 Aug 2010</td>
<td>105</td>
</tr>
</tbody>
</table>

To create the climatology, we compute the fraction of time \( f(x, y) \) where EVIL exceeds a prescribed threshold at each horizontal grid point in the radar scan. Two such thresholds are considered: \( \alpha_{15} = 15 \text{ kg m}^{-2} \) (as in BZ03) and \( \alpha_5 = 5 \text{ kg m}^{-2} \). While the former detects more intense convective cells, the latter detects most, if not all, deep convection and thus provides more robust sampling. The numerator \( n(x, y) \) of \( f \) is initially set to zero at all grid points. As the algorithm chronologically steps through each radar image over the period of record, \( n \) is incremented by one at each grid point where EVIL(x, y) > \( \alpha \). The denominator \( d(x, y) \) is simply the total number of radar images (excluding downtime) over the same period. For consistency, \( d \) is held fixed throughout the observational analysis, so that \( f \) always represents a fractional occurrence over the total climatology.

Figure 2 presents \( f \) for the two above-mentioned EVIL thresholds (\( f_{15} \) and \( f_5 \), respectively). These fractions are generally small (not exceeding \( 6 \times 10^{-4} \) within the scan area) because deep convection over this region is uncommon and intermittent. Since \( f_{15} \) identifies only the deepest and most intense cells, its statistical sampling is small and, as a result, its distribution is noisy (Fig. 2a). Nonetheless, this 22-yr analysis agrees with BZ03’s 9-yr analysis, in that convection maxima are found (i) along the Champlain valley extending into the southern SLRV, (ii) over a broad area in the western SLRV including Ottawa, and (iii) north of Trois-Rivières along the base of the Laurentians (see Fig. 1 for geographic references). Similarly, the extended climatology reproduces the broad, half-hourglass-shaped convection minimum extending south–north from Lake Ontario and the Adirondacks, through the McGill radar, and into the Laurentians.

For \( f_5 \) in Fig. 2b, the admission of more data provides a clearer picture of the regional deep-convection distribution. This enhanced clarity, coupled with our interest in all deep-convection events (rather than just the severe ones), renders \( f_5 \) preferable to \( f_{15} \) for subsequent analysis. Interestingly, the extrema remain in similar locations as those in the \( f_{15} \) field in Fig. 2a, suggesting that deep-convection occurrence follows similar spatial patterns regardless of its intensity. The Montreal region is characterized by strong variability in convection occurrence, as sharply suppressed \( f_5 \) west of the city transitions to values commensurate with the SLRV-wide mean to the east. The minimum to the west (and, hence, upwind) of the city is consistent with Niyogi et al. (2011), who found a similar suppression to the west of Indianapolis. However, the limited enhancement to the east of Montreal contrasts with the strong enhancements found downwind of other cities (e.g., Niyogi et al. 2011; Haberlie et al. 2015).

**Random decompositions**

It is fair to question whether the local extrema in Fig. 2b are owing to a repeatable climatological signal or are just dominated by a handful of large, random events. To evaluate that question, we randomly decompose the full database of events (524 total) into two mutually exclusive sets and compare their \( f_i \) distributions. If the patterns in Fig. 2b are robust, they should reappear over these randomized sets, albeit with some variation due to the reduced sampling. Such an analysis is performed twice herein, the results of which are shown in Fig. 3.
Each row compares the $f_5$ distribution from two independent sets of 262 random events. The only difference between the two rows is the seed of the random-number generator used for the decomposition.

When comparing the left and right columns of Fig. 3, some discrepancies appear among the different event subsets, particularly in the representation of the $f_5$ maximum near Ottawa. This maximum is pronounced in Figs. 3b and 3c but weakens substantially in Figs. 3a and 3d, suggesting that it may be owing to a chance combination of a few large events rather than a repeatable climatological signal. In contrast, the two other above-mentioned maxima (in the Champlain valley and north of Trois-Rivières), and the north–south minimum extending through Montreal, clearly persist in all four random subsets. Additional random decompositions of the events provide similar results (not shown). Thus, while the details of the $f_5$ extrema differ in each decomposition because of interevent variability, the majority of these extrema appear to be climatologically robust.

**c. EVIL composites**

Synoptic-scale forcing and variability influences, among other things, the potential for deep convection on a given day, along with broader organization and propagation of the storms. While our focus on topographic forcing precludes a thorough analysis of these factors, we briefly consider the impacts of varying ambient
winds on the regional distribution of storm occurrence. These winds affect the strength and location of low-level convergence zones driven by mesoscale topography (e.g., Crook and Tucker 2005; Kirshbaum and Wang 2014), and thus the preferred sites of convection initiation. They also control the mode of thunderstorm organization (ordinary cells, multicells, or supercells) and the direction of storm motion (Markowski and Richardson 2010). As a result, storm occurrence exhibits strong sensitivity to ambient-wind direction over complex topography (e.g., Kuo and Orville 1973; Wasula et al. 2002). To examine these impacts over southern Quebec, we composite the events based on the mean geostrophic winds over the 1000–800-hPa layer and at the 700-hPa level [the latter following Wasula et al. (2002)]. The former controls the patterns of boundary layer convergence over complex topography while the latter more influences the direction of incipient storm motion.

For ease of analysis, we restrict consideration to the days where intense and persistent convection occurred in southern Quebec. We consider only the days where the domain-maximum EVIL continuously exceeded $a_{15}$ for longer than 90 consecutive minutes (19 scans). In total, 164 such days were identified over 1993–2014, which constitutes less than half of the 524 days where EVIL exceeded $a_5$ in at least one scan. Nevertheless, the filtering still retains the majority (73%) of the convection scans where EVIL exceeded $a_5$ because of the increased persistence of convection in these 164 events. Hence, the $f_{15}$ and $f_5$ patterns in this 164-event subset are nearly indistinguishable from those obtained 

FIG. 3. As in Fig. 2, but for two random decompositions of the 524 convection days. (a),(b) The first and (c),(d) the second decomposition. These two decompositions vary only in the seed of the random-number generator used to subdivide the events.
in the full dataset (cf. Figs. 2a,b and Figs. 2c,d), except for a general decrease in $f_5$ owing to reduced $n(x, y)$.

For each of the 164 convection days, we determine the mean 1800 UTC geostrophic wind from the National Centers for Environmental Prediction North American Regional Reanalysis (NARR; Mesinger et al. 2006). These data use 45 vertical layers and 32-km horizontal grid spacing, which is sufficient to estimate the larger-scale flow properties. The geostrophic winds are calculated as mean isobaric gradients in geopotential height at each NARR grid point over a 192 km $\times$ 192 km horizontal box centered on Montreal. Over the 1000–800-hPa layer, the NARR geostrophic winds were westerly (247.5°–292.5°) on 41%, southwesterly (202.5°–247.5°) on 32%, northwesterly (292.5°–337.5°) on 15%, and from all other directions on 12%, respectively, of the days. Because westerly or southwesterly geostrophic winds prevailed the vast majority (73%) of the time, we henceforth focus on these two directions.

Figure 4 compares $f_5$ composites for westerly and southwesterly geostrophic winds over 1000–800 hPa. One notable difference is a northward shift in the $f_5$ maximum within the Champlain valley as the wind rotates from westerly to southwesterly. Whereas this maximum is confined to the central parts of the valley and nearby Green Mountains in the westerly events, it spreads northward into the SLRV in southwesterly events. Similarly, the $f_5$ maximum near Ottawa spreads northward from the SLRV into the southwest Laurentian foothills. The third prominent $f_5$ maximum north of Trois-Rivières shifts northward and weakens in the southwesterly events. For the most part, the broad convection minimum extending south–north from Lake Ontario and the Adirondacks, through Montreal, and into the central Laurentians, remains intact in the two composites. However, this feature is interrupted in the southwest composite by a thin west–east strip of enhanced $f_5$ extending from the Laurentian foothills eastward across the SLRV (Fig. 4b).

Because of the climatological dominance of westerly upper-level winds, the fraction of westerly events increases at the 700-hPa level (55%), in part at the expense of the southwesterly events (25%). Nevertheless, because these two directions still make up the vast majority of all events, they remain the focus of our 700-hPa analysis. The associated $f_5$ composites (Fig. 5) are similar to the 1000–800-hPa composites (Fig. 4) in that the maxima generally shift northward when the winds possess a more southerly component, while the two convection minima remain largely intact. These similarities reflect that events dominated by southwesterly flow at 700 hPa also tend to exhibit southwesterly flow over 1000–800 hPa (see, e.g., Fig. 6).

The above analysis is consistent with studies over different regions (e.g., Kuo and Orville 1973; Wasula et al. 2002) in that the ambient winds control the structure, but not necessarily the existence, of regional convection maxima and minima. The northward shift of these features under southwesterly winds likely relates to displacements in the locations of subcloud vertical motion driving convective initiation and/or changes in storm motion. We examine the impacts of ambient-wind...
direction on regional convection more closely through numerical simulation in section 3.

3. Numerical simulations

To gain insight into the contribution of the southern Quebec topography to the regional convection climatology (Fig. 2b), we conduct quasi-idealized numerical simulations of convection events using the Advanced Research version of the Weather Research and Forecasting Model, version 3.5. Provided that the simulations reasonably reproduce some of the observed convection extrema from section 2, their four-dimensional gridded datasets can help to pinpoint the physical origins of these features.

a. Setup

The most obvious method for reproducing the regional convection climatology over southern Quebec would involve brute force: real-case, explicit-convection simulations of all 524 observed events. However, such an endeavor is prohibitively expensive given our current computational resources and still potentially undermined by model uncertainties. While subsets of the events would be more affordable, their reduced sampling may cause climatological signals of interest to be obscured by the interevent variability. Following Kirshbaum et al. (2016), a far simpler and more affordable (though still imperfect) approach followed herein is to perform a small number of quasi-idealized simulations to broadly capture the topographic impacts on summertime convection over southern Quebec, under a range of environmental conditions.

The simulations use a horizontally homogeneous single-sounding initialization in conjunction with the real topography of southern Quebec. They integrate the diurnal flow evolution from 1200 UTC [0700 local standard time (LST)] to 0000 UTC (1900 LST) the following day, thus encompassing the 1700–0000 UTC afternoon analysis period. Importantly, this simplified approach neglects synoptic-scale forcing as well as the meridional baroclinicity that often prevails in the summer, with warmer air to the south and cooler air to the north. For these reasons, the experiments cannot capture the full range of processes controlling southern Quebec convection, and should not be expected to reproduce the full observational climatology of Fig. 2. However, this simplified numerical framework helps to isolate topographic effects in the absence of other complexities that might obscure them, while also limiting computational expense. A detailed investigation of the impacts of synoptic forcing and variability on southern Quebec convection is deferred to future work.

An ensemble of 15 simulations is performed, all of which are identical except for their initial sounding profiles. The soundings sample a realistic range of ambient wind directions, wind speeds, and thermodynamic conditions. To obtain “control” soundings for different principal directions (only westerly, southwesterly, and northwesterly are considered because they make up over 90% of all events), we composite the NARR
soundings based on the same 1000–800-hPa geostrophic wind classification scheme from section 2c. However, we use the 1200 UTC (rather than the 1800 UTC) NARR data on the morning of convection events to represent the conditions at model startup. Once the wind directions are classified for each sounding, we group the soundings according to their wind direction and calculate statistics of relevant variables (winds, temperatures, etc.) at each vertical level. The soundings use the geostrophic, rather than the actual, winds so that the simulations can develop ageostrophic motions naturally in response to topographic forcing. Only the 1993–2007 time period is considered in the sounding design, as the simulations were conducted prior to the completion of the 2008–14 radar analysis.

Because the temperature $T$ and dewpoint $T_d$ profiles for the three wind directions are not significantly different (not shown), we use the westerly thermodynamic sounding for all cases (Fig. 6). This profile exhibits a surface-based radiative inversion and conditional instability between 900 and 500 hPa, with a most-unstable convective available potential energy (CAPE) of 151 J kg$^{-1}$, and a corresponding convective inhibition (CIN) of 82 J kg$^{-1}$, for an air parcel originating at 700 m. As will be seen, surface-based instability develops during the day in response to diurnal heating. The mean westerly, southwesterly, and northwesterly wind profiles are shown in various colors in Fig. 6. Despite their different low-level wind directions, the mid- to upper-level winds are westerly in all three cases. Relative to individual events, the wind magnitudes (and, hence, the vertical shear) in these composite profiles is reduced via the averaging procedure.

To broadly represent the interevent variability in environmental conditions, we consider five different soundings for each wind direction:

(i) Control: mean $T$ and $T_d$ profiles, mean geostrophic wind profile;
(ii) MOIST: same as control but relative humidity (RH) increased by 1 standard deviation;
(iii) DRY: same as control but RH decreased by 1 standard deviation;
(iv) STRONG: same as control but wind speed increased by 1 standard deviation;
(v) WEAK: same as control but wind speed decreased by 1 standard deviation.

The RH and wind statistics are analyzed separately at each vertical level, so that the above standard deviations are functions of height. To illustrate the range of soundings thus produced, the moist and dry $T_d$ profiles are overlaid on Fig. 6. Although these initial soundings are based on real events and sample a reasonable range of environmental conditions, our limited suite of reduced numerical simulations still cannot capture the true diversity and time evolution of such events.

The model configuration includes two one-way nested grids both centered on Montreal (Fig. 7). Grid 1, with a horizontal grid spacing $\Delta$ of 3 km and open (radiative) lateral boundaries, covers most of the northeast United States and southeast Canada. Grid 2 is nested inside grid 1 with $\Delta = 1$ km, covering the McGill radar scan area and adjacent regions over New York and New England. Both grids use a stretched, terrain-following vertical grid with 81 levels, with a nominal spacing ranging from 108 m at the surface to 680 m at the model top (17 km). A 5-km-deep Rayleigh sponge layer is placed at the model top to limit spurious reflection of gravity waves. The model terrain gradually decays to zero toward the lateral boundaries of grid 1 using a Gaussian function with a 200-km half-width (Fig. 7). This terrain decay prevents spurious large-amplitude gravity waves from developing at the open lateral boundaries and propagating into the domain interior. Sensitivity experiments reveal that the model results within grid 2 are largely insensitive to the half-width of the terrain decay on grid 1 (not shown). The regional land use, provided by the USGS (Fig. 1a), is maintained over the entire domain.

The simulations use the Noah land surface model to represent the surface energy balance, with a single-layer urban canopy model (UCM). The land use, vegetation fraction, soil type, soil moisture, and vegetation type are initialized as USGS summer climatological mean values (some of these fields are shown in Fig. 1). The relatively simple UCM is consistent with the other major idealizations in our model setup. Other parameterizations include the Goddard shortwave and Rapid Radiative Transfer Model longwave radiation schemes, with the shortwave forcing based on a midsummer day (12 July). The surface layer is parameterized by Monin–Obukhov similarity theory, and the boundary layer is parameterized by the Mellor–Yamada–Janjic scheme. Cloud microphysics are parameterized by the two-moment Morrison scheme, with a default continental cloud-droplet concentration of 250 cm$^{-3}$. Given that the initial flow is in geostrophic balance, we apply the Coriolis force only to perturbations from that state using an $f$-plane approximation with a Coriolis parameter of $f = 10^{-5}$ s$^{-1}$. The land surface is initialized
with a skin temperature of 292 K at sea level and a vertical lapse rate of 4 K km\(^{-1}\) over sloping terrain.

For ease of reference, the simulations are henceforth identified based on their wind direction [west (WEST), southwest (SWEST), and northwest (NWEST)] and their sounding type (MOIST, DRY, STRONG, or WEAK). For example, the westerly control simulation is called WEST and the westerly simulation with reduced humidity is called WEST.DRY.

b. Results

To give a first impression of the flow and precipitation patterns in the simulations, Fig. 8 presents snapshots of simulated surface winds and radar reflectivity on grid 2 of the WEST, SWEST, and NWEST simulations at 2000 UTC (or 1500 LST, just before the 1600 LST climatological peak of convective occurrence found in BZ03). While all three cases develop isolated convection over the Laurentians, they exhibit differing low-level flows and storm development farther south. The WEST case exhibits valley-parallel low-level flow with convective clusters along the western flank of the SLRV and over the Green Mountains. The SWEST case also exhibits SLRV-parallel surface flow, but convective cells are more concentrated in the upper Champlain valley, Green Mountains, and lower St. Maurice River valley. In the NWEST case, pockets of convection develop to the northeast of Lake Ontario, to the southeast of Ottawa, and within the eastern SLRV.

To compare the simulations with the observations of section 2, we present a simulated \(f_5\) “climatology” for (i) the ensemble of 15 simulations, (ii) the five WEST simulations, and (iii) the five SWEST simulations. To this end, we obtain simulated EVIL via (1) by integrating the precipitation density (the product of air density and the sum of the precipitation mass mixing ratios) vertically from 5 km to the cloud top. Each simulation is assigned a weight \(w_{\text{sim}}\) corresponding to the occurrence frequency of its 1000–800-hPa wind direction over the 164 convection days (WEST: 0.41, SWEST: 0.32, NWEST: 0.15). As in the observational analysis, we loop through all model output times between 1700 and 0000 UTC of each simulation and increment the numerator \(n(x, y)\) of \(f_5\) each time EVIL exceeds \(a_2\). However, rather than incrementing \(n(x, y)\) by one each time the EVIL surpasses \(a_2\), we increment it by \(w_{\text{sim}}\). The ratio \(n/d\) thus gives the frequency of simulated convection occurrence weighted by \(w_{\text{sim}}\). Finally, to account for the fact that the simulations include convection days only while the observations include all days, we scale \(n/d\) by the fraction of convective days in our 22-yr climatology (0.089) to get the simulated \(f_5\).

Comparing the simulated \(f_5\) climatology (Fig. 9) with the observational \(f_5\) climatology (Figs. 2b and 4), the simulated \(f_5\) maxima are slightly more intense, with peak magnitudes about 30% larger. This intensification may arise from diminished sampling and/or the absence of synoptic variability in the simulations, the latter leaving topographic heterogeneities as the primary driver of storm initiation. Hence, areas of forced low-level convergence and associated vertical motion, and thus the locations of convective initiation, remain largely stationary throughout the simulation. Under synoptic forcing, by contrast, storms tend to develop within
pockets of enhanced larger-scale ascent and moist instability that migrate across the study region (associated with fronts, shortwave jet disturbances, etc.), which distributes their occurrence more evenly in space.

Despite the above-mentioned magnitude errors, the simulated climatology impressively reproduces the structure and sensitivities of the $f_5$ maximum within the Champlain valley, which expands from the southern part of the valley (WEST) northward into the southern SLRV (SWEST). It also reasonably reproduces the half-hourglass-shaped $f_5$ minimum extending meridionally from upstate New York, through the McGill radar, and into the Laurentians (except for some interruption within the SLRV, as in the observations).

While the two other relative maxima in the Ottawa vicinity and to the north of Trois-Rivières are also produced by the 15-member ensemble, they exhibit different sensitivities to wind direction than those observed. Whereas the former persists within the SLRV under both westerly and southwesterly winds in the observations (Figs. 4b,c), it shifts out of the SLRV and into the Laurentian foothills in the SWEST simulation (Figs. 9b,c). This disagreement may stem in part from the lack of climatological robustness of this feature (section 2b)—if the observed signal is dominated by a few extreme events, our composite approach cannot be expected to accurately reproduce it. Similarly, the maximum north of Trois-Rivières is reproduced in the 15-member ensemble but exhibits different sensitivity to wind direction than that observed. Rather than shifting northward and weakening as the winds rotate from westerly to southwesterly, the simulated maximum shifts westward and maintains its strength. This discrepancy may result, at least in part, from the use of Montreal as the centerpoint of the NARR-based initial soundings, which may not reliably represent the flow ∼200 km to the northeast.

4. Discussion

a. Physical interpretation of $f_5$ extrema

While the simulated $f_5$ climatology in Fig. 9 is imperfect, it does reproduce several of the key observed features reasonably well. Hence, the simulations provide a useful framework for interpreting some key mechanisms behind these convection extrema. To that end, we analyze several simulated quantities that bear on the initiation and intensity of deep convection. For brevity we focus on the WEST and SWEST simulations, which are broadly representative of the other four simulations within the same wind-direction bins. The quantities of interest are the mean-layer [0–500-m above ground level (AGL)] CIN (Figs. 10a,c), CAPE (Figs. 10b,d), water vapor mixing ratio $q_v$ (Fig. 11), moisture divergence (Figs. 12a,b), and wind divergence (Figs. 12c,d). These quantities are averaged over the preconvective period (1500–1700 UTC) to minimize complicated latent heat feedbacks.

The simulated CIN is generally low (<10 J kg$^{-1}$) over most of the domain except for the prominent water
bodies of Lake Ontario, Lake Champlain, the St. Lawrence River, and the Ottawa River (Figs. 10a,c). A broken strip of enhanced CIN extends along the axis of the St. Lawrence River, with maxima running streamwise through both Montreal and Trois-Rivières. These two CIN maxima are associated with lakes interspersed along the St. Lawrence River (Fig. 1c), where suppressed heat fluxes limit the destabilization of the low-level flow. The simulated CAPE is lowest over these same water bodies, and highest (>400 J kg⁻¹) in locations with large \( q_v \) and wind/moisture convergence (Figs. 10b,d, 11, and 12). The apparent correlation between \( q_v \) and CAPE arises from the potential for increased latent heat release, and hence buoyancy, within moister air parcels. In contrast, the correlation between \( q_v \) and convergence likely arises from reduced vertical mixing of dry air aloft in areas of organized ascent (e.g., Markowski and Richardson 2010).

In the WEST case, CAPE and \( q_v \) maxima tend to occur where moist valley flow collides with airstreams of different origin, for example along the Champlain valley where southerly flow converges with dry, westerly flow crossing the Adirondacks (Figs. 10b and 11a). Because of suppressed heat fluxes over Lake Champlain, however, this local maximum does not extend farther north. Similarly, the CAPE/\( q_v \) maximum along the western edge of the SLRV near Trois-Rivières coincides with the collision of southwesterly SLRV-parallel flow with...
drier, westerly flow over the Laurentians. These leeside convergence zones are promoted by mountain-wave-induced lee troughs along the downwind (eastern) side of prominent mountain ridges (e.g., Bluestein 2008). Finally, the diffuse maximum in CAPE/$q_v$ in the SLRV between Ottawa and Lake Ontario arises from a combination of (i) broad convergence owing to a narrowing of the SLRV, (ii) collision of channeled flows within the SLRV and the Ottawa River valley, and (iii) a lake breeze originating along the northern coast of Lake Ontario (see section 4b for further details).

The SWEST case features a more prominent and elongated maximum within the Champlain valley (Figs. 10d and 11b), where an increased geostrophic-wind component along the valley axis supports a stronger air current that maintains the low-level humidity despite the suppressed latent heat fluxes over Lake Champlain. On its northward journey, this air first converges with dry, southwesterly flow descending the Adirondacks, then collides with moist, southwesterly flow within the SLRV. As a result, its associated CAPE/$q_v$ maximum extends through the length of the Champlain valley and into the SLRV, to points just southeast of Montreal. Another notable CAPE/$q_v$ maximum is found within the southern St. Maurice valley (47°N, 73°W), where southerly flow channeled into the valley converges with more southwesterly flow crossing the Laurentians. While a CAPE/$q_v$ maximum is again found between Ottawa and Lake Ontario, the mass and moisture convergence in this region is substantially reduced. This reduction is owing to a weaker along-valley flow component through the Ottawa valley as well as a suppressed lake breeze due to enhanced onshore flow over the northern coastline of Lake Ontario.

While the $f_5$ maxima in Fig. 9 generally coincide with the areas of low CIN, high CAPE, and strong mass and/or moisture convergence in Figs. 10–12, the $f_5$ minima form where one or more of the above elements is lacking. Despite the existence of broad regions with little to no CIN in Figs. 10a and 10b, sustained deep convection only frequents the areas that also exhibit organized low-level convergence. The lack of storm development in the absence of such convergence may stem from the inability of isolated thermals to withstand the effects of dry-air entrainment (e.g., Khairoutdinov and Randall 2006; Kirshbaum 2011). While CIN is nearly zero over the central Laurentians and Adirondacks, both areas exhibit limited $q_v$ (<11 g kg$^{-1}$) and CAPE (<400 J kg$^{-1}$), along with disorganized mass/moisture convergence, which together suppresses the initiation and intensity of the convection. Other areas of suppressed storm activity include southwestern parts of the domain, where reduced heat fluxes over Lake Ontario maximize the CIN.

The majority of simulated convective storms decay within an hour after their initiation, as they drift eastward under westerly or southwesterly flow aloft. Thus, most of the simulated $f_5$ distribution can be explained...
by the above arguments pertaining to convection initiation. However, some multicell complexes develop that persist for multiple hours in the WEST suite of simulations. These systems form in environments sharing three characteristics: (i) organized, quasi-stationary horizontal convergence, (ii) substantial CAPE \( (>500 \text{ J kg}^{-1}) \), and (iii) locally enhanced low-level vertical shear. Figure 13 presents radar reflectivity and 0–2-km-AGL bulk shear vectors at three times during the WEST simulation. A multicell complex develops just east of Ottawa, where persistent convergence repeatedly initiates deep convection. This convection is organized by locally enhanced northwesterly low-level wind shear because of valley channeling (and backing) of near-surface flow within the SLRV (this channeling is apparent in Figs. 8a and 10–12). Once formed, this complex propagates downshear (southeastward), as predicted by the theoretical arguments of Rotunno et al. (1988). Such complexes only develop in the WEST simulations because of their stronger valley flow backing and hence stronger low-level vertical wind shear.

b. Sensitivity tests

To determine the relative contributions of different topographic features to the simulated deep-convection
patterns over southern Quebec, we perform the following sensitivity tests:

- FLAT: the orography is flattened;
- UNILU: the land use is homogenized as the dominant land-use type within the St. Lawrence valley (deciduous broadleaf forest);
- NOWTR: all water bodies are replaced by deciduous broadleaf forest;
- NOURB: urban regions are replaced by deciduous broadleaf forest.

For brevity only four sensitivity tests are discussed, all using the initial sounding from the WEST case. However, the ensuing results are consistent with additional experiments using the SWEST case as the reference (not shown). As before, the names of the sensitivity experiments are appended to that of the reference simulation (e.g., the WEST case with flat terrain is termed WEST_FLAT). We compare the cumulative precipitation $P$ over 1700–0000 UTC of these four sensitivity tests with the WEST simulation in Fig. 14. For this comparison, $P$ is preferable to $f_5$ because of its smoother appearance, although the two yield similar spatial patterns (cf. Figs. 14a and 9b).

While the $P$ distributions in the WEST, WEST_UNILU, and WEST_NOWTR cases are broadly similar, some land-use-related sensitivities appear when comparing the WEST and WEST_UNILU simulations (Figs. 14a,b). First, the $P$ maximum in the Champlain valley and Green Mountains reaches farther north in the WEST_UNILU case because of the removal of Lake Champlain and an attendant local increase in surface heat fluxes. Second, the $P$ maximum within the SLRV between Ottawa and Montreal weakens substantially because of the removal of Lake Ontario and the Ottawa urban region. While the former eliminates the coastal convergence along the lake’s eastern edge and the lake breeze streaming past its northern shoreline, the latter eliminates the UHI over Ottawa. The consequently weakened moist instability and flow convergence, in turn, limit the prevalence of deep convection. By contrast, the removal of Montreal in the WEST_UNILU case has no effect on $P$ (this finding is explained shortly).

The influence of water bodies alone is demonstrated by the WEST_NOWTR simulation (Figs. 14d), the $P$ field of which differs from the WEST and WEST_UNILU cases mainly in the two convection maxima downwind of Ottawa and northeast of Trois-Rivières. Compared to the WEST_UNILU case, the WEST_NOWTR case shows a more pronounced $P$ maximum to the east of Ottawa because of convection associated with the UHI there. Moreover, the $P$ maximum to the northeast of Trois-Rivières in the WEST_NOWTR case is weaker than that in the WEST_UNILU case because of competing effects of orography and land-use variations. As hypothesized in section 4a, this maximum is dominated by orographically forced convergence of westerly flow crossing the Laurentians with southwesterly flow within the SLRV. While land-use variations make no contribution to this convergence in the WEST_UNILU case, the contrasting land use between the Laurentians and the SLRV (as seen in Fig. 1a) sharpens the convergence in the WEST_NOWTR case. Indeed, further analysis of these cases reveals that the land-use-induced convergence patterns interfere destructively with the orographically forced convergence in this region (not shown), thus weakening the local $P$ maximum. While such destructive interference also occurs in the WEST simulation, it is offset by a convergent river breeze past the locally widened St. Lawrence River (see Fig. 1b).
Relative to the WEST case, precipitation is generally reduced in the WEST.FLAT case except along the flanks of Lake Ontario (and downwind) where lake breezes initiate convection, and downwind of Ottawa and Montreal, where the associated UHIs initiate convection. Overall, the $P$ distribution from this simulation shares little in common with that of the other cases in Fig. 14. Given that the key difference in this case is the removal of the orography, its strong disagreement with the other cases suggests that orography dominates the regional convection pattern.

The impacts of regional urbanization are revealed by comparing the $P$ fields in the WEST.NOURB and WEST simulations in Figs. 14a,e. The most obvious contrast between these cases is a reduction in $P$ just east of Ottawa that is attributable to the removal of its UHI. In the Montreal region, by contrast, $P$ is minimal in both cases. The generally suppressed precipitation around Montreal arises from two effects: (i) increased CIN resulting from reduced surface heat fluxes over the widened St. Lawrence River just upwind (Figs. 10a,c), and (ii) a local widening of the SLRV as it opens into the Champlain valley to the south. The latter effect is apparent in Fig. 15a, which subtracts the 10-m winds and divergence of the WEST.FLAT simulation from that of the WEST simulations, thus isolating the orographic imprint on the wind field. Along the SLRV, a west–east dipole of valley-wide convergence and divergence is found, with the transition occurring near the McGill radar where the valley begins to widen to the east. A similar comparison of the WEST.NOURB and WEST simulations (Fig. 15b) reveals that both cities exhibit UHIs exceeding 1 K, each giving rise to organized convergence downwind. However, only the Ottawa UHI leaves a signature on the $P$ (and $f_3$) patterns downwind (Figs. 14a,e). The differing impacts of these two UHIs may be attributed to the broader flows in which they reside: whereas Ottawa is located within a generally convergent flow with locally weak CIN and large CAPE, Montreal is embedded within a generally divergent flow with relatively large CIN and small CAPE (Fig. 12).
c. Limitations of the numerical results

While the topographic forcing mechanisms inferred from the simulations are physically plausible, they cannot provide a complete understanding of the observed $f_i$ pattern in Fig. 2. As discussed in section 3a, the numerical setup contained major idealizations to limit computational expense and isolate topographic influences. The key simplification was the neglect of synoptic-scale forcing: whereas convection initiation in the simulations is exclusively tied to mesoscale topographic features, it in reality requires a suitable combination of synoptic and topographic forcing. The former may influence both storm initiation and longevity, in that storms tend to persist longer and/or regenerate within areas of sustained large-scale ascent. Moreover, it may give rise to organized convective systems that exhibit different sensitivities to regional topography (e.g., Frame and Markowski 2006).

We suspect that the inclusion of synoptic forcing and variability, realized through real-case simulation of many observed convection events, would at least add a mean component to the purely topographic signature in Fig. 9. It may also alter the structures of the convection maxima and minima, if the flow response to the topography significantly changes in a dynamically active environment. Moreover, meridional baroclinicity, which is suggested by the veering WEST and SWEST soundings in Fig. 6 but neglected in the simulations, may cause a general north–south gradient in convection occurrence and/or intensity [as found in the lightning climatology of Kovacs (2015)]. Thus, to obtain a comprehensive picture of the mechanisms influencing the regional convective pattern, the contributions of synoptic forcing and variability must be considered in conjunction with the topographic forcing.

Another limitation is the single-layer bulk UCM, which introduces error by simplifying the highly complex water and energy budgets, and the effects of three-dimensional buildings, within urban areas (e.g., Chen et al. 2011; Wang et al. 2013; Li et al. 2013). Such deficiencies may limit the realism of the response of the simulated convection to the larger urban regions of Ottawa and Montreal. Although the minimal enhancement in simulated deep convection downwind of Montreal is not necessarily a spurious result (see discussion of section 4b), a more sophisticated UCM may be needed to properly represent the regional UHIs.

5. Conclusions

This study examined the spatial distribution of convective occurrence on summer afternoons over southern Québec, from both observational and numerical perspectives. Using data from the McGill radar in Ste. Anne de Bellevue, Quebec (30 km west of downtown Montreal), the 22-yr observational analysis (1993–2014) extended and complemented a previous 9-yr analysis by BZ03. Investigation of 524 convection events over this period indicated three robust regional mesoscale convection hotspots: (i) a meridional axis extending from the Champlain valley of New York and Vermont (United States) into the St. Lawrence River valley of Quebec, (ii) a southwest–northeast axis along the base of the Laurentian Mountains north of Trois-Rivières, and (iii) a broad area to the west of Montreal (including Ottawa) in the western SLRV (see Fig. 1 for regional geographic maps). By contrast, a broad convection minimum was observed within a meridionally oriented, half-hourglass-shaped region extending from Lake Ontario and the
Adirondack Mountains of New York north through the Montreal region and into the central Laurentians. Composites of events based on the low-level (1000–800 hPa) and midlevel (700 hPa) geostrophic wind directions indicated that the two convection minima were less sensitive to ambient wind direction than the three maxima. As might be expected from basic physical considerations, the latter features shifted northward as the geostrophic winds rotated from westerly to southwesterly.

An ensemble of 15 explicit-convection numerical simulations was conducted to interpret the observed patterns of convection occurrence. The simulations used a horizontally homogeneous initialization based on reanalysis soundings, along with the real topography of the region (with some modification near the open lateral boundaries) and a full-physics configuration with a single-layer urban canopy scheme. Despite their simplifications, the simulations reasonably reproduced all four of the above-mentioned convection extrema. Model analysis revealed that all of the convection maxima formed in locations possessing three environmental properties: near-zero convective inhibition, substantial convective available potential energy (CAPE > 500 J kg⁻¹), and organized horizontal wind and/or moisture convergence, while the minima lacked one or more of these properties. Using this model guidance, the various convection extrema are explained as follows:

- The Champlain valley maximum is caused by the collision of westerly or southwesterly flow traversing the Adirondacks with moist, southerly flow channeled through the valley.
- The maximum north of Trois-Rivières is caused by the collision of westerly or southwesterly flow traversing the Laurentians with moist, southwesterly flow channeled through the SLRV and/or the St. Maurice valley.
- The broad maximum to the west of Montreal (including Ottawa) is caused by a combination of (i) flow convergence associated with the gradual narrowing of the SLRV, (ii) convergence at the junction of the SLRV and Ottawa River valleys, (iii) lake breezes past Lake Ontario, and (iv) an urban heat island (UHI) past Ottawa. Enhanced low-level vertical shear associated with valley flow channeling helps to organize these storms into long-lived multicell complexes.
- The broad minimum from Lake Ontario and the Adirondacks, through Montreal, and into the Laurentians arises primarily from suppressed heat fluxes over water surfaces (Lake Ontario and the widened St. Lawrence River upwind of Montreal) and drier and less unstable airflow over the mountains.
- In contrast to Ottawa, no obvious local enhancement in deep convection is found downhill of Montreal, despite the existence of a 1–2-K UHI there. The inability of convection to initiate past Montreal is owing to locally increased CIN and broad terrain-forced divergence (a local widening of the SLRV), which offsets the UHI-forced horizontal convergence there.

The above findings focus exclusively on topographic impacts on convective-storm occurrence. They were facilitated by the simplicity of our quasi-idealized numerical framework, which permitted the isolation of topographic impacts in the absence of synoptic forcing and variability. As a result, however, the simulations do not capture the full range of processes that influence regional storm occurrence. Thus, to complement the current findings, we intend to broaden our investigation in the future to include synoptic forcing and variability. We will perform an observational synoptic analysis of regional storm events, complemented by real-case, convection-permitting simulations of many events (including, at the least, the 164 events used for the composite analysis). These simulations will also use a more sophisticated urban canopy model to better represent the diverse interactions (on the synoptic, meso-, and storm scales) that control regional convective-storm development.

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