

# Observational and modelling analysis of Canada's only F5/EF5 tornado

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## Key Points:

- Weather observations and numerical simulations were used to diagnose the pre-storm environment and triggering of the Elie, Manitoba tornado.
- Pre-storm conditions were found favorable for tornadic supercells; enhanced by the interaction between a remnant cold pool and ambient flow.
- The interaction between a trough and boundary-layer thermals was the primary triggering mechanism of the Elie, Manitoba tornadic supercell.

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**Abstract**

Canada's first and only F5/EF5 tornado associated with a supercell touched down near Elie, Manitoba in the late afternoon of 22 June 2007. An observational and numerical simulation analysis with the Weather Research and Forecasting (WRF) model was undertaken to characterize the pre-storm environment and processes leading to storm initiation. WRF sufficiently reproduced the synoptic and mesoscale features, including a supercell-like storm in the region of interest, and supplemented available observations. Synthesis of observational and simulation data suggests that the environment near Elie immediately before storm initiation was primed for tornadic supercells, with large most-unstable and mixed-layer convective available potential energy ( $> 4000 \text{ J kg}^{-1}$ ) and sufficient vertical shear (effective bulk wind shear  $\sim 40 \text{ kt}$ ; effective storm-relative helicity  $> 200 \text{ m}^2 \text{ s}^{-2}$ ). Despite enhancement owing to a cold pool left behind by passing early-afternoon convection, shear remained weaker than those typically found in other North American significant tornadic supercell events. The interaction between a surface trough and convective boundary-layer thermals was the primary triggering mechanism of the Elie supercell. The former appeared to be associated with a low pressure arising from the juxtaposition of lower-troposphere cyclonic differential vorticity advection and lee troughing over the western Red River Valley. More observational analysis and numerical sensitivity experiments are required to better diagnose Manitoba terrain's contribution to the Elie supercell initiation.

**Plain Language Summary**

A severe thunderstorm produced the strongest tornado ever recorded in Canada on 22 June 2007 that struck Elie, Manitoba, Canada. To better understand and characterize the conditions leading to the storm, weather observations and the data produced by a specialized computer model were analyzed. We found that the conditions were overall favorable for the formation of severe thunderstorms that would produce tornadoes. These conditions were enhanced by a cluster of showers that passed over the area in the early afternoon. An external lifting mechanism was also required to initiate the storm, which we attributed to the combined lift associated with a surface air mass boundary and updrafts that formed due to daytime solar heating. The air mass boundary may have been associated with a low pressure system resulting from airflow interacting with the shallow western Manitoba terrain. Additional observational analysis and computer modelling experiments are needed to gain further insights into the effects of Manitoba's terrain on the formation of the Elie, Manitoba tornadic thunderstorm.

**1 Introduction**

Although not as well-known as the U.S. for significant tornado events, Canada has experienced several notable modern-day tornado disasters including the 1985 Barrie tornado (Etkin et al., 2002), the 1987 Edmonton tornado (Bullas & Wallace, 1988; Charlton et al., 1995), and the 2000 Pine Lake tornado (Joe & Dudley, 2000; Erfani et al., 2003). These events caused property damage ranging from a few to hundreds of millions of CAD, with fatalities as many as 30 and injuries as many as 300. Therefore, it is in the public's interest to better understand Canadian tornado environments to improve knowledge gaps and prediction of such events. Such is one of the goals of the Northern Tornadoes Project (NTP; Sills et al., 2020), which was established in 2017 to focus on improving the detection and documentation of Canadian tornadoes using various existing and new data sources and technologies.

As part of NTP science, we present a meteorological analysis of the Elie, Manitoba F5 tornado that occurred on 22 June 2007. This event deserves a detailed case study since it is still the strongest tornado in Canada, and has several unresolved questions about its formation and evolution. The tornado occurred between 2320 UTC 22 June 2007 to

65 0000 UTC 23 June 2007 (CDT=UTC-5). It was narrow ( $\sim 200$  m wide) and slow-moving  
66 (i.e., traveling at  $\sim 2$  m s $^{-1}$ ; Hobson (2011)), wiping several houses off their foundation  
67 and tossing a cargo van along its path. It was rated F5 in the final damage assessment  
68 report, which was rather unusual given its narrow width (Brooks, 2004), and was still  
69 rated five on the Enhanced Fujita (EF) scale (MacDonald et al., 2004) after Canada adopted  
70 the scale in 2013. The full damage survey, rating process, and estimated path of this tor-  
71 nado can be found in McCarthy et al. (2008).

72 Because of its uniqueness among the documented Canadian tornadoes, an exam-  
73 ination of what the environment was like prior to storm initiation and what the storm-  
74 triggering mechanisms were, is warranted. Key tornadic supercell environmental ingre-  
75 dients include large conditional instability and low-level moisture, which are indicated  
76 by convective available potential energy (CAPE; e.g., Maddox, 1976; Brooks et al., 1994;  
77 R. L. Thompson et al., 2003, 2004a) and strong vertical wind shear, measured by bulk  
78 wind shear (or bulk wind difference; BWD) and/or storm-relative helicity (SRH; e.g.,  
79 Johns & Doswell III, 1992; Johns et al., 1993; E. N. Rasmussen & Blanchard, 1998; Markowski  
80 et al., 2003; R. L. Thompson et al., 2003). Low-level triggers of any deep moist convec-  
81 tion can be in the form of mesoscale boundaries (and their interactions) as well as synoptic-  
82 scale systems such as warm and cold fronts (e.g., Kingsmill, 1995; Koch & Ray, 1997;  
83 Ziegler & Rasmussen, 1998; Weckwerth & Parsons, 2005; Wakimoto & Murphey, 2010;  
84 Wang & Kirshbaum, 2015; Wilson et al., 2018). Some observational studies have shown  
85 that boundaries can also modify the local wind pattern and lead to more favorable con-  
86 ditions for tornadoes (e.g., Maddox et al., 1980; Sills & King, 2000; Giaiotti & Stel, 2007;  
87 Taszarek et al., 2016; Pilguy et al., 2019).

88 In this work we offer a subsequent opportunity to compare the Elie event to other  
89 significant tornado cases in Canada and the U.S. due to its significance in the nation’s  
90 tornado history. Studies of tornado environments based on historical events in the U.S.  
91 have been extensive (e.g., E. N. Rasmussen & Blanchard, 1998; Brooks et al., 2003; R. L. Thomp-  
92 son et al., 2003, 2007, 2012), with only limited work done in Canada. In Canada, Dupilka  
93 and Reuter (2006b) and Dupilka and Reuter (2006a) studied Alberta’s severe thunder-  
94 storm environments during a handful of cases, including those that were tornadic. Hanesiak  
95 et al. (2023) recently compared the storm environments during significant tornado (F2/EF2+)  
96 events in different provinces across Canada.

97 As a witness to the tornado’s entire life cycle, J. Hobson was the first to conduct  
98 an observational analysis and comparison of the Elie event to a few other significant tor-  
99 nado events in the U.S. and Canada (see Hobson (2011)). However, due to the lack of  
100 meteorological observations immediately before the supercell initiation, the full mesoscale  
101 environment and the physical mechanisms of the storm trigger(s) remain inadequately  
102 understood.

103 Building on Hobson (2011), the present study will utilize a numerical weather pre-  
104 diction (NWP) model to obtain the three-dimensional flow evolution during the Elie event.  
105 Numerical simulations have been used in many studies of notable tornado events around  
106 the world (e.g., Litta et al., 2010, 2012; Matsangouras et al., 2011, 2016; Taszarek et al.,  
107 2016; Miglietta et al., 2017; Pilguy et al., 2022). These studies successfully reproduced  
108 the observed synoptic and mesoscale flow and severe convection development to a rea-  
109 sonable degree using  $\mathcal{O}(1)$  km model grid spacings. Observational-modelling studies of  
110 Canadian events have been limited. A few exceptions include Erfani et al. (2003), who  
111 found that ascent and moisture transport associated with a mountain-plain circulation,  
112 coupled with deep-layer shear and destabilization ahead of an upper-level trough, led to  
113 the Pine Lake, Alberta tornadic supercell. Bisson and Paola (2000) found that the late-  
114 day low-level jet development over southern Manitoba created sufficient low-level shear  
115 to produce the 2000 Brunkild, Manitoba tornadic supercell in an otherwise suboptimal  
116 environment featuring only large conditional instability.

117 The objective of the present study is to identify and characterize the synoptic and  
 118 mesoscale features that contributed to producing the Elie tornadic supercell using all avail-  
 119 able observations and model simulations. This study will lay the groundwork for future  
 120 case studies of other Canadian significant tornado events, with the goal of improving the  
 121 understanding of their environments and physical mechanisms. The paper layout is as  
 122 follows: section 2 describes the observational datasets and numerical simulation setups.  
 123 Observational analyses of the pre-storm environment and storm evolution are shown in  
 124 section 3. Section 4 evaluates the model performance against available and proxy obser-  
 125 vations. Section 5 presents the analyses of the simulated flow immediately before the su-  
 126 percell initiation. The findings are summarized in section 6.

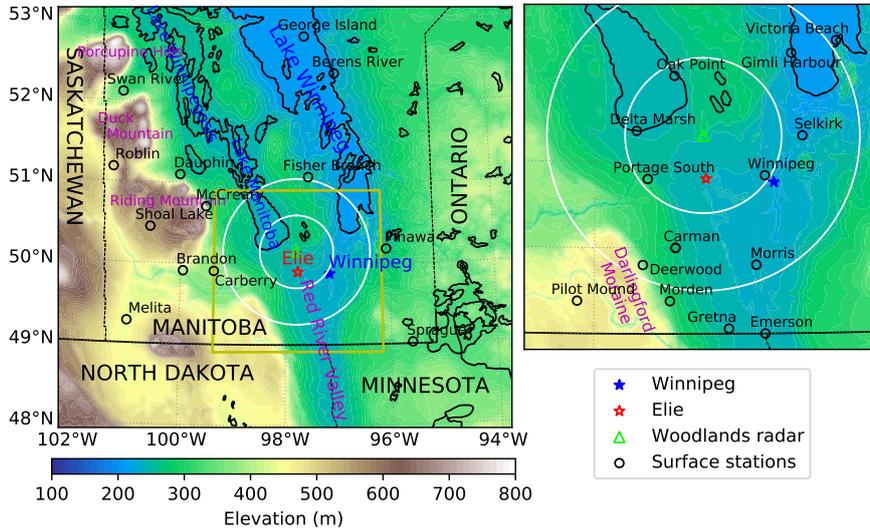
127 **2 Data and Methods**

128 **2.1 Observational datasets**

129 *2.1.1 The study area*

130 The topography of south-central Manitoba is characterized by the relatively flat  
 131 croplands in the Red River Valley (RRV hereafter; Fig. 1). The provincial capital, Win-  
 132 nipeg, is situated near the valley base (elevation  $\sim 230$  m above sea level, ASL), with  
 133 Elie located about 40 km to the west of the city. Two large, south-north-oriented lakes,  
 134 Lakes Winnipeg and Manitoba, lay  $\sim 60$  km to the north-northeast and  $\sim 80$  km to  
 135 the northwest of Winnipeg, respectively. Lake breezes frequently occur during the sum-  
 136 mer along the lake shores (Curry, 2015), with their frontal updrafts having the poten-  
 137 tial to trigger deep moist convection.

138 The western slope of the RRV features taller terrain than its eastern slope, with  
 139 the Porcupine Hills, Duck Mountain, and Riding Mountain, west of the RRV. Each moun-  
 140 tain complex rises to 600-800 m ASL (or 400-600 m from the valley base). Manitoba ter-  
 141 rain generally has been thought to have limited impacts on the regional convection pat-  
 142 tern due to its shallowness (Erfani, 1999).



**Figure 1.** Elevation maps (based on the 2-min U.S. Geological Survey topography data used in the WRF simulation) of southern Manitoba and locations where surface station data was available. The right panel shows the zoomed-in area within the yellow box in the left panel. The white circles denote the 50- and 100-km Woodlands, MB radar range rings.

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### 2.1.2 Radar

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The Environment and Climate Change Canada’s (ECCC), 5-cm, C-band Doppler operational weather radar at Woodlands, MB provided radar coverage for much of south-central Manitoba (Fig. 1). 10-min scans of reflectivity and radial velocity at the  $0.5^\circ$  elevation angle between 1500 UTC 22 June 2007 to 0200 UTC 23 June 2007 were used in this study. These fields have  $0.5^\circ$  azimuthal and 0.5 km radial resolutions and a maximum range of about 113 km from the radar. Radar signals can be contaminated by ground clutter, velocity aliasing, or other artifacts such as dual pulse repetition frequency velocity errors (Joe & May, 2003; Fabry, 2015). ECCC filtered out ground clutter from the reflectivity data and unfolded the radial velocity up to  $48 \text{ m s}^{-1}$ . To eliminate spurious noise in the velocity data, a 3-by-3 median filter like that implemented in Mahalik et al. (2019) was applied. Although volumetric scans were also available from this radar, they were not used because they were too noisy and did not provide any additional insight into identifying the storm-triggering mechanism(s). The volumetric data were not used to investigate the 3D storm structure since this is beyond the scope of this paper.

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### 2.1.3 Satellite

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The visible channel imagery from the National Oceanographic and Atmospheric Administration’s (NOAA) Geostationary Operational Environmental Satellite (GOES) 12 was examined to identify the presence of any shallow cumulus field where the low-level lift is locally enhanced, which may indicate a low-level mesoscale boundary (Purdom, 1976; Sills et al., 2011; Alexander et al., 2018). In this study, the satellite images produced by NOAA between 2000 UTC 22 June 2007 and 0000 UTC 23 June 2007 were used.

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### 2.1.4 Sounding

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The closest operational rawinsonde observations are International Falls, MN, and Bismarck, ND. However, both are  $\sim 400$  km away from Elie, thus the environment sampled there might not be representative. Fortunately, the Prairie and Arctic Storm Prediction Centre (PASPC) released a special sounding from Winnipeg (XWI) at 1800 UTC of the Elie, MB tornado day (22 June 2007). Air temperature, relative humidity, pressure, wind, and height profiles were sampled.

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### 2.1.5 Surface stations

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Hourly surface observations at 27 stations in the southern half of Manitoba (south of  $53^\circ\text{N}$ ) were used to diagnose the regional surface weather conditions before storm initiation between 1200 UTC and 2100 UTC 22 June 2007 (Fig. 1). Most of these stations were standard automated weather stations at airports that collected quality-controlled surface air temperature, dew point, pressure, wind speed, and wind direction. A few were temporary stations used in other field studies that no longer operate today (e.g., Delta Marsh) or private stations installed by the local farmers (e.g., Morris). All fields are assumed taken at the standard heights above ground level (2 m AGL for temperature, dew point, and pressure; 10 m AGL for wind).

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## 2.2 WRF simulation setup

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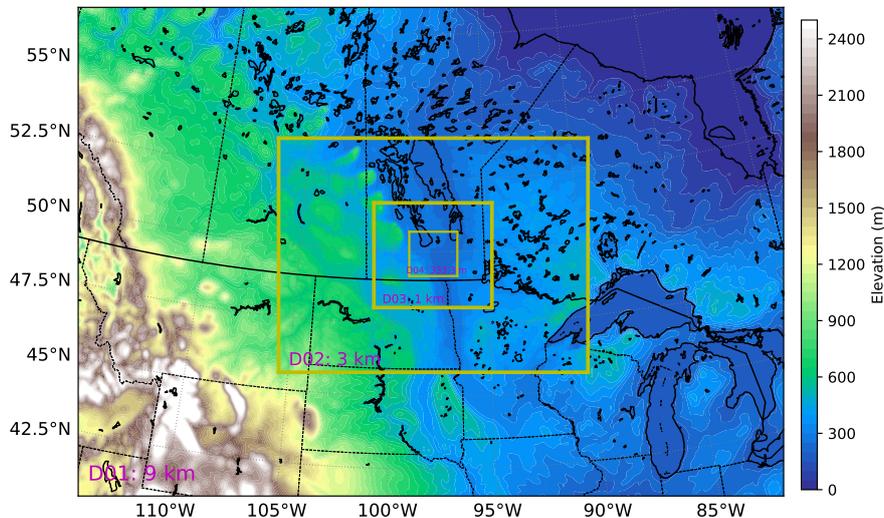
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The Weather Research and Forecasting (WRF) model version 4.2.1 (Skamarock et al., 2019) was used to simulate the three-dimensional synoptic and mesoscale flows during the Elie tornado event. The Runge-Kutta 3rd-order scheme was used to integrate the model’s 3D moist atmospheric equations in time, with horizontal and vertical advection computed by 5th- and 3rd-order schemes, respectively. The scalar advection is positive-definite. The 6-hourly,  $0.5^\circ$  resolution Global Forecasting System (GFS) model analyses were used to initialize the simulation and update the lateral boundary condi-

190 tions. The model was integrated for 18 hr starting at 1200 UTC 22 June 2007 to cap-  
 191 ture the morning to late evening periods.



**Figure 2.** The WRF simulation domains (the nested domains are represented by the boxed regions) with the terrain elevation indicated in filled contours.

192 Four, two-way nested domains (D01-D04) were used in the simulation (Fig. 2), with  
 193 horizontal grid spacings  $\Delta x = \Delta y$  decreasing from 9 km, 3 km, 1 km, to 333.3 m, re-  
 194 spectively. A similar grid-spacing configuration had been used in other numerical inves-  
 195 tigations of tornadic supercells (e.g., Pilguy et al., 2019). The inclusion of the sub-kilometer  
 196 D04 was intended to resolve large intracloud eddies (Bryan et al., 2003), lake breeze fronts,  
 197 which are similar to sea breeze fronts (Lyons & Olsson, 1973; Chiba, 1993; Wood et al.,  
 198 1999; Curry et al., 2016), and other boundary layer drafts such as thermals and horizon-  
 199 tal convective rolls (Balaji & Clark, 1998; Weckwerth et al., 1997; Dailey & Fovell, 1999;  
 200 Bryan et al., 2003). A hybrid vertical coordinate was used in that the model levels are  
 201 roughly terrain-following at the ground and gradually relax to isobaric at upper tropo-  
 202 sphere. 115 user-specified model levels were used up to 50 hPa, yielding a nominal verti-  
 203 cal resolution of  $\sim 30$  m in the lowest 1 km and the lowest de-staggered level at  $\sim 20$   
 204 m above the ground. A rigid boundary caps the model top with an implicit gravity wave  
 205 damping layer specified in the uppermost 5 km to prevent spurious wave reflections.

206 Long- and short-wave radiation were parameterized using the Rapid Radiative Trans-  
 207 fer Model for Global Climate Models (RRTM-G) schemes. Land surface processes were  
 208 modeled using the Noah land surface scheme. A Smagorinsky-type closure was used to  
 209 represent horizontal turbulent mixing, whereas vertical mixing was handled by the Mellor-  
 210 Yamada-Janjic (MYJ) planetary boundary layer (PBL) scheme. The surface layer was  
 211 parameterized using the Janjic scheme based on the Monin-Obukhov similarity theory  
 212 and Zilitinkevich thermal roughness length. A sensitivity experiment by varying the PBL-  
 213 surface layer scheme pairs, namely the Yonsei University, quasi-normal scale elimination  
 214 (QNSE), Mellor-Yamada Nakanishi and Niino, and Shin-Hong schemes, was also con-  
 215 ducted. The Thompson graupel scheme was chosen to parametrize the microphysics, with  
 216 the cloud droplet concentration set to the typical continental value of  $300 \text{ cm}^{-3}$  accord-  
 217 ing to the scheme’s recommendation. The Grell-Freitas cumulus parameterization was  
 218 used for the 9-km resolution domain (D01) only.

219 As mentioned, Manitoba lakes frequently generate lake breezes during summer. Be-  
 220 cause lake breezes are driven by the land-water temperature contrast (Lyons, 1972; Cros-  
 221 sman & Horel, 2012), the proper initialization of the lake surface temperatures ( $T_{lake}$ )  
 222 in the WRF simulation is crucial to reasonably capture lake effects.  $T_{lake}$  was initialized  
 223 using the daily-averaged surface air temperature using the 6-hourly GFS analyses be-  
 224 tween 0000 UTC and 1800 UTC 22 June 2007. This was done because the lakes in Man-  
 225 itoba are either too small or narrow ( $< 0.5^\circ$  latitudinally or longitudinally) to be resolved  
 226 in the  $0.5^\circ$  resolution GFS analyses, introducing the potential for WRF to poorly ini-  
 227 tialize  $T_{lake}$  (readers are referred to Chapter 3-27 of the WRF Users' Guide: [https://](https://www2.mmm.ucar.edu/wrf/users/docs/user_guide_v4/v4.2/WRFUsersGuide_v42.pdf)  
 228 [www2.mmm.ucar.edu/wrf/users/docs/user\\_guide\\_v4/v4.2/WRFUsersGuide\\_v42.pdf](https://www2.mmm.ucar.edu/wrf/users/docs/user_guide_v4/v4.2/WRFUsersGuide_v42.pdf)  
 229 for a detailed explanation of this issue). The comparison of  $T_{lake}$  in Manitoba initial-  
 230 ized with this approach vs. those without against the National Aeronautics and Space  
 231 Administration (NASA)'s Group for High-Resolution Sea Surface Temperature (GHRSSST;  
 232 <https://worldview.earthdata.nasa.gov/>) dataset values on 22 June 2007 shows that  
 233 the former were generally within  $2^\circ\text{C}$  of the GHRSSST values, whereas the latter were  
 234 about  $10^\circ\text{C}$  too cold. Therefore, the surface air temperature-initialized  $T_{lake}$  was rea-  
 235 sonable and was fixed throughout the simulation.

### 236 3 Event Overview

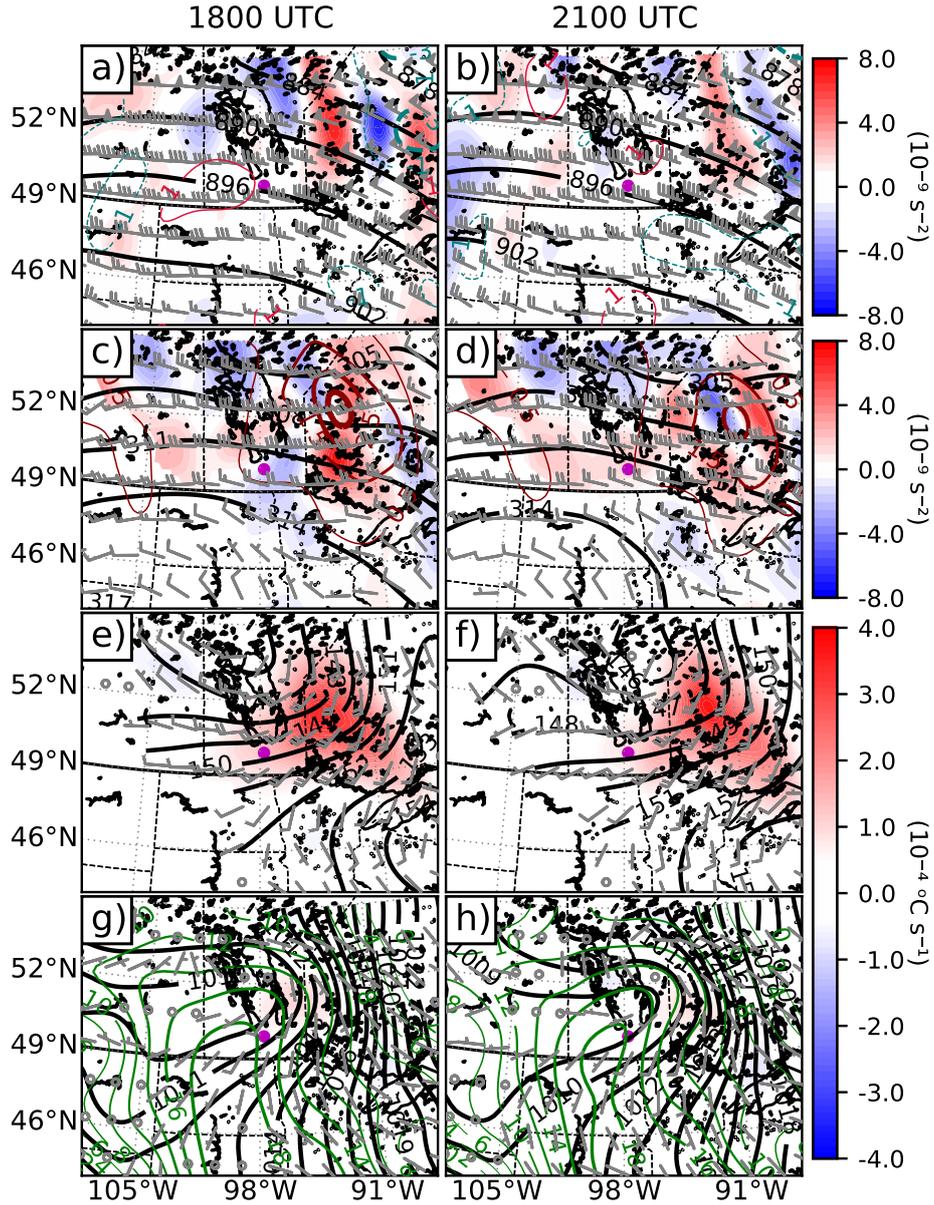
#### 237 3.1 Large-scale pattern

238 The hourly, fifth-generation European Centre for Medium-Range Weather Fore-  
 239 casts reanalysis (ERA5; Hersbach et al., 2020) was used to examine the regional synoptic-  
 240 scale flow on 22 June 2007. ERA5 has been found to reasonably depict observed weather  
 241 patterns and convective environments in various parts of the world (e.g., Balsamo et al.,  
 242 2018; Coffey et al., 2020; F. Li et al., 2020; Taszarek et al., 2021; Pilguy et al., 2022). The  
 243 20-km Rapid Update Cycle (RUC) model analysis was also examined as additional sup-  
 244 porting data. Both datasets showed similar diurnal evolution of large-scale patterns, there-  
 245 fore, only the ERA5 is discussed.

246 The Elie event featured a broad upper-level ridge above 500 hPa over the Cana-  
 247 dian Prairies (Fig. 3a, b), with no significant diffluence and jet streak influences at 200  
 248 hPa or vorticity advection at 500 hPa. Thus, very little to no upper-level (500 hPa or  
 249 above) forcing for vertical motion was present before storm initiation.

250 At 850 (700) hPa, 20-25 kt southwesterly (west-northwesterly) at  $220^\circ$  ( $290^\circ$ ) winds  
 251 advected warm air into southern Manitoba behind a shortwave trough moving eastward  
 252 across northern Manitoba in the morning (Figs. 3c-f), with associated isolated precip-  
 253 itation passing over the RRV (not shown). The warm air advection (WAA) partly con-  
 254 tributed to the stout 900-700 hPa capping inversion in the 1800 UTC XWI sounding (Fig. 4a).  
 255 Analysis of 850-700 hPa vorticity advection reveals that anti-cyclonic vorticity advec-  
 256 tion (AVA) with height existed over southern Manitoba until 1900 UTC, favoring large-  
 257 scale subsidence and mid-level capping (Fig. 3c). Differential vorticity advection flipped  
 258 to cyclonic across the 850-700 hPa layer by 2100 UTC, leading to forcing for ascent (Fig. 3d).  
 259 Cyclonic differential vorticity advection, solar heating, and low-level warm air and mois-  
 260 ture advection following an early-morning warm frontal passage (Figs. 3g, h) were likely  
 261 the dominant mechanisms of lower-troposphere lift and destabilization in the afternoon  
 262 over southern Manitoba.

263 By 2100 UTC, the warm front's parent low pressure was approaching the Elie area  
 264 from the north along with its associated cold front. Southern Manitoba was situated within  
 265 the warm sector of this surface low.



**Figure 3.** Left column: maps of a) 500-hPa vorticity advection (filled), 200-hPa divergence ( $10^{-5} \text{ s}^{-1}$ ), 200 and 500 hPa averaged geopotential heights (black solid; in dm), and wind barbs (full barbs denote 10 kt and half barbs denote 5 kt; all plotted barbs will have this same convention), c) 700-hPa geopotential heights (black solid), temperature advection ( $10^{-4} \text{ }^\circ\text{C s}^{-1}$ ), the differential absolute vorticity advection between 850-700 hPa (filled), and wind barbs, e) 850-hPa temperature advection (filled), geopotential heights (solid), and wind barbs, and g) surface temperature advection (filled), sea-level pressure (black solid), dew point (green solid;  $^\circ\text{C}$ ), and wind barbs at 1800 UTC 22 June 2007. Right column: the same as in the left column, but for 2100 UTC 22 June 2007.

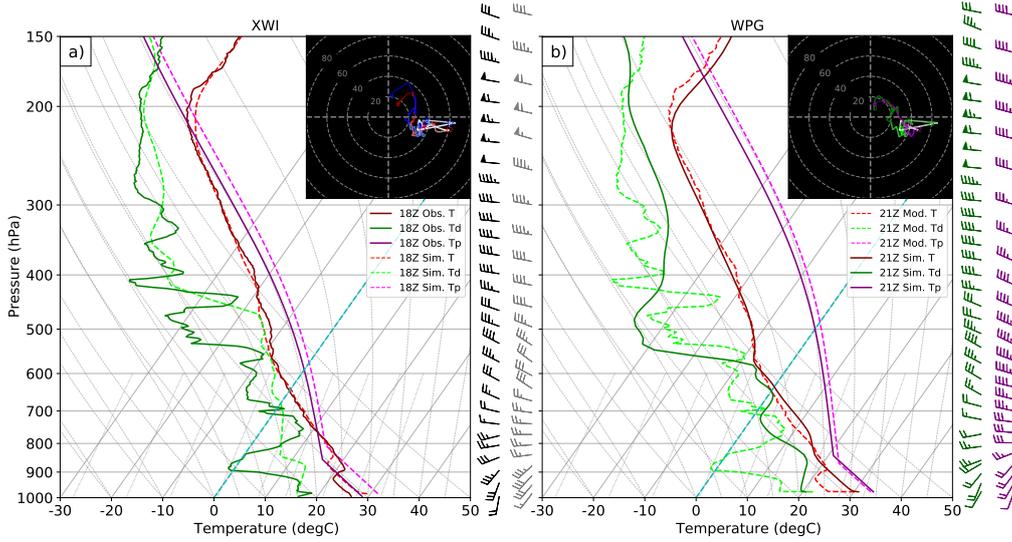
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### 3.2 Pre-storm initiation environment

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CAPE and CIN were computed for the surface-based (SB), lowest 100-mb mixed layer (ML), and lowest 300-mb most-unstable (MU) parcels for the observed XWI sound-



**Figure 4.** a) Observed and WRF simulated 1800 UTC 22 June 2007 soundings from XWI. The observed wind profile (hodograph) is indicated in black (blue) while the simulated wind profile is indicated in gray (red). b) The modified 1800 UTC XWI (with 2100 UTC WPG surface observations and Woodlands radar VAD winds up to 1.5 km ASL) and simulated soundings at 2100 UTC 22 June 2007 at WPG. The modified wind profile and hodograph are indicated in green while the simulated ones are shown in purple. All parcel temperature profiles are for surface-based parcels.

269 ing (Table 1). BWD and SRH were calculated for various fixed layers above the ground  
 270 (0-1 km and 0-6 km for BWD; 0-3 km for SRH), as well as the ‘effective’ layer (EBWD  
 271 and ESRH (R. L. Thompson et al., 2007). SRH was obtained using the Bunkers et al.  
 272 (2014) storm motion of 17 kt ( $\sim 9 \text{ m s}^{-1}$ ) from  $300^\circ$  derived from the 1800 UTC XWI  
 273 sounding. Lastly, a lower mixed-layer lifting condensation level (MLLCL) often suggests  
 274 a higher boundary-layer relative humidity, which prevents strong outflow from forming  
 275 to cut off storm inflow and increase tornado probability (E. N. Rasmussen & Blanchard,  
 276 1998; Markowski et al., 2002). This parameter was also computed. The convective pa-  
 277 rameters were calculated using SHARPPy (Blumberg et al., 2017).

278 Previous proximity sounding studies of tornadic supercell environments in North  
 279 America have found that these storms are typically associated with environmental ML-  
 280 CAPE or MUCAPE  $> 1000 \text{ J kg}^{-1}$ , SBCIN or MLCIN  $> -50 \text{ J kg}^{-1}$ , 0-1 km BWD  
 281  $> 10 \text{ kt}$ , 0-6 km BWD or EBWD 30-40 kt, 0-3 km SRH or ESRH  $> 100 \text{ m}^2 \text{ s}^{-2}$ , and  
 282 MLLCL  $< 1500 \text{ m AGL}$  (R. L. Thompson et al., 2003, 2012; Taszarek et al., 2020; Hane-  
 283 siak et al., 2023). Based on the parameters shown in Table 1, local weather forecasters  
 284 were concerned about tornadic supercells developing. However, a major uncertainty was  
 285 whether the large cap (MLCIN  $< -50 \text{ J kg}^{-1}$ ) would be reduced enough for convection  
 286 initiation. The combination of large CIN and high MLLCL ( $> 1500 \text{ m}$ ) also suggested  
 287 that any developed supercell is likely going to be elevated (Coleman, 1990) and produce  
 288 a strong outflow, both of which reduce tornado probability (E. N. Rasmussen & Blan-  
 289 chard, 1998; Davies, 2004; R. L. Thompson et al., 2012).

290 To project the mesoscale environment just before the Elie supercell initiation, we  
 291 modified the 1800 UTC observed sounding using the 2100 UTC surface observations at  
 292 Portage la Prairie, MB (WPG) and the Woodlands radar velocity azimuthal display (VAD)-

**Table 1.** Mesoscale storm parameters calculated using the observed, WRF, and ERA5 Winnipeg (XWI) soundings at 1800 UTC 22 June 2007. The 2100 UTC modified and WRF simulated sounding parameters at Portage la Prairie (WPG) are also shown.

	1800 UTC XWI obs.	1800 UTC XWI WRF	1800 UTC XWI ERA5	2100 UTC WPG mod.	2100 UTC WPG WRF
SBCAPE/SBCIN (J kg <sup>-1</sup> )	1010/-162	1641/-32	2004/-71	5581/0	4738/0
MLCAPE/MLCIN (J kg <sup>-1</sup> )	309/-226	1299/-73	713/-153	43/-286	3687/-2
MUCAPE/MUCIN (J kg <sup>-1</sup> )	2187/-71	1641/-32	2004/-71	5581/0	4738/0
0 – 1 km BWD (kt)	23	18	19	15	10
0 – 6 km BWD (kt)	39	33	36	30	43
EBWD (kt)	39	32	36	34	42
0 – 3 km SRH (m <sup>2</sup> s <sup>-2</sup> )	296	200	219	175	230
ESRH (m <sup>2</sup> s <sup>-2</sup> )	130	200	144	98	215
MLLCL (m AGL)	1595	1529	1434	1996	1319

293 derived winds (VAD data quality above 1.5 km ASL appears to worsen rapidly and thus  
 294 was discarded and substituted with the 1800 UTC XWI sounding winds). The surface  
 295 site and time were chosen since they were the closest to the Elie supercell initiation (see  
 296 section 3.3.1) yet still free of convection contamination.

297 The modified sounding shows that the capping likely eroded away by late afternoon  
 298 (Fig. 4b), with the SBCAPE and MUCAPE exceeding 5500 J kg<sup>-1</sup> (Table 1). The 0-  
 299 6 km BWD and EBWD (30-35 kt) and 0-3 km SRH and ESRH (100-200 m<sup>2</sup> s<sup>-2</sup>) remained  
 300 sufficient for tornadic supercells. However, because the modified sounding neglects the  
 301 afternoon boundary-layer warming and moistening as well as the upper-level flow evo-  
 302 lution, it only serves as a first guess of the local convective environment. Further anal-  
 303 ysis of the environment immediately before the Elie supercell initiation is provided in  
 304 section 5.1 using the WRF simulation.

### 305 **3.3 Storm initiation and the subsequent mesoscale evolution**

306 In this subsection, we describe the radar and satellite observations during the late  
 307 afternoon to early evening of 22 June 2007 with a focus on identifying the potential low-  
 308 level trigger(s) of the Elie supercell. The broad storm evolution after the Elie supercell  
 309 initiation is also briefly described. For a detailed description of the observed storm struc-  
 310 ture and evolution, readers are referred to Hobson (2011).

#### 311 **3.3.1 Storm initiation**

312 At 2030 UTC 22 June 2007, a low-level mesoscale boundary (B1) was detected by  
 313 radar to the southwest of Lake Manitoba (Figs. 5b, c). Its reflectivity and radial veloc-  
 314 ity shift signatures were faint at this time due to the radar beam altitude in relation to  
 315 the depth of the low-level convergence (see the radar animation in the supplemental ma-  
 316 terial). To the east of B1, lines of weak reflectivity, which we identified as horizontal con-  
 317 vective rolls (HCRs; Weckwerth et al., 1997; Yang & Geerts, 2006), can be seen roughly

318 aligned with the southwesterly surface flow across south-central Manitoba (only one is  
319 labeled as B2 for simplicity).

320 As B1 moved eastward, it became more apparent in the reflectivity and radial ve-  
321 locity scans (Figs. 5e, f). Deep cumulus convection began to develop along B1 (Fig. 5d).  
322 The Elie supercell (S1) was first detected on radar at 2200 UTC (Fig. 5e), rooted along  
323 B1 about 10 km to the northeast of WPG (Fig. 1). Two LBFs were identified along the  
324 southern shore of Lake Manitoba (labeled as B3 and B4). B3 brought an onshore (south-  
325 westerly to west-northwesterly) wind shift, a 1°C temperature drop, and a 3% relative  
326 humidity increase at Delta Marsh (Fig. 1) after it passed between 2100 UTC and 2200  
327 UTC (not shown). A cold front approached the area from the north (see section 3.1) but  
328 it was not detected on radar well after S1 had initiated (B5 in Figs. 5h, k). Ahead of the  
329 cold front, the surface temperatures soared into low 30°Cs while the dew point reached  
330 low 20°Cs across southern Manitoba.

### 331 **3.3.2 Storm evolution after initiation**

332 After initiating, S1 intensified and moved eastward until it was located about 10  
333 km north-northwest of Elie, with its anvil and precipitation shield blown to the south-  
334 east (Figs. 5g-i). S1 turned right relative to its original motion (east to south-southeast)  
335 between 2310 UTC and 2320 UTC while forming supercell structures, including a rota-  
336 tional couplet, hook echo, and wall cloud (Figs. 5k, l, 6a; Browning, 1977; Rotunno &  
337 Klemp, 1982; Weisman & Klemp, 1984; Klemp, 1987; Burgess & Lemon, 1990; Weisman  
338 & Rotunno, 2000; Davies-Jones, 2002; Markowski & Richardson, 2010). Two other promi-  
339 nent thunderstorms also had formed near S1 (S2 and S3; Fig. 5h). Between 2320 UTC  
340 and 2330 UTC, S1 produced a funnel cloud that reached the ground shortly after to form  
341 the Elie tornado (Figs. 6b, c). The tornado then struck Elie at 2350 UTC (Fig. 6d). At  
342 around the same time, S3 also matured into a supercell and produced an F3 tornado near  
343 Oakville, MB (Fig. 6e), while S2 weakened as a supercell without producing a tornado  
344 (not shown).

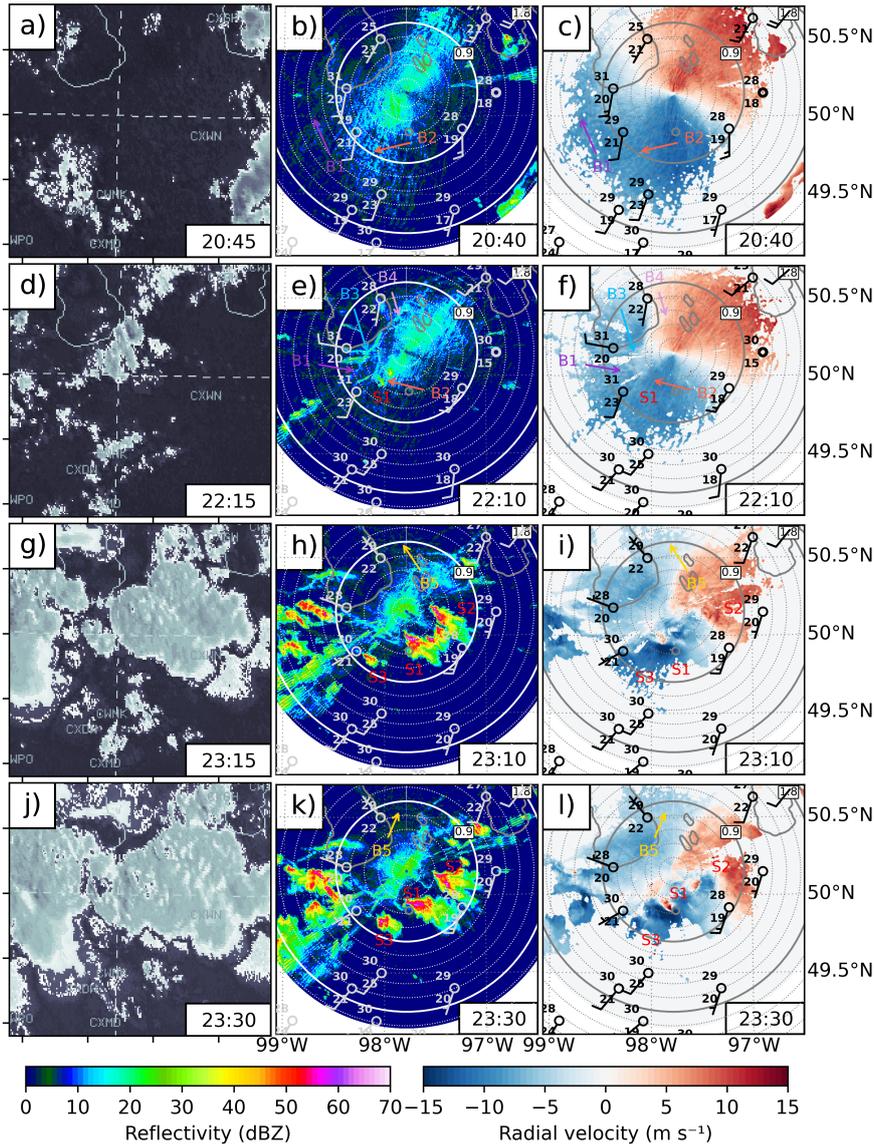
345 The radar analysis suggests that a low-level mesoscale boundary may be the pri-  
346 mary triggering mechanism of the Elie supercell. However, the identity of this trigger-  
347 ing boundary and the relative importance of the other boundaries (i.e., HCRs and LBFs)  
348 in storm initiation remain unclear. The real-case WRF simulation of this event was used  
349 to address two outstanding issues due to the lack of observations: 1) to better under-  
350 stand the local convective environment immediately before the Elie supercell initiation  
351 and 2) to identify and characterize the storm-triggering mechanism(s) of the Elie super-  
352 cell. The simulation results follow.

## 353 **4 Simulation overview and verification**

354 We devote this section to the WRF simulation’s performance in reproducing the  
355 Elie tornado event, focusing on the flow evolution leading up to the Elie supercell ini-  
356 tiation. All PBL scheme sensitivity experiments exhibited similar synoptic- and mesoscale  
357 features (not shown), with the MYJ and QNSE scheme producing the strongest discrete  
358 supercells. The results from a simulation employing only D01-D03 with the MYJ and  
359 Eta schemes were found to be similar to those using all four domains, suggesting that  
360 1 km grid-spacing was sufficient to resolve the processes contributing to the supercell ini-  
361 tiation. For the ease of computation, the former was chosen for in-depth analysis.

### 362 **4.1 Overview of the simulated flow**

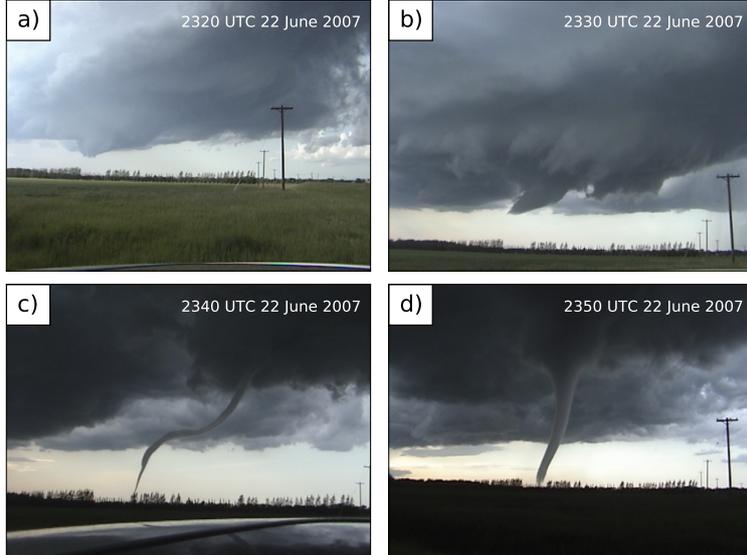
363 We begin by providing an overview of the simulated flow in southern Manitoba on  
364 22 June 2007, focusing on the cloud and precipitation patterns during the late afternoon  
365 to early-evening periods. The simulated cloud top temperature was used to compare with



**Figure 5.** Observed visible satellite imagery (left column), radar reflectivity (center column), and radial velocity (right column) at the  $0.5^\circ$  elevation angle at a-c) 2040, d-f) 2210, g-i) 2310, and j-l) 2330 UTC 22 June 2007. Radar-detected mesoscale boundaries and supercells are annotated and labeled. Elie, MB is indicated by a gray circle. Station models in the area are also displayed. In the radar columns, the thick circles indicate the 50- and 100-km range rings, with the altitude (km ASL) at these ranges indicated in the boxes.

366 the visible satellite observations, while the simulated reflectivity at 1 km ASL was used  
 367 to compare with the observed  $0.5^\circ$  reflectivity.

368 At 2040 UTC, a low-level convergence line with a few clouds was simulated to the  
 369 southwest of Lake Manitoba (Fig. 7a, b), with the simulated 2-m air temperature and  
 370 dew point around Elie reaching  $28\text{--}31^\circ\text{C}$  and  $18\text{--}21^\circ\text{C}$ , respectively, similar to observa-  
 371 tions (next section). Deep moist convection, including the precursor simulated Elie storm  
 372 (SS1), began to initiate along the low-level convergence line at 2130 UTC,  $\sim 40$  min ear-



**Figure 6.** Photos of a) the Elie tornado’s parent supercell, b) the Elie supercell with a funnel cloud, c) the Elie tornado touching down just north of Elie, d) and the Elie tornado in Elie. The camera shot times are indicated on the top right of each photo.

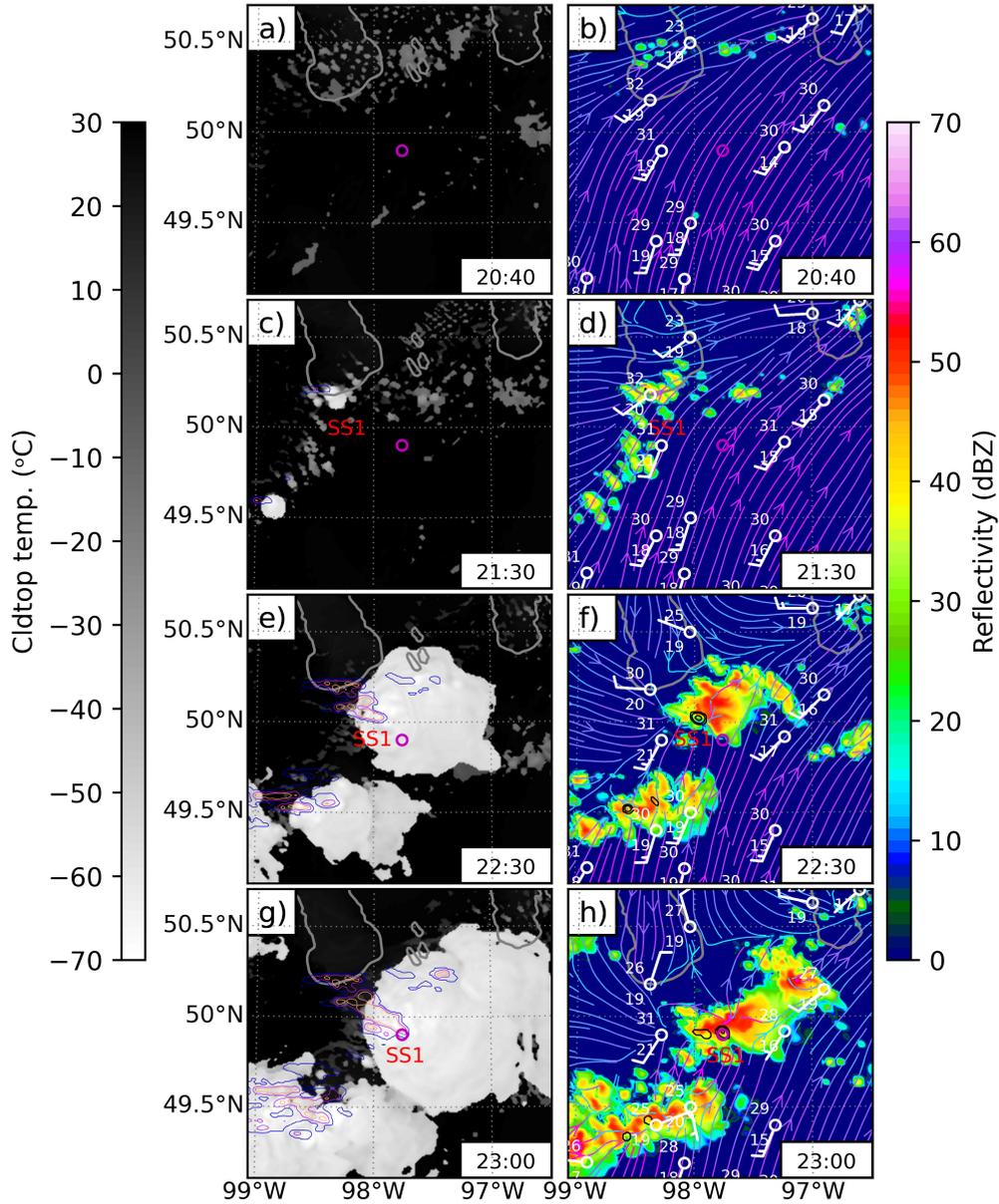
373 lier and  $\sim 40$  km farther west than that observed (Figs. 7c, d). The winds near Elie to  
 374 the east of the convergence line were generally southwesterly with more westerly winds  
 375 west of the line.

376 The simulated 2-5 km AGL updraft helicity (UH; Sobash et al., 2011; Naylor et  
 377 al., 2012; Loken et al., 2017) was used to indicate mid-level mesocyclone presence in our  
 378 simulation. After initiating, SS1 propagated eastward and developed an UH core on its  
 379 western flank by 2230 UTC, with its anvil sheared to the southeast (Figs. 7e, f). As with  
 380 the observed storm, the accumulated 10-min maximum UH track (since 2000 UTC 22  
 381 June 2007) suggests that SS1 traveled eastward and then southeastward (crossref Figs. 7  
 382 and 5); reaching Elie at around 2300 UTC,  $\sim 50$  min earlier than the observed (Fig. 7g,  
 383 h). Some vigorous cells were also simulated to the north and south of SS1. A few of these  
 384 cells briefly exhibited mid-level rotation, agreeing with the observed convection distri-  
 385 bution where multiple supercells existed simultaneously around the Elie cell.

#### 386 4.2 Simulated surface conditions

387 The simulation root-mean-squared error (RMSE), bias, and index of agreement (IOA)  
 388 for surface air temperature ( $T_{2m}$ ), dew point ( $Td_{2m}$ ), wind speed ( $V_{10m}$ ), and wind di-  
 389 rection ( $\angle V_{10m}$ ) were computed using the hourly observed and simulated values at all  
 390 surface stations described in section 2.1.5 and presented in Table 2. The differences be-  
 391 tween the observed and forecast  $\angle V_{10m}$  were adjusted using the method described in (Lascaux  
 392 et al., 2013) to account for angular measurements. Wind measurements at non-standard  
 393 meteorological stations (i.e., Morris and Selkirk, MB) were excluded to avoid possible  
 394 data quality issues. The evaluation period was between 1200 UTC to 2100 UTC 22 June  
 395 2007 to focus on the pre-Elie storm period.

396 Based on the values in Table 2, WRF appeared to reasonably capture the observed  
 397 surface conditions during the period examined, with the variables’ RMSE and bias gener-  
 398 ally small and IOA  $> 0.70$ . The largest errors exist in  $Td_{2m}$  and  $V_{10m}$  likely due to



**Figure 7.** Left: The WRF simulated cloud top temperature (filled) and accumulated 10-min maximum 2-5 km AGL updraft helicity (solid; in contours of 100, 300, 500 and 700  $\text{m}^2 \text{s}^{-2}$ ). Right: The WRF simulated 1 km ASL reflectivity (filled), 2-5 km AGL updraft helicity (black solid; in contours of 75, 150, 300, and 500  $\text{m}^2 \text{s}^{-2}$ ), 10-m streamlines, and the simulated conditions on the hour at the surface stations in the area (see Fig. 1). Elie is indicated by a magenta circle. The plotted times (UTC) are indicated on the lower right of each panel.

399 their highly locally varying nature, which was exacerbated by the lack of observations.  
 400 The model's misrepresentation of the observed early-day precipitation likely partly con-  
 401 tributed to the remaining errors. Immediately before the observed Elie supercell initi-  
 402 ation, the WRF simulated  $T_{2m}$ ,  $Td_{2m}$ ,  $V_{10m}$ , and  $\angle V_{10m}$  appear to agree well with the  
 403 observations near the storm initiation region (not shown).

**Table 2.** Simulation RMSE, bias, and IOA for  $T_{2m}$ ,  $Td_{2m}$ ,  $V_{10m}$ , and  $\angle V_{10m}$  evaluated between 1200 UTC and 2100 UTC 22 June 2007.

	$T_{2m}$	$Td_{2m}$	$V_{10m}$	$\angle V_{10m}$
RMSE	1.9	2.4	1.9	46.1
Bias	0.3	-1.3	0.4	5.8
IOA	0.96	0.80	0.70	0.80

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### 4.3 Winnipeg sounding and the large-scale pattern

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The 1800 UTC WRF simulated sounding at the nearest model grid point to the Winnipeg radiosonde launch site was used to compare against the observed sounding (Fig. 4a). The simulated 1800 UTC CAPE (CIN) at Winnipeg were generally 600-1000 (50-150)  $J\ kg^{-1}$  larger (smaller) than the observed, except for MUCAPE (Table 1). These discrepancies may be partly owing to a 2-3°C simulation warm bias below 900 hPa and mesoscale ascent ahead of the simulated early-afternoon convection near Winnipeg temporarily weakening the low-level cap (not shown). After the simulated convection exited the area by 1830 UTC, the cap quickly rebuilt over Winnipeg before eroding away (not shown).

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The simulated BWD and SRH at XWI at 1800 UTC 22 June 2007 were overall weaker than the observed (except for ESRH), especially for the simulated 0-3 km SRH ( $\sim 30\%$  too weak). The error may be partly due to the simulated wind direction being more westerly than the observed from the surface up to 800 hPa (hence less veering; Fig. 4a). In their studies of European and Canadian tornadic storm environments, Taszarek et al. (2021) and Hanesiak et al. (2023) found that ERA5 reasonably captured the observed vertical atmospheric profiles and the derived convective parameters in these regions. For the Elie event, the 1800 UTC WRF-derived BWD and SRH at XWI were comparable to those calculated from the ERA5 sounding (Table 1). The temperature advection, pressure (or geopotential height), and convective parameter patterns over the Canadian Prairies were also broadly similar between WRF and ERA5 throughout the day (not shown). Thus, WRF appeared to adequately reproduce the event's large-scale flow.

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## 5 Immediate storm environment and storm-triggering mechanisms

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### 5.1 Immediate storm environment

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To evaluate the local storm environment just before the simulated Elie supercell initiation at 2130 UTC 22 June 2007, we examined the 2100 UTC 22 June 2007 simulated convective parameters near the storm initiation site at WPG. For the sounding analysis, the data from D03 was used, while the data from D01 was used for the map analysis unless noted otherwise. To demonstrate the significance of the simulated environmental parameters in terms of the tornadic supercell environments, we also computed the supercell composite parameter (SCP) and significant tornado parameter (STP):

$$SCP = \frac{MUCAPE}{1000\ J\ kg^{-1}} \frac{EBWD}{20\ m\ s^{-1}} \frac{ESRH}{50\ m^2\ s^{-2}} \quad (1)$$

$$STP = \frac{MLCAPE}{1500\ J\ kg^{-1}} \frac{EBWD}{20\ m\ s^{-1}} \frac{ESRH}{150\ m^2\ s^{-2}} \frac{2000\ m - MLLCL}{1000\ m} \frac{MLCIN + 200\ J\ kg^{-1}}{150\ J\ kg^{-1}} \quad (2)$$

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, where all terms have been previously defined. Readers shall refer to R. L. Thompson et al. (2004c) and R. L. Thompson et al. (2012) for the conditions that apply to the right-hand-side terms of both equations.

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**5.1.1 The 2100 UTC simulated WPG sounding**

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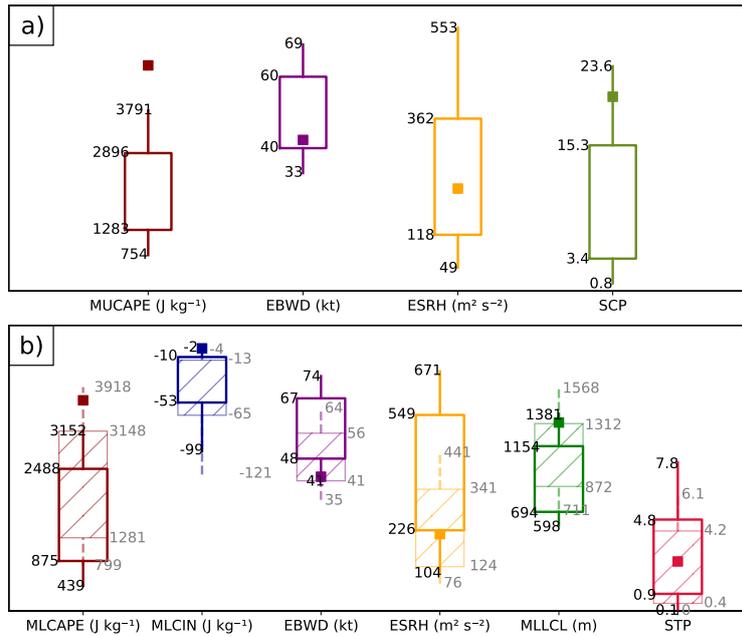
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The SCP and STP computed based on the 2100 UTC simulated sounding (Figs. 4b and Table 1) were 20 and 2.6, respectively. Compared to other tornado events in the U.S., the former was well above the 25th-75th percentile range (the interquartile range, hereafter) for discrete, right-moving tornadic (RMdT) supercell cases, while the latter fell within the interquartile range (i.e., the typical range) for significantly tornadic (EF2+) RMdT (sigtor) supercells (Fig. 8; R. L. Thompson et al., 2012). The large SCP and STP were mainly a result of the large MUCAPE and MLCAPE as they both exceeded the typical values found in the RMdT and sigtor supercell cases in the U.S. ( $> 3500 \text{ J kg}^{-1}$ ), while EBWD and ESRH both fell towards the lower end of the distributions ( $< 45 \text{ kt}$ ), especially for sigtor supercells (Fig. 8). The Elie event also featured a higher MLLCL ( $> 1300 \text{ m AGL}$ ) than the majority of sigtor supercell events in the U.S. (Fig. 8b). Similar trends were found when comparing the 2100 UTC simulated SCP and STP and their constituent parameters to those derived based on the Canadian sigtor events (see Hanesiak et al. (2023)). However, if comparing only against the summertime sigtor supercell cases in the U.S., the 2100 UTC simulated EBWD, ESRH, and MLLCL at WPG were rather typical for these events (Fig. 8b; R. L. Thompson et al., 2012). The middle-to-high-end simulated SCP and STP compared to those typically found in other significant tornado events in North America suggests that the late-afternoon environment near Elie posed a substantial threat of sigtor supercells.



**Figure 8.** Box-whiskers (the boxes denote the 25th-75th percentiles; the whisker tips denote the 10th-90th percentiles) of a) SCP and its component parameters for RMdT (discrete, right-moving tornadic) supercells in the U.S. and b) STP and its component parameters for U.S. sigtor (significantly tornadic) supercell events. In b), the hollow boxes with solid whiskers denote the distribution derived based on all-year events regardless of the season and the hatched boxes with dashed whiskers represent those derived based on summertime events only. The 2100 UTC WRF simulated values (texts omitted) at WPG are marked by filled squares. Adapted from R. L. Thompson et al. (2012).

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### 5.1.2 The 2100 UTC simulated convective parameter spatial patterns

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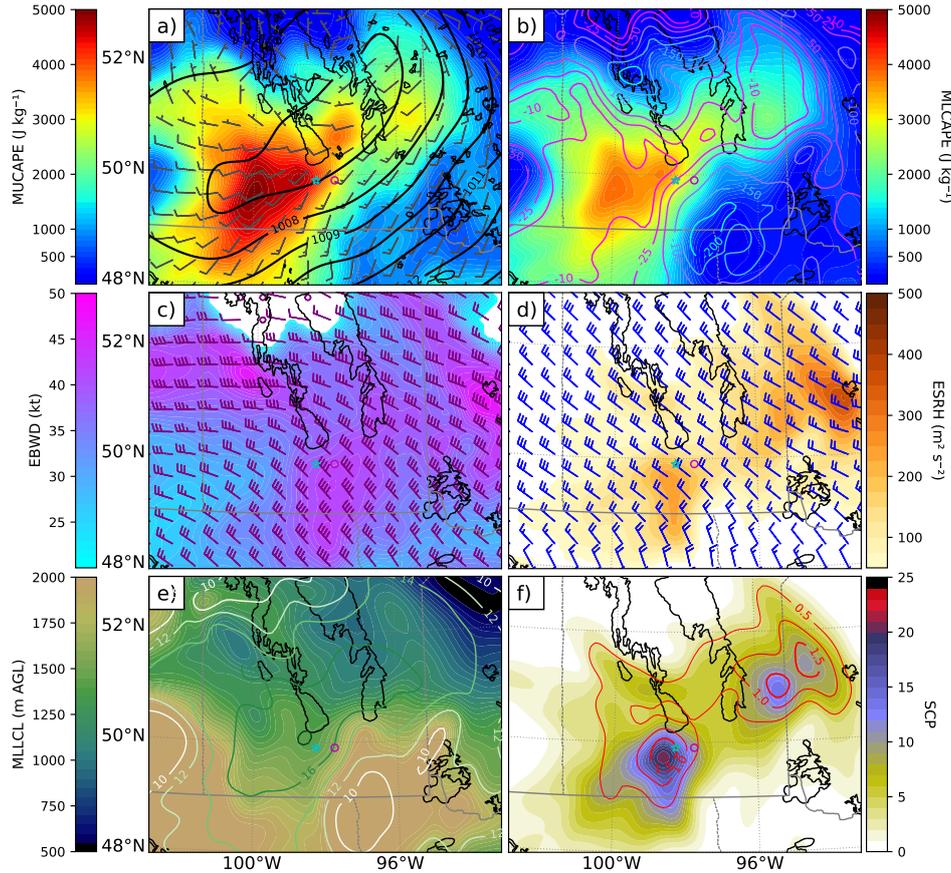
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Figure 9 shows that low-level moisture and wind shear were enhanced over a narrow band stretching from the Manitoba-North Dakota border, over WPG, and into southern Lake Manitoba, hence the locally larger CAPE, lower LCL, and greater EBWD and ESRH. This feature was likely induced by an outflow boundary associated with the simulated convection that moved across southern Manitoba between 1600 UTC and 1800 UTC. Outflow boundaries can locally enhance moisture and/or shear, thereby increasing the likelihood of tornadic supercells (e.g., Maddox et al., 1980). The simulated SCP and STP were indeed enhanced within the band of large conditional instability and shear near WPG relative to the west-east adjacent areas (Fig. 9f).



**Figure 9.** Maps of the WRF simulated a) MUCAPE (filled), sea level pressure (solid), and 10-m wind barbs, b) MLCAPE (filled), MLCIN (solid), c) EBWD speed (filled) and barbs, d) ESRH (filled) and Bunkers storm motion barbs, e) MLLCL (filled) and the lowest 100-mb averaged dew point (solid; in °C), and f) SCP (filled) and STP (solid) at 2100 UTC 22 June 2007. Elie is indicated by a circle, while WPG is indicated by a star.

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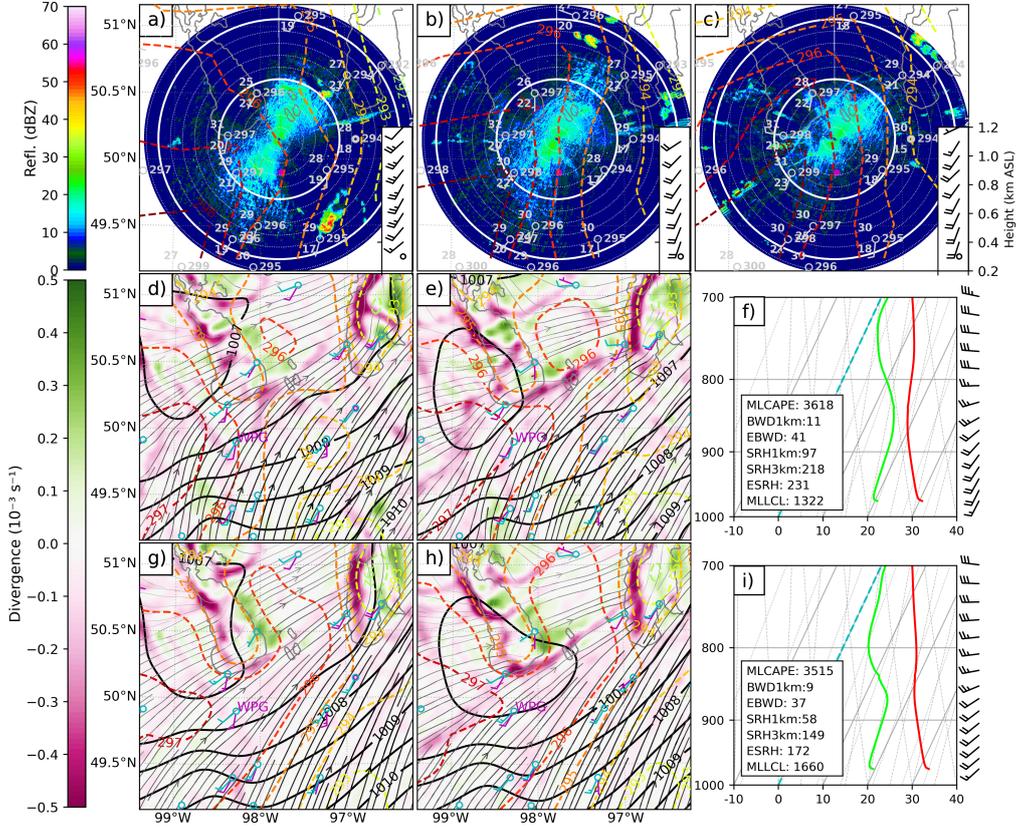
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A cluster of convection was observed propagating across southern Manitoba between 1800 UTC to 2000 UTC 22 June 2007. Here, we investigate whether this convection also locally enhanced the immediate tornado environment near Elie. At 2000 UTC, an area of reduced reflectivity (suppressed HCRs) denoting the cold pool produced by this convection was detected (Figs. 10a, b). The gridded surface wet-bulb potential temperature ( $\theta_w$ ) calculated using the surface observations shows that the cold pool may have tight-

473 ened and orientated the surface  $\theta_w$  contours latitudinally between WPG and Elie, thereby  
 474 influencing the local pressure gradient and wind patterns. The radar VAD shows that  
 475 the low-level winds turned southwesterly within the storm's outflow (Fig. 10a), shifted  
 476 back to south-southwesterly as the cold pool moved east out of the area, and remained  
 477 so until storm initiation (Figs. 10b, c).



**Figure 10.** Observed radar reflectivity (filled), surface  $\theta_w$  (dashed), and station models (temperature: top left, dew point: bottom left, and  $\theta_w$ : center right) at a) 2000 UTC, b) 2100 UTC, and c) 2200 UTC 22 June 2007. The Woodlands, MB radar VAD-derived wind profiles at these times are also shown at the bottom right of these panels. d, e) Simulated 10-m divergence (filled), surface  $\theta_w$  (dashed), SLP (black solid), 10-m streamlines, and the observed (purple) and simulated (cyan) surface station wind barbs at 2000 UTC and 2100 UTC, respectively, from the MP simulation. g, h) Same as in d, e), but from the NOMP simulation. f) Simulated WPG sounding and selected convective parameters from the MP simulation at 2100 UTC 22 June 2007. i) Same as in f), but from the NOMP simulation.

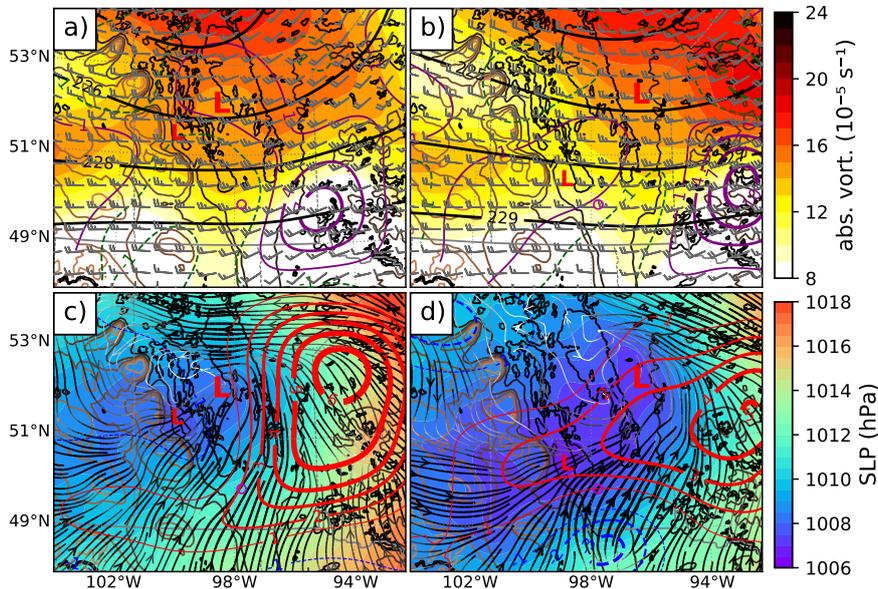
478 To isolate the outflow's effects on the local storm environment, we performed a  
 479 another WRF simulation with the microphysics turned off (NOMP). D02 data was used  
 480 for this analysis. The original simulation (MP) produced similarly orientated surface  $\theta_w$   
 481 contours behind the simulated afternoon convection (Figs. 10d, e), while the NOMP simu-  
 482 lation showed less latitudinally-oriented and farther east-located tight  $\theta_w$  contours (Figs. 10g,  
 483 h). Perhaps as a result, the 2100 UTC simulated winds between the surface and 800 hPa  
 484 at WPG were more southerly in the MP simulation than those in the NOMP simula-  
 485 tion, better agreeing with the observations (Figs. 10e, h; crossref Figs. 10f, i with Figs. 10b,

486 c). The greater low-level veering likely resulted in larger simulated BWD and SRH near  
 487 WPG in the MP simulation vs. the NOMP simulation (Figs. 10f, i).

488 Both the MP and NOMP simulations produced a mesoscale boundary that later  
 489 triggered the Elie storm in the former. The more southerly flow in the MP simulation  
 490 more strongly interacted with the boundary, thereby producing greater low-level con-  
 491 vergence (hence moisture convergence; not shown), larger MLCAPE, and lower MLLCL  
 492 near WPG than in the NOMP simulation (Figs. 10e, h, f, i). Overall, the observed rem-  
 493 nant outflow boundary likely enhanced the immediate storm environment near Elie and  
 494 the simulation appeared to capture this effect.

## 495 5.2 Triggering mechanism

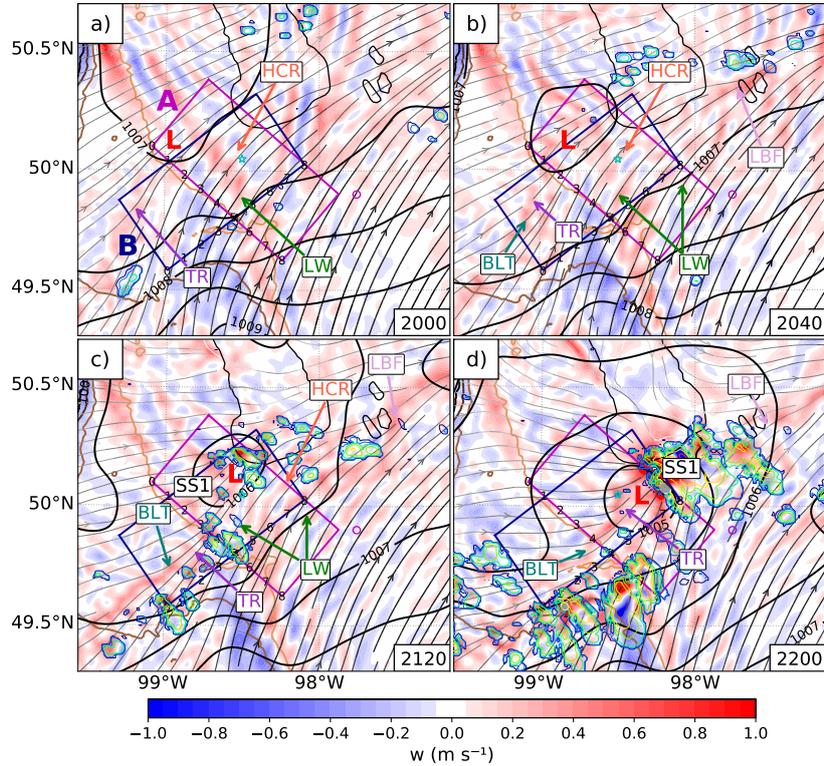
496 In this section, we investigate the potential triggering mechanism(s) of the Elie su-  
 497 percell using the simulated low-level flow in the afternoon of 22 June 2023 leading up  
 498 to the simulated convection initiation time. Two averaged cross sections (along their short  
 499 axes) each composed of 40 individual transects spaced  $\sim 1$  km apart were created to  
 500 diagnose the vertical structure of the simulated low-level flow (see Fig. 12 for locations).



**Figure 11.** Top panels: averaged 850-700 hPa simulated absolute vorticity (filled), geopotential height (black solid; in decameters), wind barbs, and 850-700 hPa differential absolute vorticity advection (purple solid for cyclonic, green for anticyclonic; in  $10^{-9} \text{ s}^{-2}$ ) at a) 1600 UTC and b) 2000 UTC 22 June 2007. Bottom panels: simulated SLP (filled), averaged 850-700 hPa vertical velocity (red for ascent, blue for descent; in  $\text{cm s}^{-1}$ ), and 10-m streamlines at c) 1600 UTC and d) 2000 UTC 22 June 2007. The synoptic low pressure system is labeled with a big ‘L’, while the lee low is marked with a small ‘L’.

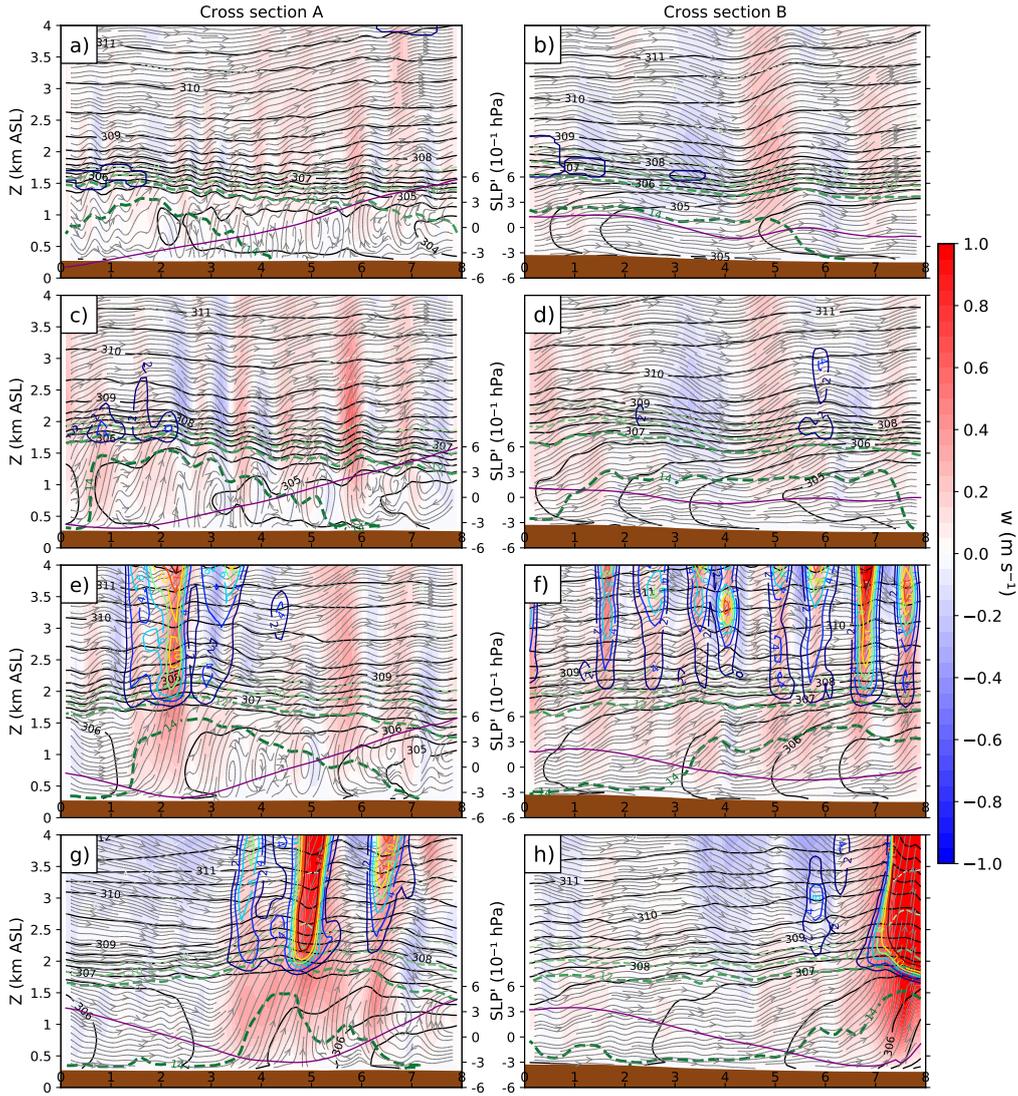
501 At 1600 UTC, a mesoscale low developed downwind (to the east) of Duck Moun-  
 502 tain (Fig. 1). We suspect that this low may have been a lee low (Palmén & Newton, 1969;  
 503 McGinley, 1982; Steenburgh & Mass, 1994; Holton, 2004) that formed within the syn-  
 504 optoc low as low-level cyclonic flow over eastern Saskatchewan moved east and descended  
 505 into the lower-laying RRV (Figs. 2, 11a). As in ERA5, 850-700 hPa cyclonic differential  
 506 vorticity advection and the associated mid-level ascent occurred above the mesoscale low

507 in the late afternoon (Figs. 3d, 11b), promoting further deepening of the low as it prop-  
 508 agated southeastward along the western slope of RRV (Figs. 1, 11d).



**Figure 12.** Simulated vertical motion at 1 km AGL (filled), 1-km ASL reflectivity (rainbow contours), SLP (black contours), and 10-m streamlines. The simulated boundaries (TR=surface trough, HCR=horizontal convective rolls, BLT=boundary layer thermals, LBF=lake breeze front, LW=lee waves) and Elie supercell (SS1) are also labeled. The locations of the averaged cross sections (A and B) are indicated by the boxes. ‘L’ indicates the approximate location of the mesoscale low pressure center. The times shown are a) 2000 UTC, b) 2040 UTC, c) 2120 UTC, and d) 2200 UTC 22 June 2007.

509 A trough (TR in Fig. 12) developed to the southwest of the low near 2000 UTC,  
 510 with CBL thermals (BLTs) propagating northeastward along it. Beginning at 2040 UTC,  
 511 a few shallow cumuli formed where the TR-BLT forcing interacted above the moisture  
 512 pool between the TR and the remnant outflow (see section 5.1.2; Figs. 12b, c, and 13c-  
 513 f). One of the simulated cells underwent rapid growth after coinciding with a simulated  
 514 lee-wave crest and matured into the simulated Elie supercell (SS1 in Fig. 12). Lee waves  
 515 were not observed over southern Manitoba around storm initiation time, possibly because  
 516 they were weak, resided above the lowest radar scan, and peaked above the CBL top where  
 517 backscatters were scarce (1.5-2 km vs. 1 km ASL; crossref Figs. 13, 5). HCRs and the  
 518 Lake Manitoba lake breeze front (LBF) were also simulated around this time (Figs. 12  
 519 and 13). Based on the simulation results, lee waves likely invigorated the convection in-  
 520 stead of triggered it on this day, similarly for the HCRs. The Lake Manitoba LBFs and  
 521 the cold front did not appear to contribute directly to the Elie storm initiation, as both  
 522 the observation and simulation suggest that the storm’s initial updrafts originated else-  
 523 where. After initiation, SS1 propagated eastward towards Elie (Fig. 12d)



**Figure 13.** Averaged simulated cross sections (see Fig. 12 for their locations) of vertical motion (filled), cloud and ice water mixing ratio (rainbow contours;  $10^{-2} \text{ g kg}^{-1}$ ), potential temperature (black contours), water vapor mixing ratio (dashed contours;  $\text{g kg}^{-1}$ ), and streamlines. The SLP perturbation relative to the cross-section mean is also plotted as purple lines. The times shown are a-b) 2000 UTC, c-d) 2040 UTC, e-f) 2120 UTC, and g-h) 2200 UTC 22 June 2007.

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## 6 Conclusions

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To identify and characterize the synoptic scale and mesoscale features that contributed to the formation of the 22 June 2007 Elie, Manitoba F5/EF5 tornado (the first and only F5/EF5 tornado in Canada to this date), a comprehensive meteorological analysis was undertaken using available observations and a high-resolution, cloud-resolving, real-case WRF simulation of this event. This study complements Hobson (2011) observational analysis of this event by better characterizing the storm environment immediately before the storm initiation and the storm-triggering mechanism(s). The main results are summarized below:

- 533 1. The Elie supercell formed under a quiescent upper-level regime with little upper-  
 534 level forcing for ascent, suggesting the importance of lower-troposphere large-scale  
 535 and mesoscale forcing in priming the convective environment.
- 536 2. The convective environment near Elie immediately before storm initiation was fa-  
 537 vorable for the development of tornadic supercells, highlighted by the very large  
 538 conditional instability (convective available potential energy  $> 4000 \text{ J kg}^{-1}$ ) and  
 539 moderate deep-layer shear (effective wind shear  $\sim 40 \text{ kt}$ ; effective storm-relative  
 540 helicity  $> 200 \text{ m}^2 \text{ s}^{-2}$ ). However, an important result is that both ingredients were  
 541 enhanced by a remnant cold pool left behind by early afternoon convection and  
 542 the storm-triggering boundary itself.
- 543 3. Despite the local enhancement, the shear parameters were not particularly impres-  
 544 sive compared to those typically found in the significantly tornadic supercell cases  
 545 in the United States considering all seasons, but rather typical compared to just  
 546 the summertime cases (R. L. Thompson et al., 2012).
- 547 4. The primary triggering mechanism of the Elie supercell was the interaction be-  
 548 tween boundary-layer thermals and a surface trough associated with a mesoscale  
 549 low pressure system that developed over the western slope of the Red River Val-  
 550 ley. The western Manitoba terrain (lee troughing effect) and 850-700 hPa cyclonic  
 551 differential vorticity advection may both have played roles in the formation and/or  
 552 intensification of the low.
- 553 5. Other mesoscale features such as lee waves and horizontal convective rolls may have  
 554 provided additional lifting that invigorated convection, including the Elie super-  
 555 cell.

556 Terrain lee effects have been observed and studied near many major mountains around  
 557 the world, including the Alps (e.g., Speranza, 1975; Buzzi & Tibaldi, 1978; McGinley,  
 558 1982), the Canadian and U.S. Rockies (e.g., Chung et al., 1976; Steenburgh & Mass, 1994),  
 559 the Tibetan Plateau in China (e.g., Chung et al., 1976; Q. Li et al., 2016), and the An-  
 560 des (e.g., Chung et al., 1977; K. L. Rasmussen & Houze Jr., 2016). They also have been  
 561 documented to the lee of shallower mountains such as the Appalachians (e.g., D. B. Thomp-  
 562 son, 2012). To the authors' knowledge, orographic lee phenomena induced by the even  
 563 shallower Manitoba terrain have not received any attention. Contrary to the common  
 564 notion that Manitoba terrain have minimal influences on the regional convection pat-  
 565 tern (e.g., Erfani, 1999), this study suggests that terrain-induced features may indeed  
 566 have effects in Manitoba. More numerical experiments by varying the regional topog-  
 567 raphy and land cover are required to better understand the topography's impacts on con-  
 568 vection initiation during the Elie event. Like the climatology study by Kovacs and Kir-  
 569 shbaum (2016) near Montréal, Québec, more meteorological observations and numeri-  
 570 cal experiments may be needed to diagnose the effects of Manitoba orography on the re-  
 571 gional convection pattern.

## 572 Open Research Section

### 573 Data availability statement

574 The complete ERA5 data (Hersbach et al., 2017) is publicly available at the Coper-  
 575 nicus Climate Change Service (C3S) Data Store (CDS) at [https://cds.climate.copernicus](https://cds.climate.copernicus.eu/cdsapp#!/dataset/reanalysis-era5-complete?tab=overview)  
 576 [.eu/cdsapp#!/dataset/reanalysis-era5-complete?tab=overview](https://cds.climate.copernicus.eu/cdsapp#!/dataset/reanalysis-era5-complete?tab=overview). The radar and  
 577 sounding observations are publicly available upon request from ECCO. The surface ob-  
 578 servations are publicly available for download from the ECCO Historical Climate Archive  
 579 at [https://climate.weather.gc.ca/historical\\_data/search\\_historic\\_data\\_e.html](https://climate.weather.gc.ca/historical_data/search_historic_data_e.html).  
 580 The GFS analysis used for WRF initialization can be download from the National Cen-  
 581 ter for Environmental Information (NCEI) data server at [https://www.ncei.noaa.gov/](https://www.ncei.noaa.gov/products/weather-climate-models/global-forecast)  
 582 [products/weather-climate-models/global-forecast](https://www.ncei.noaa.gov/products/weather-climate-models/global-forecast). The GHRSSST dataset can be  
 583 viewed on National Aeronautics and Space Administration's Worldview data viewer at

584 <https://worldview.earthdata.nasa.gov/> by searching for the GHRSSST dataset in  
 585 the data layer tab (under sea surface temperature) and select the correct date. The WRF  
 586 output and processed data used in this paper can be accessed at [https://doi.org/10](https://doi.org/10.5281/zenodo.10125834)  
 587 [.5281/zenodo.10125834](https://doi.org/10.5281/zenodo.10125834).

### 588 Software availability statement

589 The WRF model and its preprocessing system (WRF Preprocessing System; WPS)  
 590 used in this study can be downloaded from [https://github.com/wrf-model/WRF/releases](https://github.com/wrf-model/WRF/releases?page=2)  
 591 [?page=2](https://github.com/wrf-model/WRF/releases?page=2) and <https://github.com/wrf-model/WPS/releases>. The calculations and fig-  
 592 ures in this paper were performed/produced using Python v3.8.5 (Python Software Foun-  
 593 dation, 2020). The WRF output was analyzed using WRF-Python v1.3.2 (Ladwig, 2017),  
 594 licensed under the Apache License v2.0. Many of the thermodynamic and kinematic cal-  
 595 culations (excluding the convective parameters) and Skew-T plots were made using MetPy  
 596 v0.12 (May et al., 2020), licensed under the BSD 3 Clause license. The radar observa-  
 597 tions were analyzed using ARM-PyART v1.11.3 (Helmus & Collis, 2016), licensed un-  
 598 der the BSD 3 Clause license. The convective parameters were computed using SHARPy  
 599 v1.4.0 (Blumberg et al., 2017), licensed by the SHARPy license at [https://sharppy](https://sharppy.github.io/SHARPy/license.html)  
 600 [.github.io/SHARPy/license.html](https://sharppy.github.io/SHARPy/license.html). All figures were made using Matplotlib v3.2.2 (Hunter,  
 601 2007; Caswell et al., 2020), available under the Matplotlib license at [https://matplotlib](https://matplotlib.org/)  
 602 [.org/](https://matplotlib.org/). Map plots were made using Basemap v1.2.2 (Whitaker, 2020), available under  
 603 the Basemap license at <https://github.com/matplotlib/basemap>.

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