A Convection Parameterization for Low-CAPE Environments

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ABSTRACT: Numerical models that are unable to resolve moist convection in the atmosphere employ physical parameterizations to represent the effects of the associated processes on the resolved-scale state. Most of these schemes are designed to represent the dominant class of cumulus convection that is driven by latent heat release in a conditionally unstable profile with a surplus of convective available potential energy (CAPE). However, an important subset of events occurs in low-CAPE environments in which potential and symmetric instabilities can sustain moist convective motions. Convection schemes that are dependent on the presence of CAPE are unable to depict accurately the effects of cumulus convection in these cases. A mass-flux parameterization is developed to represent such events, with triggering and closure components that are specifically designed to depict subgrid-scale convection in low-CAPE profiles. Case studies show that the scheme eliminates the “bull’s-eyes” in precipitation guidance that develop in the absence of parameterized convection, and that it can represent the initiation of elevated convection that organizes squall-line structure. The introduction of the parameterization leads to significant improvements in the quality of quantitative precipitation forecasts, including a large reduction in the frequency of spurious heavy-precipitation events predicted by the model. An evaluation of surface and upper-air guidance shows that the scheme systematically improves the model solution in the warm season, a result that suggests that the parameterization is capable of accurately representing the effects of moist convection in a range of low-CAPE environments.

KEYWORDS: Convective storms; Cumulus clouds; Convective parameterization; Model evaluation/performance; Numerical weather prediction/forecasting; Parameterization

1. Introduction

The representation of moist convection in atmospheric numerical models is an important source of uncertainty and potential error (Molinari and Dudek 1992; Holloway et al. 2014). Although the march toward finer grid spacings that are capable of resolving convective processes proceeds apace, many large-scale regional models, global NWP systems, and climate models will be forced to parameterize convection for the foreseeable future (Arakawa et al. 2016). Many of these systems employ schemes that relate the deep convective activity directly to the convective available potential energy (CAPE) of the near-storm environment (Yano et al. 2013). In contrast, the topic of representing convection in environments with limited conditional instability (low CAPE) has received relatively little attention. However, accurately depicting such events is shown here to have the potential to impact the quality of the model solution significantly.

The potential for vigorous moist convection in low-instability environments represents a predictability and detection challenge because traditional conceptual models and severe weather

indices fail to identify such regions of possible convective activity (McCann 1978; Vescio and Thompson 2001; Davis and Parker 2014). Adopting the ingredients-based view of convection (Doswell 1987; Doswell et al. 1996), all developing cells require ascent, access to moisture, and a source of instability. In the low-CAPE context (including the zero-CAPE limit), the first two of these are typically associated with the warm sectors and frontal boundaries of midlatitude cyclones (Jewett and Wilhelmson 2006; Clark 2009; Evans 2010). The importance of strong synoptic and mesoscale forcing for ascent in the pre-convective environment is emphasized by van Den Broeke et al. (2005) in relation to the quasi-linear mesoscale convective systems that can develop parallel to cold fronts under low-CAPE conditions. Along the leading edge of these fronts, the poleward transport of moisture in the warm conveyor belt (Carlson 1980; Eckhardt et al. 2004) contributes to satisfying the moisture requirement of the ingredients-based perspective (King et al. 2017). These associations are reinforced by recent studies of high-shear, low-CAPE (HSLC) convective storms (Sherburn and Parker 2014) and the environments in which they form (Sherburn et al. 2016).

Identifying the source of instability, the final necessary ingredient for convection, can be challenging in the low-CAPE environments considered here. However, Schultz and Schumacher (1999) provide a framework for understanding how potential and symmetric instabilities can be tapped by convection under low-CAPE conditions. Convective case studies by Carr and Millard (1985) and Clark (2009) highlight

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the possible importance of an elevated dry intrusion, which can reduce the equivalent potential temperature (θ_e) aloft to create a potentially unstable layer. Such a feature is also identified by [Evans \(2010\)](#) to be related primarily to warm-season low-CAPE events in eastern North America, while [Sherburn and Parker \(2014\)](#) conclude that most HSLC-related severe weather reports occur in environments characterized by at least moderate potential instability.

The importance of potential instability to moist convection was first recognized by [Rossby \(1932\)](#) in his consideration of layers in which $\partial\theta_e/\partial p > 0 \text{ K Pa}^{-1}$ (his definition of “convective instability” is identical to the “potential instability” terminology adopted here). [Schultz et al. \(2000\)](#) provide a detailed review of early studies on the subject, along with a description of the process that converts this near-instability to one that leads to the buoyant acceleration of ascending parcels ([Sherwood 2000](#)). The key to realizing potential instability is sustained lifting that leads to condensation at the base of the potentially unstable layer. Continued ascent steepens the lapse rate within the layer, a change that leads to the production of CAPE ([Schultz et al. 2000](#)). However, this layer is likely to be neither horizontally homogeneous nor lifted uniformly to saturation. As a result, condensation occurs at the convective cloud scale before forced ascent can convert potential instability into CAPE over a large area ([Sherwood 2000](#)). [Kreitzberg and Perkey \(1976\)](#) use a Lagrangian updraft (plume) model to describe how this “releasable instability” fuels quasi-equilibrium convection in which CAPE values remain low throughout the convective life cycle.

The dynamically forced vertical layer displacements required to convert potential instability into CAPE are typically largest in the free troposphere, leading to the development of elevated convection as the potential instability is released ([Rosenow et al. 2018](#)). [Corfidi et al. \(2008\)](#) define elevated convection as any convection whose inflow source lies above the boundary layer, a characteristic that distinguishes this class of event from its more common counterpart that relies on surface heating for initiation. The close relationship between low-CAPE and elevated convection in environments of potential instability was initially described by [Colman \(1990a,b\)](#), and more recently investigated by [Sherburn and Parker \(2014\)](#). [Moore et al. \(2003\)](#) provide a process-level description of this connection, concluding that potential instabilities are likely to be most relevant for elevated convection of the “frontal overrunning” type. This archetype bears a strong resemblance to the “escalator–elevator” concept illustrated by [Neiman et al. \(1993\)](#), in which a sloping warm conveyor belt ascent is punctuated by upright convection as localized pockets of potential instability are released. These conceptual models provide further evidence of the strong links between potential instability and elevated convection in low-CAPE environments.

The representation of deep convection under low-CAPE conditions presents a challenge to atmospheric models that are run at resolutions that are too coarse to depict explicitly the convective ingredients involved ([Randall et al. 1997](#); [Anderson et al. 2002](#)). The problem is particularly acute for models that employ convection parameterization schemes that are sensitive

to the conditional instability represented by CAPE (for example, [Fritsch and Chappell 1980](#); [Betts 1986](#); [Kain and Fritsch 1990](#); [Zhang and McFarlane 1995](#); [Bechtold et al. 2001](#)). In environments with little to no CAPE, such schemes either will fail to trigger or will depict exceedingly weak convection. A numerical model that fails to parameterize the effects of subgrid-scale convection will suffer from systematic errors in the form of delayed convective initiation ([Weisman et al. 1997](#)), insufficient triggering ([Petch et al. 2002](#)), and excessive vertical mass fluxes and precipitation rates generated by the gridscale condensation scheme ([Kato and Saito 1995](#); [Weisman et al. 1997](#); [Roberts and Lean 2008](#)).

In low-CAPE environments with sufficient potential instability and moisture, sustained lifting may lead to convective initiation. In a model with an inactive CAPE-dependent deep convection scheme, no parameterization represents this subgrid-scale process. Instead, the gridscale condensation scheme depicts latent heating at the base of the potentially unstable layer as the grid cell-averaged relative humidity nears saturation in the ascending airstream ([Teixeira 2001](#); [Quass 2012](#)). The associated buoyancy increase fuels further ascent, leading to continued condensation as any CAPE generated by the realized potential instability is consumed via the “gridpoint storm” mechanism ([Scinocca and McFarlane 2004](#)). This unphysical process results in precipitation “bull’s-eyes” as the gridscale condensation scheme couples with the dynamical core to develop moist convective motions at the minimum length scale of the model. This problem is a direct result of the model’s inability to simulate accurately the release of low-CAPE instabilities (potential instability in this example) via subgrid-scale processes.

Models that employ a class of convective schemes whose activity depend on moisture convergence ([Bougeault 1985](#); [Tiedtke 1989](#); [Banacos and Schultz 2005](#)) have the potential to represent better the effects of convection in low-CAPE environments. This assertion is supported by results of [Suhas and Zhang \(2015\)](#), who conclude that closures based on this quantity yields the highest correlation with observed convection; however, they note that all relationships to the large-scale state weaken as model resolution increases. Schemes that are particularly well suited to represent the release of potential instability are those that can act in low-CAPE environments in response to elevated synoptic and mesoscale forcings for ascent. Such “midlevel” convection schemes are employed in operational systems run at both ECMWF ([Tiedtke 1989](#)) and UKMO ([Gregory and Rowntree 1990](#)); however, their impact on predictions of convection and overall guidance quality have not been documented extensively in the literature.

The purpose of this study is to describe the implementation of a convection scheme designed to represent the effects of convection in low-CAPE environments. The scheme’s development was motivated by concerns expressed by forecasters about the prevalence of precipitation bull’s-eyes in guidance generated by the Canadian Regional Deterministic Prediction System ([Caron et al. 2015](#)). This error mode is consistent with insufficient triggering of the CAPE-dependent [Kain and Fritsch \(1993\)](#)-based scheme that is used to parameterize deep convection in the model. The trigger and closure elements of

the low-CAPE scheme draw heavily from existing midlevel convection parameterizations (Tiedtke 1989; Gregory and Rowntree 1990), consistent with the relationship between elevated and low-CAPE convection. However, in keeping with a process-oriented viewpoint and the ingredients-based framework, the new scheme will be referred to hereafter as a “low-CAPE” convection parameterization, with the “mid-level” terminology adopted only in specific reference to these predecessor formulations.

The investigation begins with an overview of the datasets and numerical model employed in this research in section 2, followed by a description of the parameterization itself in section 3. The impact of representing low-CAPE convection on summer precipitation events is evaluated in section 4, whereas the overall influence of the parameterization on guidance quality is assessed in section 5. The study concludes with a discussion of the broader implications of these results in section 6.

2. Data and model description

This study employs data from several observing networks for case study descriptions (section 4) and model evaluations (section 5). In situ data consist primarily of radiosonde and surface observations available over the Global Telecommunication System. Included in the surface observational dataset are precipitation accumulations measured at synoptic weather stations that have passed the quality control stage of the Canadian Precipitation Analysis (Lespinas et al. 2015) to ensure a baseline of validity.

Spatial precipitation structures are evaluated using the quantitative precipitation estimate from the Multi-Radar Multi-Sensor (MRMS) system, derived from gauge-corrected radar retrievals (Zhang et al. 2016). The atmospheric state is assessed using primarily the ERA-Interim (Dee et al. 2011), although the higher-resolution ERA5 (Hersbach et al. 2018) is employed for the analysis of mesoscale structures (section 4a).

The Global Environmental Multiscale (GEM) model is used in all operational applications at the Canadian Meteorological Centre. The implicit, semi-Lagrangian dynamical core solves the full governing equations on a latitude–longitude grid (Côté et al. 1998; Husain and Girard 2017). The terrain following vertical coordinate is of the log-hydrostatic-pressure type (Girard et al. 2014).

This study employs the Regional Deterministic Prediction System, in which the GEM model is configured to run hydrostatically with 0.09° (~10 km) grid spacing and a 300-s time step. In the vertical 84 levels are distributed between the surface and the 0.1-hPa model lid, with 12 levels below 850 hPa in a standard atmosphere and the lowest thermodynamic level at ~10 m above ground.

The physical parameterizations adopted by the Regional Deterministic Prediction System recently underwent a major upgrade (McTaggart-Cowan et al. 2019a). Of direct relevance here are the moist atmospheric processes that are handled primarily by either the gridscale condensation scheme (Sundqvist 1978), or by one of the model’s convection parameterizations. The latter include the shallow convection scheme proposed by

Bechtold et al. (2001), a modified version of the Kain and Fritsch (1993) deep convection scheme (McTaggart-Cowan et al. 2019b), and the parameterization for low-CAPE convection described here. In addition to the direct impact of these schemes on state variables, each of these parameterizations interacts with the radiation scheme through the cloud-radiation interface described by McTaggart-Cowan et al. (2019a).

3. A low-CAPE convection scheme

The formulation of the low-CAPE convection scheme described here is based loosely on the ECMWF (2018) description of the midlevel variant of the Tiedtke (1989) scheme. An overview of the convective plume-based mass-flux scheme is provided in Fig. 1 for reference. The utility of such a Lagrangian updraft model in low-CAPE environments was first demonstrated by Kreitzberg and Perkey (1976), who investigated the parameterization of moist convection in the presence of potential instability. To avoid the double-counting of moist convective effects, the new scheme is only tested for possible activity where CAPE values are below 250 J kg^{−1}. This limit that is more restrictive than the 500 J kg^{−1} typically used to identify HSLC events (Sherburn and Parker 2014), a reflection of the precedence given to the deep convection scheme in environments with moderate conditional instability (ECMWF 2018; McTaggart-Cowan et al. 2019a). Unlike HSLC severe storm studies; however, no wind shear condition is applied because the intent of the scheme is to represent the full range of low-CAPE convective intensities.

a. Trigger function

The trigger function for the low-CAPE scheme is relatively simple in comparison with the highly parameterized triggers used for deep convection (Suhas and Zhang 2014). Consistent with the conceptual model of “frontal overrunning” (Weckwerth et al. 2019), resolved vertical motions form the basis for convective initiation. These vertical wind speeds (w) can become large as gravitational or symmetric instabilities are released, particularly when the gridscale condensation scheme predicts latent heating in ascending airstreams.

The first stage of the trigger is therefore a requirement on the minimum resolved mass flux (M_e), wherein,

$$M_e \equiv \rho w \Delta x^2 > M_e^{\min}, \quad (1)$$

where ρ is air density and Δx is the model grid length. Evaluation of a spatially uniform $M_e^{\min}$ threshold on the global domain suggested that the scheme was excessively active in tropical regions, an error corrected by the adoption of a heuristic scaling by the Coriolis parameter (f):

$$M_e^{\min} = \frac{m_o}{|f|}, \quad (2)$$

in which $m_o = 7 \times 10^6 \text{ kg s}^{-2}$ and $f \geq 1 \times 10^{-6} \text{ s}^{-1}$. The justification for this scaling is based on the presence of off-equatorial ascent driven by inertial instability that is not associated with low-CAPE convection (Tomas and Webster 1997). Although effective, further efforts should be made to

FIG. 1. Schematic of the major components of the low-CAPE convection scheme. Yellow cylinders represent exchanges of information with other model components (specific quantities are itemized with orange boxes and dashed lines at the bottom of the plot), blue boxes represent major elements of the parameterization, and red parallelograms represent logical branches or conditions. Symbol definitions and values for the Regional Deterministic Prediction System are defined in Table 1.

develop a trigger modulation that is more directly linked to convective processes.

Air at the first level at which the Eq. (1) inequality is satisfied is lifted to its lifting condensation level to seed the updraft (section 3b). Unlike the equivalent procedure in standard deep

TABLE 1. Definition of symbols used in Fig. 1, with values defined for the Regional Deterministic Prediction System for applicable parameters.

Symbol	Description	Parameter value
H_c	Cloud depth	
$H_{\min}$	Minimum cloud depth	2000 m
LCL	Lifting condensation level	
LFS	Level of free sink	
m_o	Factor for $M_e^{\min}$	$7 \times 10^6 \text{ kg s}^{-2}$
M_e	Environmental mass flux	
$M_e^{\min}$	Minimum environmental mass flux	Follows Eq. (2)
z	Height above ground	
$z_{\max}$	Maximum mixing-layer base height	10 000 m
$z_{\min}$	Minimum mixing-layer base height	500 m

convection schemes, multiple layers are not mixed to create updraft parcel properties because there is no requirement for turbulent inflow as layers are dynamically lifted to initiate low-CAPE convection.

b. Updraft

The cloud model used to compute the properties of a prototypical updraft is initialized with the cloud-base mass flux:

$$M_{\text{LCL}}^u = \rho_{\text{LCL}} \pi r^2 w_{\text{LCL}}, \quad (3)$$

where the subscript “LCL” represents values at the lifting condensation level (here equivalent to cloud base), $r = 500 \text{ m}$ is the prescribed radius of the updraft (Bechtold et al. 2001), and M^u is the updraft mass flux (a superscript u denotes updraft properties hereafter). The resolved vertical motion is used here because it represents the mean vertical velocity within the grid cell, characterizing a distribution that is expected to be narrow in coherent airstreams undergoing forced ascent. The implications of this assumption for the closure are discussed in section 3d.

The construction of the updraft by the plume model proceeds following the strategy described by Kain and Fritsch (1990). Rudimentary microphysical assumptions are used to generate hydrometeor profiles, with phase transitions affecting updraft acceleration through changes to the parcel buoyancy as described by Simpson and Wiggert (1969):

$$\Delta(w^u)^2 = \frac{2g\Delta z}{1+\lambda} \left(\frac{\overline{T_v^u} - \overline{T_v}}{\overline{T_v}} - D_p \right) - 2 \frac{\varepsilon^u}{M^u} (w^u)^2. \quad (4)$$

In this equation, g is gravitational acceleration, T_v is the virtual temperature, ε is the entrainment rate (in mass flux units of kg s^{-1}) and $\lambda = 0.5$ is an empirically derived constant. The Δ operator signifies a difference between two adjacent model levels, while the overbar ($\overline{}$) indicates an average over the same layer. The D_p term in Eq. (4) is a modification to the traditional form that represents the drag on updraft acceleration resulting from condensate loading (Ogura and Cho 1973). Parcel buoyancy [the first term on the rhs of Eq. (4)] is generally small in low-CAPE profiles, such that the cloud-base updraft speed (w_{LCL}) that serves as the lower boundary condition plays an important role in determining the depth of the cloud layer.

FIG. 2. Synoptic environment at 0000 UTC 6 Jul 2016 as depicted in the ERA-Interim. (a) The upper-tropospheric state is shown at 250 hPa with heights (black contours; 12-dam intervals), winds (wind barbs, with short, long, and pennant barbs representing 5, 10, and 25 m s^{-1} wind speeds, respectively; barbs are only plotted for wind speeds exceeding 10 m s^{-1}) and wind speed (shaded; in m s^{-1} , as indicated on the color bar). (b) The lower-tropospheric state is shown with sea level pressure (black contours; 4-hPa intervals) and fronts from the operational surface analysis produced by the Weather Prediction Center, as well as surface cyclone (“L”) and anticyclone (“H”) positions from the same source. In both panels, the location of the Pickle Lake sounding station is shown with a cyan crosshair symbol, the Aberdeen station is shown with a yellow crosshair symbol, and the domains of interest for the case studies are outlined: the northern Ontario domain in red in (a), and the midwestern domain in dark magenta in (b). Also shown in (b) is the first-echo location of the convective features highlighted in the case studies (green thunderstorm symbols), both of which appear prior to the valid time of this analysis.

The updraft mass flux (M^u) of the cloud prototype (also called the unit updraft) needs to be specified in order to determine the updraft velocity from Eq. (4). Beginning with the cloud-base mass flux [Eq. (3)], the M^u profile is computed as

$$\Delta M^u = \overline{\varepsilon^u} - \overline{\delta^u}, \quad (5)$$

which depends directly on the representation of entrainment and detrainment (δ) processes (Bechtold et al. 2001). Lateral mixing (ε and δ) is represented in the low-CAPE scheme using the buoyancy sorting technique introduced by Kain and Fritsch (1990), with inflow into the mixing zone parameterized as $\gamma M_{\text{LCL}}^u/r$ [$\gamma = 0.2$ m as in Kain and Fritsch (1990) following Simpson (1983)]. A zone of forced detrainment lies above the level of neutral buoyancy ($T_v^u = \overline{T}_v$), followed by the overshooting cloud top where $w^u = 0 \text{ m s}^{-1}$. Over this layer, the updraft horizontal momentum is adjusted to account for the pressure gradient across the ascending plume (Gregory 1996), as described for the model's deep convection scheme by McTaggart-Cowan et al. (2019b). A minimum cloud depth of 2000 m is required for this form of precipitating convection (Niziol et al. 1995; Kain 2004), with shallower prototypes failing this component of the trigger requirement for precipitating convection and returning for further evaluation (section 3a) until the potential triggering level exceeds 10 km above ground (ECMWF 2018).

c. Downdraft

Downdrafts resulting from elevated, low-CAPE convection typically occur in environments with both significant downdraft convective available potential energy and downdraft convective inhibition (Market et al. 2017). The latter limits the downdraft's ability to reach the surface, such that downdraft detrainment may occur within the stable layer that lies beneath the updraft source level. This behavior distinguishes such convection from its boundary layer-based counterpart, for which downdraft detrainment typically occurs over a relatively shallow near-surface layer.

Although such differences in downdraft characteristics will have an impact on cold pool formation and potential secondary initiation (Kain and Fritsch 1998; Marion and Trapp 2019), there is no distinction from the perspective of the underlying physical processes. As such, a downdraft model that is similar to the one described by Kain and Fritsch (1993) is employed in the low-CAPE convection scheme. The top of the downdraft [the level of free sink (LFS)] coincides with the level possessing the minimum environmental saturation equivalent potential temperature in the cloud layer. Detrained moisture is evaporated into environmental air at this level to establish θ_e^d (superscript d denotes downdraft quantities hereafter), the quantity that is conserved in the downdraft.

The initial downdraft mass flux estimate is based on an extraction of updraft air scaled by precipitation efficiency (E_p) as

$$M_{\text{LFS}}^d = (1 - E_p) \rho_{\text{LFS}} \pi r^2. \quad (6)$$

The downdraft penetrates vertically to the level at which it becomes neutrally buoyant, below which detrainment occurs

over a 60-hPa layer. Between the LFS and the top of the detrainment layer, the downdraft grows through a constant fractional entrainment rate of 0.03.

The full profiles of M^d , ε^d , and δ^d are rescaled based on the amount of condensate available for evaporation in the downdraft. The latter is related to $(1 - E_p)P$, where P represents the precipitation flux. In the Regional Deterministic Prediction System, E_p is set to 0.4 as an intermediate estimate of the efficiencies of convective clouds with bases more than 1.5 km above the ground (Fig. 12b of Fankhauser 1988). Alternative formulations for E_p that account for the subcloud environmental state either directly (Bluestein and Parks 1983) or indirectly via gridscale evaporation rates, failed to improve guidance skill beyond that which was achieved with the constant efficiency estimate.

d. Closure

The closure of a convection scheme relates M^u at a specified level to gridscale quantities. The closure is very direct in the low-CAPE convection parameterization:

$$M_{\text{LCL}}^u = M_{e\text{LCL}}, \quad (7)$$

where M_e is defined in Eq. (1). Physically, this means that all air entering the grid volume at the LCL is processed by moist convection. This closure is consistent with the philosophy adopted for the trigger function (section 3a) in its use of gridscale vertical motion as a key ingredient for low-CAPE convection.

The prescription of M_{LCL}^u renders the closure diagnostic, thus eliminating the need for the iterations that are standard with CAPE-based closures. Prototype cloud properties are simply scaled by the ratio of Eqs. (3) and (7), a factor of $\Delta x^2/(\pi r^2)$ that represents the cloud population size required to draw the full resolved mass flux into convective updrafts. This relationship suggests that w_{LCL} in the trigger [Eq. (3)] is an underestimate of the true updraft velocity because of the expectation that subgrid-scale updrafts occupy a small fraction of the domain. Although the use of empirical estimates of this quantity (w_{LCL}) are common in updraft models (typically constants Bechtold et al. 2001; Chikira and Sugiyama 2010), the direct link between the trigger and closure in this scheme suggests that further investigation of the appropriate value for w_{LCL} based on subgrid-scale variability is needed (Baba 2019). Despite this simplification, the closure [Eq. (7)] is successful in promoting convective activity where the release of instabilities not detected by the deep convection scheme leads to the development of strong vertical motions in the model.

e. Solver

The mass-flux framework is adopted for this parameterization, consistent with the treatment of shallow and deep convection in the model. The underlying equations for this scheme therefore take the following form:

$$\frac{\partial \Psi}{\partial t} = \frac{1}{\rho(\Delta x)^2} \left\{ \frac{\partial}{\partial z} [(M^u + M^d)\Psi] - (\varepsilon^u + \varepsilon^d)\Psi + \delta^u \Psi^u + \delta^d \Psi^d \right\}, \quad (8)$$

FIG. 3. Quantitative precipitation estimates (shaded; in mm h^{-1} , as shown on the color bar) from the MRMS dataset at hourly intervals from (a) 2100 UTC 5 Jul to (h) 0400 UTC 6 Jul 2016. The domain employed for subsequent Ontario case study maps is outlined with a black box in each panel, and the location of the Pickle Lake sounding station is shown with a cyan crosshair symbol in (d) for reference. Surface fronts are plotted as in Fig. 2 when frontal analyses are available.

FIG. 4. Sounding from Pickle Lake for 0000 UTC 6 Jul 2016. The temperature is shown in red and the dewpoint is shown in blue. Winds are plotted with barbs as in Fig. 2. The sounding location is shown in Figs. 2 and 3d.

which is valid for any gridscale variable Ψ (Bechtold et al. 2001). In the low-CAPE convection scheme the potential temperature, water vapor mixing ratio and wind components are treated directly by this equation to yield tendencies for gridscale temperature, moisture, and winds.

Equation (8) is solved implicitly within the parameterization. This numerical decision reduces the need for time-splitting and improves the convergence properties of the scheme with respect to the small time-step limit.

4. Heavy precipitation case study

An unusual flow pattern over North America provides a useful test bed for illustrating the behavior of the low-CAPE convection scheme described in the previous section. On 6 July 2016, low-CAPE convection occurred both in the form of isolated convective cells in northern Ontario and as part of a long-lived squall line in the midwestern United States. A focus on this date therefore serves to document the performance of the parameterization in the context of different observed convective modes.

Each case study begins with an analysis of observational data related to convective activity. These comparisons with observed storm evolution represent attempts to ensure that the

new scheme is acting in a physically reasonable manner, and that improvements in the model solution are not the result of error compensation within the system.

A jet streak centered over Quebec places both of the areas of interest in the confluent entrance region (Fig. 2a). The midwestern squall line develops in the equatorward jet entrance, an optimal location for quasigeostrophic forcing for midtropospheric ascent (Namias and Clapp 1949; Uccellini and Johnson 1979; Bluestein and Thomas 1984). The baroclinicity that supports the anomalous jet appears in the form of a pair of zonally oriented fronts that straddle the international border (Fig. 2b). Both convective events are initiated in close proximity to these fronts, consistent with conceptual models of convective development in low-instability environments (van Den Broeke et al. 2005; Sherburn et al. 2016).

The GEM simulations described in this section are based on integrations of the 0000 UTC 6 July 2016 Regional Deterministic Prediction System (section 2), run with the low-CAPE scheme either activated (LC) or turned off (CTL). The latter is considered the control configuration because the operational atmospheric physics suite did not contain a parameterization for low-CAPE convection until mid-2019 (McTaggart-Cowan et al. 2019a). Because an incremental analysis update

technique based on a 6-h assimilation window is employed in this system (Buehner et al. 2015), the model is initialized at 2100 UTC 5 July for all integrations. This strategy permits a “hot start” of the model, in which physical quantities (including cloud and turbulence fields) are recycled within a continuous integration framework. The associated reduction of model spinup permits analyses of short-range predictions, a behavior that is useful in the current context because the divergence of solutions between model integrations is minimized even for the relatively small-scale features considered here (Zhang and Wu 2003).

To ensure the robustness of the results described in this section, multiple simulations were carried out on this case with different domain configurations, all of which yield qualitatively similar results. This suggests strongly that differences between the integrations are physically robust and not the result of “chaos seeding” (Ancell et al. 2018).

a. Isolated convection in northern Ontario

The origins of scattered, intense convection in northern Ontario are assessed in this section (quantitative precipitation estimates and relevant locations are shown in Fig. 3). The release of elevated potential instability leads to precipitation “bull’s-eyes” in simulations run without the low-CAPE convection scheme. This case highlights the direct influence that the scheme can have on the reduction of small-scale precipitation extremes.

1) OBSERVED EVOLUTION

Isolated convective cells develop over northwestern Ontario at 2100 UTC 5 July (1700 local time), immediately poleward of the warm-frontal boundary in the region (Fig. 3a). The storms remain anchored to this feature at 0000 UTC (Fig. 3b), despite their rapid eastward progression under a strong westerly flow aloft (Fig. 2). By 0400 UTC (local midnight), the cells have largely dissipated, although scattered showers remain throughout the area.

The 0000 UTC 6 July Pickle Lake sounding is taken to be representative of conditions poleward of the surface front in the preconvective environment (Fig. 3d), wherein the radiosonde intersects the baroclinic zone near 850 hPa (Fig. 4). The sounding provides preliminary evidence that the three ingredients for convective initiation are present. The saturated layer above the frontal inversion contains ample moisture for convection, while the veering wind profile across this layer implies warm advection (“frontal overrunning”) and attendant quasigeostrophic forcing for ascent. The source of instability is less clear, as daytime heating has failed to warm the boundary layer sufficiently to overcome the convective inhibition of the 850-hPa inversion. The profile is nearly moist adiabatic above this level, devoid of appreciable CAPE. However, drying in the 750–550-hPa layer leads to a negative vertical gradient in equivalent potential temperature (θ_e) that is indicative of potential instability (Fig. 5).

The release of potential instability requires layer lifting via forced ascent that may occur at synoptic scales or at the mesoscale. The ERA5 is therefore used to assess convective ingredients in the near-storm environment over the 850–750-hPa

FIG. 5. (left) Profile of equivalent potential temperature θ_e labeled on the abscissa as “Eq. Pot. Temp.” and shown in the plot with a solid black line in the Pickle Lake sounding for 0000 UTC 6 Jul 2016 (Fig. 4). Potentially unstable layers, highlighted with a red background, are computed as positive values of the vertical gradient of θ_e ($\partial\theta_e/\partial p > 0 \text{ K Pa}^{-1}$) following the application of a digital filter to remove structures less than 40 hPa in depth. (right) The resulting gradient profile is shown with a dashed gray line at neutral potential stability for reference.

layer that is likely to represent the convective inflow (Fig. 6). A local maximum in relative humidity is indicative of available moisture (Fig. 6a); however, values remain generally below 80% in the region of interest and may serve as a limiting factor on the areal extent of convection. A similar maximum in frontogenesis (Fig. 6b) implies a forcing for mesoscale ascent in the secondary circulation across the baroclinic zone (Keyser et al. 1988). The final convective ingredient, instability, is assessed as $\partial\theta_e/\partial p$, which is positive in areas of potential instability (Fig. 6c). Combining these three ingredients with a set of thresholds determined by inspection of this case,¹ it becomes clear that the bulk of convective activity occurs in the area where the local extremes of the convective ingredients overlap

¹ The thresholds employed in Fig. 6 are not intended to be universally applicable. Instead, they are meant simply to highlight the region in which local extremes in the ingredients overlap to create favorable conditions for convective activity.

FIG. 6. Ingredients for convection at 0000 UTC 6 Jul 2016 over the northern Ontario region outlined in Fig. 3. All quantities represent averages over the 850–750-hPa layer as diagnosed from ERA5. (a) Moisture availability is shown using relative humidity (plotted as a percentage, with a heavy red line at 70%), (b) forcing for mesoscale ascent is shown using quasi-horizontal frontogenesis (Petterssen 1936) (plotted as $1 \times 10^{-10} \text{ K m}^{-1} \text{ s}^{-1}$ with a heavy green line at $0 \times 10^{-10} \text{ K m}^{-1} \text{ s}^{-1}$), and (c) instability is shown using $\partial\theta/\partial p$ (plotted as $1 \times 10^{-4} \text{ K kg m}^{-1} \text{ s}^{-2}$ with a heavy blue line at $5.5 \times 10^{-4} \text{ K kg m}^{-1} \text{ s}^{-2}$). (d) The thresholds for the ingredients are superposed, with light, medium, and dark gray shading representing one, two, and three threshold exceedences, respectively. Also shown in (d) is the MRMS precipitation rate estimate (mm h^{-1} , as shown on the color bar) and a 10-km scale for reference.

(Fig. 6d). This suggests that the observed convection in this low-CAPE region forms in response to frontogenetic forcing in a relatively moist, potentially unstable environment.

Before comparing the precipitation-rate estimates from the MRMS dataset with model predictions, the scale mismatch between the 1-km observations and the 10-km model grid must be addressed, particularly given the small-scale nature of the convective elements (Fig. 6d). Because the heavy-precipitation cores cover only a small fraction of the model grid, peak precipitation rates in the model should not match those of the observations. Arguably, any model-generated precipitation structure that is below about $7\Delta x$ is underresolved; however, for the purposes of addressing solely the representativeness issue, conservative remapping of the MRMS data to the model grid suffices (Fig. 7). The impact of this remapping on the precipitation maxima is large, with peak rainfall rates dropping to below 20 mm h^{-1} . The strong influence of grid size in this case is physically relevant because it indicates that the scales associated with the convective motions are much smaller than those that can be explicitly represented by the model.

2) SIMULATED EVOLUTION

The CTL model integration for this event successfully simulates the development of precipitation in northern Ontario in

the short-range forecast (Fig. 8a). However, bull's-eyes in which precipitation rates exceed 30 mm h^{-1} are apparent in the precipitation field. Although such rainfall intensities are consistent with the localized maxima estimated in the full-resolution MRMS dataset (Fig. 7a), they are clearly excessive on a 10-km grid (Fig. 7b). This broadening of the precipitation cores leads to area-averaged CTL rainfall rates that nearly double the observational estimate (Figs. 7b and 8a).

In a region of convective inhibition ($\text{CIN} = -25 \text{ J kg}^{-1}$) and small CAPE (90 J kg^{-1}), the CTL model is releasing the potential instability via the gridpoint storm mechanism. Despite being clearly convective in nature (Fig. 7), precipitation in the CTL integration results primarily from the gridscale scheme, with only trace contributions from the deep convection parameterization because of limited CAPE (Fig. 8b). The underresolved updrafts tap potential instability at the finest scales possible in the 10-km model, but these dimensions remain much larger than reality in this case. The result is unrealistically broad updrafts that generate excessive precipitation rates.

When the low-CAPE convection scheme is activated (the LC integration), the isolated precipitation maxima of the CTL integration are replaced by an extended area over which rainfall rates approach 20 mm h^{-1} , in line with the MRMS estimate (cf. Figs. 7b and 8c). Although the regionally averaged

FIG. 7. Precipitation rate (in mm h^{-1} , as shown on the color bar) at 0000 UTC 6 Jul 2016 over the northern Ontario region outlined in Fig. 2, as estimated by the MRMS dataset on (a) the original 1-km grid and (b) remapped onto the 10-km grid of the Regional Deterministic Prediction System. The rate shown in (b) is computed using the conservative regridding capacity of the Earth System Modeling Framework (Hill et al. 2004). The track of the precipitation maximum on the 10-km grid is shown with heavy black line in (b), with hourly markers (time is in UTC) color coded following the color bar. Also shown in each panel is a 10-km scale for reference. The plot-domain-averaged precipitation rate is shown in the top-left corner of each panel.

convective rainfall rate only increases by 0.1 mm h^{-1} , the ratio of convective to total precipitation rises from $<20\%$ to 50% in this integration (based on domain-average precipitation rates annotated in Fig. 8). The rainfall reduction and change in convective ratio are persistent differences between the CTL and LC integrations (Fig. 9) during both the primary convective period (2100 UTC 5 July–0200 UTC 6 July) and a secondary event that is overpredicted by the model (after 0200 UTC 6 July). The dramatic reduction in precipitation rates predicted by the gridscale condensation scheme is an indication that the potential instability is processed through parameterized moist turbulence in the low-CAPE convection scheme, rather than being extracted through an under-resolved interaction between gridscale condensation and model dynamics.

b. Squall line in the midwestern United States

The evolution of an intense squall line over the midwestern United States is outlined in this section (quantitative

precipitation estimates and relevant locations are shown in Fig. 10). The cold pool generated by initially elevated cells promotes the development of surface-based deep convection and plays an important role in establishing squall-line structure in simulations with the low-CAPE scheme activated. This case highlights the indirect impact that the scheme can have on convective organization.

1) OBSERVED EVOLUTION

The convective cells that represent the beginnings of the midwestern squall line form over western Minnesota at 1700 UTC 5 July 2016 (local noon; not shown). The 1800 UTC Aberdeen sounding (Fig. 11; 150 km to the west of the convective initiation) is not favorable for surface-based convection despite daytime surface heating because cold northerly flow undercuts the warmer air above the 850-hPa frontal inversion to create large CIN (-300 J kg^{-1}). Instead, the Aberdeen sounding represents an environment on the cold side of the surface front that is conducive to the development of elevated convection (Horgan et al. 2007).

Near 1800 UTC, the Morris Municipal Airport (KMOX) reports thunderstorms with cloud bases up to 2500 m above ground ($\sim 700 \text{ hPa}$), consistent with parcels that have undergone forced ascent from above the frontal inversion (Fig. 11 annotations). Parcels ascending from this level still need to overcome a 100-hPa layer of strong negative buoyancy before beginning free convective ascent in a “tall and thin” low-CAPE environment (Blanchard 1998). A decrease in equivalent potential temperature of 10 K between 850 and 650 hPa (Fig. 12) suggests that mesoscale ascent associated with the frontal circulation may have initiated elevated convection in this case (Trier and Parsons 1993), supported by synoptic-scale ascent in the equatorward jet entrance region (Fig. 2a). METAR reports indicate that convection remains elevated (2000 to $>3000 \text{ m}$ above ground) as the developing line passes through Minneapolis at 0000 UTC 6 July.

Over the next 6 h, the squall-line convection transitions from elevated to surface-based. Despite the development of a conceptual model by Rockwood and Maddox (1988), the processes involved with such a transition have not been studied extensively, particularly in the context of such linear features (Gray and Frame 2019). However, Marsham et al. (2011) show that evaporatively cooled downdrafts can play an important role in secondary convective initiation (Bennett et al. 2006). As the elevated squall line passes through La Crosse, Wisconsin (ceilings are $\sim 3000 \text{ m}$ above ground during the 0300 UTC passage), a robust cold pool is evident at the surface (Fig. 13). Screen-level temperatures drop by 6°C as the squall line passes, as winds turn to the north and increase to 18 m s^{-1} (gusting to 30 m s^{-1}). This evolution closely follows the conceptual squall-line model of Johnson and Hamilton (1988), and yields a density current that is capable of lifting near-surface air along the gust front to facilitate the initiation of surface-based convection (Wakimoto 1982). Observed cloud bases in the squall line drop to below 1 km as convection moves through Rockford, Illinois, shortly after 0600 UTC (Fig. 10), indicative of a completed transition from elevated to surface-based convection.

FIG. 8. Precipitation rate at 0000 UTC 6 Jul 2016 in the (a) CTL and (c) LC integrations, for comparison with the remapped MRMS estimate shown in Fig. 7b. (b),(d) The convective precipitation component is shown for the CTL and LC integrations, respectively. (In the LC integration, convective precipitation includes contributions from both the deep and low-CAPE convection schemes.) All fields are plotted in mm h^{-1} , as shown on the color bars at the bottom of the plot. The track of the precipitation maximum is shown with heavy black line in each panel, with hourly markers (time is in UTC) color-coded following the color bars. The 0000 UTC 6 Jul rainfall rate marker in (a) obscures the grid cell with a value $>40 \text{ mm h}^{-1}$ in the CTL integration. The average precipitation rate over the plot domain is shown the top-left corner of each panel.

The length of the squall line increases after transition, and a large region of trailing stratiform precipitation develops by 0900 UTC (Fig. 10e). By 1800 UTC the latter has disappeared, but the remnant bowing line extends for over 700 km across southern Illinois and Indiana (Fig. 10h).

2) SIMULATED EVOLUTION

At the 2100 UTC 5 July initialization time of the GEM integrations, the elevated convective cells that initiate the squall line are already active over central Minnesota (Fig. 10a). Despite the “hot start” strategy, the model’s representation of convective structure is initially relatively poor. In the CTL integration, precipitation maxima bracket the observed location of the developing squall line at 0000 UTC 6 July (cf. Figs. 10b and 14a). When the low-CAPE convection scheme is activated, a coherent precipitation maximum appears in the region, although the convective mode does not reflect the linear structure of the observed rainfall (Fig. 15a).

Despite being the only difference between the CTL and LC integrations, the low-CAPE convection scheme is not directly responsible for the difference in precipitation patterns at 0000 UTC 6 July (Figs. 14a and 15a), and is active only around the periphery of the rainfall maximum (Fig. 16b). However, the parameterization plays an important role in promoting the organization of this feature, as evidenced by its widespread activation in the region just 1 h earlier (Fig. 16a). This elevated

convective initiation is consistent with both the environmental sounding (Fig. 11) and the observed cloud bases at this early stage of the squall-line life cycle.

The cold pool processes involved in the transition from elevated to surface-based convection appear to be well represented in the LC simulation, although they occur at least 3 h

FIG. 9. Time series of maximum total (solid) and convective (dashed) precipitation rates following the convective event in northern Ontario from the model initialization time (2100 UTC 5 Jul 2016) to 0600 UTC 6 Jul (tracks are shown in Fig. 8). Rainfall estimates from MRMS on the 10-km regional grid are shown in black, while those from the CTL and LC integrations are shown in blue and red, respectively, as indicated in the legend. The 0000 UTC 6 Jul valid time considered in the discussion is identified with a vertical dashed line for reference.

FIG. 10. Rain-rate estimates (in mm h^{-1} , as shown on the color bars) from MRMS at 3-hourly intervals from (a) 2100 UTC 5 Jul to (h) 1800 UTC 6 Jul 2016. Also shown in each panel are the locations of the surface stations referred to in the text: target markers for Morris, Minnesota (KMOX); Minneapolis, Minnesota (KMSP); La Crosse, Wisconsin (KLSE); and Rockford, Illinois (KRFD). The location of the Aberdeen sounding (Fig. 11) is shown in (a).

FIG. 11. Aberdeen sounding for 1800 UTC 5 Jul, plotted as in Fig. 4. Green annotations show the effects of lifting a parcel to a 700-hPa cloud base, with a dashed line for the mixing ratio, a dotted line for the potential temperature, and a solid line for undilute moist adiabatic ascent. The sounding location is shown in Figs. 2 and 10a.

earlier than observed. The cold pool formed by downdrafts in the low-CAPE convection scheme extends across southeastern Minnesota after 2 h of integration (Fig. 17). Enhanced upward motion is evident along the leading edge of this feature, consistent with forced ascent at the head of a density current (Wakimoto 1982). The fact that precipitation is light or absent in these areas (less than 0.2 mm of accumulation after 2 h as shown in Fig. 17) suggests that the enhanced ascent is acting to relatively moisten and destabilize the preconvective environment, rather than simply identifying regions in which condensation is fueling buoyant updrafts. Consistent with the transition process described by Marsham et al. (2011), the cold pool is thereby facilitating the initiation of surface-based convection early in the LC integration. Cloud bases drop to 1000 m above ground after just 3 h of integration (0000 UTC 6 July), indicating the completion of the transition well before the developing squall line passes La Crosse, Wisconsin. This represents a transition timing error in the LC integration; however, the lack of elevated convection in the CTL configuration means that these processes are entirely absent in that simulation.

Without a robust initiating cold pool, precipitation patterns in the CTL integration remain relatively disorganized throughout the remainder of the integration (Fig. 14). This

behavior stands in stark contrast to the LC simulation, in which widespread rainfall is predicted along the path of the observed squall line (Fig. 15). The enhanced cold pool in the LC simulation, defined as the LC-minus-CTL difference in screen level temperature, evolves in a manner that aligns with the structural differences in rainfall (Fig. 18). The cold pool difference increases to 6°C in the 6-h forecast (valid at 0300 UTC), consistent with the magnitude of the storm-related cooling observed at La Crosse (Fig. 13). A precursor cold pool also appears ahead of the squall line over northeastern Missouri at this time (Fig. 18b), where the low-CAPE convection scheme accurately predicts convective activity that is absent in the CTL integration (not shown). The enhancement of the primary cold pool increases to a maximum of 8°C by the end of the squall-line life cycle, with cold air extending across Illinois and Indiana behind the well-positioned decaying squall line in the LC integration. Screen-level temperature errors are reduced by ~20% across the region at all lead times (Fig. 18), a result that is indicative of improved cold pool structure in the LC simulation.

5. Impact on NWP guidance

Analysis of the behavior of the low-CAPE convection scheme in the case study context (section 4) is useful for

FIG. 12. Profile of equivalent potential temperature (θ_e) in the Aberdeen sounding for 1800 UTC 5 Jul (Fig. 11), plotted as in Fig. 5.

confirming that the parameterization acts in a physically realistic manner. However, an evaluation of the impact of the scheme on guidance skill over a large number of cases representing different environments is essential for NWP purposes.

Performance of the Regional Deterministic Prediction System is assessed in this section using the same configurations as in the preceding analysis: the CTL integrations are run without a low-CAPE convection scheme, while the LC integrations employ the parameterization described in section 3. The July–August 2016 period is considered here because the introduction of the low-CAPE convection scheme has little systematic impact on winter guidance. This is consistent with the fact that although the bulk of low-CAPE convection in the southeastern United States occurs in the cool season (King et al. 2017), such events are more frequent at night and during the warm season across the remainder of the continent (Evans 2010; Main et al. 2018). Forecasts are initialized from operational analyses at 36-h intervals in order to reduce computational load while promoting serial independence. This dataset is further stratified by initial synoptic hour (0000 and 1200 UTC) for near-surface evaluations to avoid smoothing the diurnal cycle; the conclusions drawn here are valid for both subsets. Initialization with operational analyses implies that this evaluation assesses sensitivity to the activation of the

FIG. 13. Time series of surface METAR observations at La Crosse, Wisconsin, between 0000 and 1200 UTC 6 Jul 2016. (a) Temperature, (b) sea level pressure converted from reported altimeter setting, (c) wind speed, and (d) wind direction. The units for each plot are labeled on the ordinate of the panel. A vertical gray line in all panels identifies the passage of the squall line at 0300 UTC 6 Jul 2016, while horizontal dashed gray lines in (d) identify the cardinal points for reference.

scheme in a forecast sequence rather than a full assimilation cycle. However, results obtained with the full suite of changes documented by McTaggart-Cowan et al. (2019b) demonstrate that the large majority of the impact can be effectively assessed in this forecast-only framework.

The clearest direct impact of the low-CAPE convection scheme is on the precipitation metrics shown in Fig. 19. The equitable threat score is improved at most accumulation thresholds (Fig. 19a), largely as a result of a reduction in the number of false alarms for thresholds of 2 mm (24 h)^{−1} and above (Fig. 19b). The decrease in the false alarm ratio is not accompanied by an equivalent reduction in the probability of detection (Fig. 19c), an indication that the parameterization is acting to eliminate spurious precipitation events in a selective manner. This is particularly encouraging given the

FIG. 14. Precipitation rates at 6-hourly intervals from the CTL integration, split into (left) total and (right) convection-only rates (shaded; in mm h^{-1} following the color bars). Also shown in each panel using a heavy black contour is the position of the observed squall line as analyzed from MRMS data (Fig. 10). The left column of this figure is directly comparable to the right column of Fig. 10.

FIG. 15. As in Fig. 14, but for the LC integration. The subdomain used to investigate early storm evolution in Figs. 16 and 17 is outlined with dark green in (a).

FIG. 16. Precipitation rates (in mm h^{-1}) from the LC integration (a) 2 and (b) 3 h after initialization (valid times as indicated in the panels). The component of precipitation coming from the low-CAPE convection scheme is shown using the magenta lines of the vertical color bar, while the remainder of precipitation from all other sources is shown using the colors of the horizontal color bar. The southern Minnesota domain for this plot is indicated by the dark green outline in Fig. 15a.

dramatic reduction in the frequency of occurrence of accumulations of 5 mm (24 h)^{-1} and greater (Fig. 19d). The latter is consistent with the impact of the parameterization on the convective cells analyzed in northern Ontario case study (section 4a), determined to be a consequence of the parameterized release of potential instability and the suppression of the gridpoint storm feedback.

The impact of the low-CAPE convection scheme on predictive skill for near-surface atmospheric variables is smaller than it is for precipitation (Fig. 20). Improvements to the error standard deviation² of 2-m temperature and moisture, and 10-m winds rise to the level of statistical significance at most lead times, but the change in guidance quality is small enough to be of little practical utility. The same is generally true for changes

² The “error standard deviation” is computed as the standard deviation of model error, thereby depicting the random component of the error to complement the systematic component that is illustrated by the bias (mean error).

FIG. 17. Cold pool enhancement after 2 h of integration, valid at 2300 UTC 5 Jul 2016. Total accumulated precipitation (in mm) is shown with gray shading at the 0.2-mm threshold as indicated on the vertical color bar. Cold pool enhancement is shown as LC-minus-CTL screen-level temperatures (in K) using solid contours blue-shaded as indicated on the horizontal color bar, and vertical motion enhancement (again computed as LC minus CTL) at the 0.85 hybrid-coordinate model level (Girard et al. 2014) is red-stippled for upward motions exceeding 0.5 Pa s^{-1} . The domain for this plot matches that of Fig. 16, as indicated by the dark green outline in Fig. 15a.

to mean errors, although daytime cooling leads to a 0.3-K increase in the cold bias of the LC guidance compared to the CTL (Fig. 20a). This change in 2-m temperature is consistent with the cold pool enhancement noted in the midwestern squall-line case study (Fig. 18), although the relative contributions of such features and the effects of shading by the clouds predicted by the low-CAPE convection scheme has not been systematically assessed in the forecast sequence.

The effects of parameterized low-CAPE convection on upper-air scores are more pronounced, with improvements in error standard deviation clearly visible in height and wind predictions near the tropopause (Figs. 21a,b). However, a reduction in the midtropospheric warm bias (Fig. 21c) leads hydrostatically to a 3-m decrease in upper-tropospheric heights that is evaluated as a degradation compared to radiosonde-observed values. Uncertainty in the sign of the height bias of the CTL integration is large; however, as comparison with the ERA5 (Hersbach et al. 2018) reveals a significant high bias that renders the height reductions in the LC integration a positive result (not shown).

The significant improvements in guidance quality described here may represent an upper bound on the impact that could be expected from the introduction of a low-CAPE convection scheme. This is because deep convection in GEM is depicted using a CAPE-dependent parameterization whose activity is suppressed under conditions of limited conditional instability. In models that employ schemes with alternative closures, convection in such environments may already be adequately represented. This appears to be the case for the ECMWF system, in which the sensitivity to the activation of the equivalent midlevel form of the Tiedtke (1989) scheme is

FIG. 18. Cold pool enhancement shown as LC-minus-CTL screen-level temperatures (shaded; in $^{\circ}\text{C}$, as shown on the color bars) at 6-hourly intervals for valid times from (a) 0000 to (d) 1800 UTC 6 Jul 2016. Also shown in each panel using a heavy black contour is the position of the observed squall line as analyzed from MRMS data (Fig. 10). The root-mean-square error (RMSE) for both simulations is shown in the top-right corner of each panel, computed against analyses from the Regional Deterministic Prediction System run with the latest model configuration (McTaggart-Cowan et al. 2019a). These analyses are used for evaluation because they employ the same grid definition and model orography as the CTL and LC simulations, two factors for which inconsistencies complicate the estimation of near-surface errors.

much smaller than that reported here (P. Bechtold, personal communication).

6. Conclusions

An important subset of convective events occur in environments that are devoid of significant CAPE. Representing such events in atmospheric models that are unable to resolve moist convective processes explicitly is the objective of the low-CAPE convection parameterization introduced in this study.

The scheme is based on a convective plume model and the mass-flux equations, employing distinct triggering and closure components that enhance its ability to represent moist convection in low-CAPE environments. The direct connection of these elements to the grid-scale vertical motion ensures that the scheme is able to tap instabilities through parameterized moist convection instead of permitting the model to release them in underresolved updrafts via the gridpoint storm mechanism.

The introduction of the low-CAPE convection scheme appears to improve precipitation structure under low-CAPE conditions that are not conducive to surface-based deep

convection. The direct impact of the new scheme is apparent in the simulation of a set of small, intense convective cells in northern Ontario: precipitation bull's-eyes are eliminated in the low-CAPE region of potential instability. The indirect impact of the scheme on convective structure is evident in integrations that focus on a squall line over the midwestern United States. The enhanced cold pool generated by the low-CAPE convection scheme early in the simulation leads to the secondary initiation of surface-based convection that forms the leading edge of the system. These results combine with the positive impact of the parameterization on guidance quality to suggest that the scheme is working in a physically realistic manner to represent the effects of convection in low-CAPE environments.

Although potentially unstable profiles have been emphasized in this study, the scheme has the capacity to react to any low-CAPE instability in principle. For example, a moist symmetric instability that leads to excessive vertical motions along a slant path will lead to the scheme's activation: the trigger does not depend on the nature of the instability. Although a dedicated slantwise convection scheme would be

FIG. 19. Precipitation scores for day-2 24-h accumulations (24–48-h forecast time) over North America from 1200 UTC initializations in the summer 2016 forecast sequence (21 cases total). Forecasts from the CTL are plotted in blue, and those from LC are plotted in red for (a) the equitable threat score, (b) the false alarm ratio, (c) the probability of detection, and (d) the frequency bias. Differences between the scores that are significant at the 90% level as determined using a bootstrap test are identified with a filled line marker on the time series with the improved score. Precipitation measurements from synoptic stations that have passed the quality control stage of the [Lepinas et al. \(2015\)](#) precipitation analysis are used as observational data here. The number of observations at each accumulation threshold is annotated in (a).

best suited to represent convection in such cases ([Nordeng 1987](#); [Lindstrom and Nordeng 1992](#); [Balasubramanian and Yau 1994](#); [Schultz and Schumacher 1999](#); [Glinton et al. 2017](#)), the triggering of parameterized upright convection is more physically justified than the alternative of gridscale feedbacks ([Tiedtke 1989](#)). This is particularly true given the tendency for the two forms of moist convection to coexist ([Rasp et al. 2016](#)). The scheme's focus on the presence of lifting rather than the details of the instability is consistent with the ingredients-based view of convection ([Schultz et al. 2000](#)).

Despite the positive impact that the low-CAPE convection parameterization has on model simulations, the selection of free parameters within the scheme introduces new dimensions to the problem of determining an optimal model configuration. These parameters, which include the threshold environmental mass flux for triggering ($M_e^{\min}$), cloud radius (r), lateral entrainment (ϵ) and detrainment (δ) rates, and precipitation efficiency (E_p) all have distinct impacts on the model solution. The parametric estimates of quantities such as the cloud-base updraft velocity (w_{LCL}) represent additional sources of uncertainty. The values employed in this study represent those that yield the largest improvements in deterministic guidance quality; however, scans of the parameter space reveal significant

FIG. 20. North American near-surface scores from 1200 UTC initializations in the summer 2016 forecast sequence (21 cases total) for (a) temperature, (b) dewpoint, and (c) wind speed. The bias (dashed lines) and error standard deviation (solid lines) are used here to quantify forecast quality. Forecasts from CTL are plotted in blue, and those from LC are plotted in red, with units plotted as labeled on the ordinate of each panel. Differences between the time series that are significant at the 90% level as determined from a bootstrap test are identified at 6-hourly intervals with a filled line marker on the time series with the improved score. Observations come from a combination of the synoptic and METAR networks that includes only stations whose altitudes in the model are within 30 m of the recorded station elevation.

sensitivity, with a particularly strong relationship between $M_e^{\min}$ (triggering) and quantitative precipitation forecasts. This indicates that the details of the representation of low-CAPE convection directly influence forecast skill and that constraints on the ranges of these parameters need to be established for the probabilistic context. Further, the problem of making these quantities scale-aware in a physically relevant way represents a challenge that has not yet been investigated in detail.

The low-CAPE convection scheme was implemented in Canadian operational NWP systems as part of a major update to the physical parameterization suite in July 2019 ([McTaggart-Cowan et al. 2019b](#)). Despite the significant improvements in guidance skill that accompanied this upgrade, ongoing follow-up evaluations suggest that model's long-standing positive precipitation bias has been overreduced in convectively active regions. This is particularly true in the eastern United States, where [Lepore et al. \(2015\)](#) show that CAPE is an effective predictor for convective activity. Given the large impact of the new scheme on precipitation totals, opportunities for the refinement of the parameterization and its associated free parameters are being explored.

Despite the increases in computing power that have allowed most regional NWP systems to adopt convection-permitting configurations, most global NWP models, ensemble systems

FIG. 21. Profile of error standard deviation (solid) and bias (dashed) against North American radiosonde observations of (a) geopotential height, (b) zonal wind, (c) temperature, and (d) dewpoint depression after 48 h of forecast time for the summer 2016 period (42 cases total). Forecasts from CTL are plotted in blue, and those from LC are plotted in red, with units plotted as labeled on the abscissa of each panel. Differences between the time series that are significant at the 90% level based on a two-sided Student's t test are identified with a filled line marker on the time series with the improved score. The number of observations for each variable and level is identified on the ordinate to the right of the profile.

and climate applications will need to continue to parameterize moist convection for the foreseeable future. The positive impact of the low-CAPE convection scheme on guidance skill provides a strong incentive for further investigation of the representation of this class of convection in atmospheric models that are unable to resolve completely the moist turbulent processes that may be active in such environments.

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Data availability statement. Access to all data employed in this study is unrestricted. The reanalysis products used in this study can be obtained using via the references provided in section 2, while observational data can be retrieved from archives of the Global Telecommunication System. The large volume of data generated by the integrations described in this study precludes their entry into a public archive, but these data are stored at the Canadian Meteorological Centre and will be made available upon request.

REFERENCES

- Ancell, B. C., A. Bogusz, M. J. Lauridsen, and C. J. Nauert, 2018: Seeding chaos: The dire consequences of numerical noise in NWP perturbation experiments. *Bull. Amer. Meteor. Soc.*, **99**, 615–628, <https://doi.org/10.1175/BAMS-D-17-0129.1>.
- Anderson, C. J., W. A. J. Gallus, R. W. Arritt, and J. S. Kain, 2002: Impact of adjustments in the Kain-Fritsch convective scheme on QPF of elevated convection. *19th Conf. on Weather Analysis and Forecasting/15th Conf. on Numerical Weather Prediction*, Omaha, NE, Amer. Meteor. Soc., 2.2, https://ams.confex.com/ams/SLS_WAF_NWP/techprogram/paper_46840.htm.
- Arakawa, A., J.-H. Jung, and C.-M. Wu, 2016: Multiscale modeling of the moist-convective atmosphere. *Multiscale Convection-Coupled Systems in the Tropics: A Tribute to Dr. Michio Yanai*, Meteor. Monogr., No. 56, Amer. Meteor. Soc., <https://doi.org/10.1175/AMSMONOGRAPHIS-D-15-0014.1>.
- Baba, Y., 2019: Spectral cumulus parameterization based on cloud-resolving model. *Climate Dyn.*, **52**, 309–334, <https://doi.org/10.1007/s00382-018-4137-z>.
- Balasubramanian, G., and M. K. Yau, 1994: Baroclinic instability in a two-layer model with parameterized slantwise convection. *J. Atmos. Sci.*, **51**, 971–990, [https://doi.org/10.1175/1520-0469\(1994\)051<0971:BIATL>2.0.CO;2](https://doi.org/10.1175/1520-0469(1994)051<0971:BIATL>2.0.CO;2).
- Banacos, P. C., and D. M. Schultz, 2005: The use of moisture flux convergence in forecasting convective initiation: Historical and operational perspectives. *Wea. Forecasting*, **20**, 351–366, <https://doi.org/10.1175/WAF858.1>.
- Bechtold, P., E. Bazile, F. Guichard, P. Mascart, and E. Richard, 2001: A mass-flux convection scheme for regional and global models. *Quart. J. Roy. Meteor. Soc.*, **127**, 869–886, <https://doi.org/10.1002/qj.49712757309>.
- Bennett, L. J., K. A. Browning, A. M. Blyth, D. J. Parker, and P. A. Clark, 2006: A review of the initiation of precipitating convection in the United Kingdom. *Quart. J. Roy. Meteor. Soc.*, **132**, 1001–1020, <https://doi.org/10.1256/qj.05.54>.
- Betts, A. K., 1986: A new convective adjustment scheme. Part I: Observational and theoretical basis. *Quart. J. Roy. Meteor. Soc.*, **112**, 667–691, <https://doi.org/10.1002/QJ.49711247307>.
- Blanchard, D. O., 1998: Assessing the vertical distribution of convective available potential energy. *Wea. Forecasting*, **13**, 870–877, [https://doi.org/10.1175/1520-0434\(1998\)013<0870:ATVDOC>2.0.CO;2](https://doi.org/10.1175/1520-0434(1998)013<0870:ATVDOC>2.0.CO;2).
- Bluestein, H. B., and C. R. Parks, 1983: A synoptic and photographic climatology of low-precipitation severe thunderstorms in the Southern Plains. *Mon. Wea. Rev.*, **111**, 2034–2046, [https://doi.org/10.1175/1520-0493\(1983\)111<2034:ASAPCO>2.0.CO;2](https://doi.org/10.1175/1520-0493(1983)111<2034:ASAPCO>2.0.CO;2).
- , and K. W. Thomas, 1984: Diagnosis of a jet streak in the vicinity of a severe weather outbreak in the Texas Panhandle. *Mon. Wea. Rev.*, **112**, 2499–2520, [https://doi.org/10.1175/1520-0493\(1984\)112<2499:DOAJSI>2.0.CO;2](https://doi.org/10.1175/1520-0493(1984)112<2499:DOAJSI>2.0.CO;2).
- Bougeault, P., 1985: A simple parameterization of the large-scale effects of cumulus convection. *Mon. Wea. Rev.*, **113**, 2108–2121, [https://doi.org/10.1175/1520-0493\(1985\)113<2108:ASPTOL>2.0.CO;2](https://doi.org/10.1175/1520-0493(1985)113<2108:ASPTOL>2.0.CO;2).
- Buehner, M., and Coauthors, 2015: Implementation of deterministic weather forecast systems based on ensemble-variational data assimilation at environment Canada. Part I: The global system. *Mon. Wea. Rev.*, **143**, 2532–2559, <https://doi.org/10.1175/MWR-D-14-00354.1>.
- Carlson, T. N., 1980: Airflow through midlatitude cyclones and the comma cloud pattern. *Mon. Wea. Rev.*, **108**, 1498–1509, [https://doi.org/10.1175/1520-0493\(1980\)108<1498:ATMCAT>2.0.CO;2](https://doi.org/10.1175/1520-0493(1980)108<1498:ATMCAT>2.0.CO;2).
- Caron, J.-F., T. Milewski, M. Buehner, L. Fillion, M. M. Reszka, S. Macpherson, and J. St-James, 2015: Implementation of deterministic weather forecast systems based on ensemble-variational data assimilation at Environment Canada. Part II: The regional system. *Mon. Wea. Rev.*, **143**, 2560–2580, <https://doi.org/10.1175/MWR-D-14-00353.1>.
- Carr, F. H., and J. P. Millard, 1985: A composite study of comma clouds and their association with severe weather over the Great Plains. *Mon. Wea. Rev.*, **113**, 370–387, [https://doi.org/10.1175/1520-0493\(1985\)113<0370:ACSOCC>2.0.CO;2](https://doi.org/10.1175/1520-0493(1985)113<0370:ACSOCC>2.0.CO;2).
- Chikira, M., and M. Sugiyama, 2010: A cumulus parameterization with state-dependent entrainment rate. Part I: Description and sensitivity to temperature and humidity profiles. *J. Atmos. Sci.*, **67**, 2171–2193, <https://doi.org/10.1175/2010JAS316.1>.
- Clark, M. R., 2009: The southern England tornadoes of 30 December 2006: Case study of a tornadic storm in a low CAPE, high shear environment. *Atmos. Res.*, **93**, 50–65, <https://doi.org/10.1016/j.atmosres.2008.10.008>.
- Colman, B. R., 1990a: Thunderstorms above frontal surfaces in environments without positive CAPE. Part I: A climatology. *Mon. Wea. Rev.*, **118**, 1103–1122, [https://doi.org/10.1175/1520-0493\(1990\)118<1103:TAFSIE>2.0.CO;2](https://doi.org/10.1175/1520-0493(1990)118<1103:TAFSIE>2.0.CO;2).
- , 1990b: Thunderstorms above frontal surfaces in environments without positive CAPE. Part II: Organization and instability mechanisms. *Mon. Wea. Rev.*, **118**, 1123–1144, [https://doi.org/10.1175/1520-0493\(1990\)118<1123:TAFSIE>2.0.CO;2](https://doi.org/10.1175/1520-0493(1990)118<1123:TAFSIE>2.0.CO;2).
- Corfidi, S. F., S. J. Corfidi, and D. M. Schultz, 2008: Elevated convection and castellanus: Ambiguities, significance, and questions. *Wea. Forecasting*, **23**, 1280–1303, <https://doi.org/10.1175/2008WAF2222118.1>.
- Côté, J., S. Gravel, A. Méthot, A. Patoine, M. Roch, and A. Staniforth, 1998: The Operational CMC-MRB Global Environmental Multiscale (GEM) model. Part I: Design considerations and formulation. *Mon. Wea. Rev.*, **126**, 1373–1395, [https://doi.org/10.1175/1520-0493\(1998\)126<1373:TOCMGE>2.0.CO;2](https://doi.org/10.1175/1520-0493(1998)126<1373:TOCMGE>2.0.CO;2).
- Davis, J. M., and M. D. Parker, 2014: Radar climatology of tornadic and nontornadic vortices in high-shear, low-CAPE environments in the mid-Atlantic and southeastern United States. *Wea. Forecasting*, **29**, 828–853, <https://doi.org/10.1175/WAF-D-13-00127.1>.
- Dee, D. P., and Coauthors, 2011: The Era-Interim reanalysis: Configuration and performance of the data assimilation system. *Quart. J. Roy. Meteor. Soc.*, **137**, 553–597, <https://doi.org/10.1002/qj.828>.
- Doswell, C. A., 1987: The distinction between large-scale and mesoscale contribution to severe convection. *Wea. Forecasting*, **2**, 3–16, [https://doi.org/10.1175/1520-0434\(1987\)002<0003:TDBLSA>2.0.CO;2](https://doi.org/10.1175/1520-0434(1987)002<0003:TDBLSA>2.0.CO;2).
- , H. E. Brooks, and R. A. Maddox, 1996: Flash flood forecasting: An ingredients-based methodology. *Wea. Forecasting*, **11**, 560–581, [https://doi.org/10.1175/1520-0434\(1996\)011<0560:FFFAIB>2.0.CO;2](https://doi.org/10.1175/1520-0434(1996)011<0560:FFFAIB>2.0.CO;2).
- Eckhardt, S., A. Stohl, H. Wernli, P. James, C. Forster, and N. Spichtinger, 2004: A 15-year climatology of warm conveyor belts. *J. Climate*, **17**, 218–237, [https://doi.org/10.1175/1520-0442\(2004\)017<0218:AYCOWC>2.0.CO;2](https://doi.org/10.1175/1520-0442(2004)017<0218:AYCOWC>2.0.CO;2).
- ECMWF, 2018: *IFS Documentation—Cy45r1. Part IV: Physical Processes*. ECMWF, 223 pp.
- Evans, M., 2010: An examination of low CAPE/high shear severe convective events in the Binghamton, New York county warning area. *Natl. Wea. Dig.*, **34**, 129–144.

- Fankhauser, J. C., 1988: Estimates of thunderstorm precipitation efficiency from field measurements in CCOPE. *Mon. Wea. Rev.*, **116**, 663–684, [https://doi.org/10.1175/1520-0493\(1988\)116<0663:EOTPEF>2.0.CO;2](https://doi.org/10.1175/1520-0493(1988)116<0663:EOTPEF>2.0.CO;2).
- Fritsch, J. M., and C. F. Chappell, 1980: Numerical prediction of convectively driven mesoscale pressure systems. Part I: Convective parameterization. *J. Atmos. Sci.*, **37**, 1722–1733, [https://doi.org/10.1175/1520-0469\(1980\)037<1722:NPOCDM>2.0.CO;2](https://doi.org/10.1175/1520-0469(1980)037<1722:NPOCDM>2.0.CO;2).
- Girard, C., and Coauthors, 2014: Staggered vertical discretization of the Canadian Global Environmental Multiscale (GEM) model using a coordinate of the log-hydrostatic-pressure type. *Mon. Wea. Rev.*, **142**, 1183–1196, <https://doi.org/10.1175/MWR-D-13-00255.1>.
- Glinton, M. R., S. L. Gray, J. M. Chagnon, and C. J. Morcrette, 2017: Modulation of precipitation by conditional symmetric instability release. *Atmos. Res.*, **185**, 186–201, <https://doi.org/10.1016/j.atmosres.2016.10.013>.
- Gray, K., and J. Frame, 2019: Investigating the transition from elevated multicellular convection to surface-based supercells during the tornado outbreak of 24 August 2016 using a WRF Model simulation. *Wea. Forecasting*, **34**, 1051–1079, <https://doi.org/10.1175/WAF-D-18-0209.1>.
- Gregory, D., 1996: Parameterization of convective momentum transports in the ECMWF model: Evaluation using cloud resolving models and impact upon model climate. *Proc. Workshop on New Insights and Approaches to Convective Parameterization*, Shinfield, Reading, ECMWF, 208–227, <https://www.ecmwf.int/node/9644>.
- , and P. R. Rowntree, 1990: A mass flux convection scheme with representation of cloud ensemble characteristics and stability-dependent closure. *Mon. Wea. Rev.*, **118**, 1483–1506, [https://doi.org/10.1175/1520-0493\(1990\)118<1483:AMFCSW>2.0.CO;2](https://doi.org/10.1175/1520-0493(1990)118<1483:AMFCSW>2.0.CO;2).
- Hersbach, H., and Coauthors, 2018: Operational global reanalysis: Progress, future directions and synergies with NWP. ERA Rep. 27, 63 pp., <https://www.ecmwf.int/node/18765>.
- Hill, C., C. DeLuca, V. Balaji, M. Suarez, and A. da Silva, 2004: The architecture of the earth system modeling framework. *Comput. Sci. Eng.*, **6**, 18–28, <https://doi.org/10.1109/MCISE.2004.1255817>.
- Holloway, C. E., and Coauthors, 2014: Understanding and representing atmospheric convection across scales: Recommendations from the meeting held at Darlington Hall, Devon, UK, 28–30 January 2013. *Atmos. Sci. Lett.*, **15**, 348–353.
- Horgan, K. L., D. M. Shultz, J. E. Hales Jr., S. F. Corfidi, and R. H. Johns, 2007: A five-year climatology of elevated severe convective storms in the United States east of the Rocky Mountains. *Wea. Forecasting*, **22**, 1031–1044, <https://doi.org/10.1175/WAF1032.1>.
- Husain, S. Z., and C. Girard, 2017: Impact of consistent semi-Lagrangian trajectory calculations on numerical weather prediction performance. *Mon. Wea. Rev.*, **145**, 4127–4150, <https://doi.org/10.1175/MWR-D-17-0138.1>.
- Jewett, B. F., and R. B. Wilhelmson, 2006: The role of forcing in cell morphology and evolution within midlatitude squall lines. *Mon. Wea. Rev.*, **134**, 3714–3734, <https://doi.org/10.1175/MWR3164.1>.
- Johnson, R. H., and P. J. Hamilton, 1988: The relationship of surface pressure features to the precipitation and airflow structure of an intense midlatitude squall line. *Mon. Wea. Rev.*, **116**, 1444–1473, [https://doi.org/10.1175/1520-0493\(1988\)116<1444:TROSPF>2.0.CO;2](https://doi.org/10.1175/1520-0493(1988)116<1444:TROSPF>2.0.CO;2).
- Kain, J. S., 2004: The Kain–Fritsch convective parameterization: An update. *J. Appl. Meteor.*, **43**, 170–181, [https://doi.org/10.1175/1520-0450\(2004\)043<0170:TKCPAU>2.0.CO;2](https://doi.org/10.1175/1520-0450(2004)043<0170:TKCPAU>2.0.CO;2).
- , and J. M. Fritsch, 1990: A one-dimensional entraining/detraining plume model and its application in convective parameterization. *J. Atmos. Sci.*, **47**, 2784–2802, [https://doi.org/10.1175/1520-0469\(1990\)047<2784:AODEPM>2.0.CO;2](https://doi.org/10.1175/1520-0469(1990)047<2784:AODEPM>2.0.CO;2).
- , and —, 1993: Convective parameterization for mesoscale models: The Kain–Fritsch scheme. *The Representation of Cumulus Convection in Numerical Models of the Atmosphere*, Meteor. Monogr., No. 46, Amer. Meteor. Soc., 165–170.
- , and —, 1998: Multiscale convective overturning in mesoscale convective systems: Reconciling observations, simulations, and theory. *Mon. Wea. Rev.*, **126**, 2254–2273, [https://doi.org/10.1175/1520-0493\(1998\)126<2254:MCOIMC>2.0.CO;2](https://doi.org/10.1175/1520-0493(1998)126<2254:MCOIMC>2.0.CO;2).
- Kato, T., and K. Saito, 1995: Hydrostatic and non-hydrostatic simulations of moist convection: Applicability of the hydrostatic approximation to a higher-resolution model. *J. Meteor. Soc. Japan*, **73**, 59–77, https://doi.org/10.2151/jmsj1965.73.1_59.
- Keyser, D., M. J. Reeder, and R. J. Reed, 1988: A generalization of Pettersen's frontogenesis function and its relation to the forcing of vertical motion. *Mon. Wea. Rev.*, **116**, 762–781, [https://doi.org/10.1175/1520-0493\(1988\)116<0762:AGOPFF>2.0.CO;2](https://doi.org/10.1175/1520-0493(1988)116<0762:AGOPFF>2.0.CO;2).
- King, J. R., M. D. Parker, K. D. Sherburn, and G. M. Lackmann, 2017: Rapid evolution of cool season, low-CAPE severe thunderstorm environments. *Wea. Forecasting*, **32**, 763–779, <https://doi.org/10.1175/WAF-D-16-0141.1>.
- Kreitzberg, C. W., and D. J. Perkey, 1976: Release of potential instability. Part I: A sequential plume model within a hydrostatic primitive equation model. *J. Atmos. Sci.*, **33**, 456–475, [https://doi.org/10.1175/1520-0469\(1976\)033<0456:ROPIPI>2.0.CO;2](https://doi.org/10.1175/1520-0469(1976)033<0456:ROPIPI>2.0.CO;2).
- Lepore, C., D. Veneziano, and A. Molini, 2015: Temperature and cape dependence of rainfall extremes in the eastern United States. *Geophys. Res. Lett.*, **42**, 74–83, <https://doi.org/10.1002/2014GL062247>.
- Lespinas, F., V. Fortin, G. Roy, P. Rasmussen, and T. Stadnyk, 2015: Performance evaluation of the Canadian Precipitation Analysis (CaPA). *J. Hydrometeorol.*, **16**, 2045–2064, <https://doi.org/10.1175/JHM-D-14-0191.1>.
- Lindstrom, S. S., and T. E. Nordeng, 1992: Parameterized slantwise convection in a numerical model. *Mon. Wea. Rev.*, **120**, 742–756, [https://doi.org/10.1175/1520-0493\(1992\)120<0742:PSCIAN>2.0.CO;2](https://doi.org/10.1175/1520-0493(1992)120<0742:PSCIAN>2.0.CO;2).
- Main, M. E., R. A. Lazear, and L. F. Bosart, 2018: Differences between high shear/low CAPE environments in the Northeast US favoring straight-line damaging winds versus tornadoes. Honours thesis, Dept. of Atmospheric and Environmental Sciences, State University of New York at Albany, 50 pp., https://scholarsarchive.library.albany.edu/cgi/viewcontent.cgi?article=1014&context=honorscollege_daes.
- Marion, G. R., and R. J. Trapp, 2019: The dynamical coupling of convective updrafts, downdrafts, and cold pools in simulated supercell thunderstorms. *J. Geophys. Res. Atmos.*, **124**, 664–683, <https://doi.org/10.1029/2018JD029055>.
- Market, P. S., S. M. Rochette, J. Shewchuk, R. Difani, J. S. Kastman, C. B. Henson, and N. I. Fox, 2017: Evaluating elevated convection with the downdraft convective inhibition. *Atmos. Sci. Lett.*, **18**, 76–81, <https://doi.org/10.1002/asl.727>.
- Marshall, J. H., S. B. Trier, T. M. Weckwerth, and J. W. Wilson, 2011: Observations of elevated convection initiation leading to a surface-based squall line during 13 June IHOP_2002. *Mon. Wea. Rev.*, **139**, 247–271, <https://doi.org/10.1175/2010MWR3422.1>.
- McCann, D. W., 1978: Severe convective storms with little or no thunder. *Natl. Wea. Dig.*, **3**, 2–5.

- McTaggart-Cowan, R., and Coauthors, 2019a: Modernization of atmospheric physics parameterization in Canadian NWP. *J. Adv. Model. Earth Syst.*, **11**, 3593–3635, <https://doi.org/10.1029/2019MS001781>.
- , P. A. Vaillancourt, A. Zadra, L. Separovic, S. Corvec, and D. Kirshbaum, 2019b: A Lagrangian perspective on parameterizing deep convection. *Mon. Wea. Rev.*, **147**, 4127–4149, <https://doi.org/10.1175/MWR-D-19-0164.1>.
- Molinari, J., and M. Dudek, 1992: Parameterization of convective precipitation in mesoscale numerical models: A critical review. *Mon. Wea. Rev.*, **120**, 326–344, [https://doi.org/10.1175/1520-0493\(1992\)120<0326:POCPIM>2.0.CO;2](https://doi.org/10.1175/1520-0493(1992)120<0326:POCPIM>2.0.CO;2).
- Moore, J. T., F. H. Glass, C. E. Graves, S. M. Rochette, and M. J. Singer, 2003: The environment of warm-season elevated thunderstorms associated with heavy rainfall over the central United States. *Wea. Forecasting*, **18**, 861–878, [https://doi.org/10.1175/1520-0434\(2003\)018<0861:TEOWET>2.0.CO;2](https://doi.org/10.1175/1520-0434(2003)018<0861:TEOWET>2.0.CO;2).
- Namias, J., and P. F. Clapp, 1949: Confluence theory of the high tropopause jet stream. *J. Meteor.*, **6**, 330–336, [https://doi.org/10.1175/1520-0469\(1949\)006<0330:CTOTHT>2.0.CO;2](https://doi.org/10.1175/1520-0469(1949)006<0330:CTOTHT>2.0.CO;2).
- Neiman, P. J., M. A. Shapiro, and L. S. Fedor, 1993: The life cycle of an extratropical marine cyclone. Part II: Mesoscale structure and diagnostics. *Mon. Wea. Rev.*, **121**, 2177–2199, [https://doi.org/10.1175/1520-0493\(1993\)121<2177:TLCOAE>2.0.CO;2](https://doi.org/10.1175/1520-0493(1993)121<2177:TLCOAE>2.0.CO;2).
- Niziol, T. A., W. R. Snyder, and J. S. Waldstreicher, 1995: Winter weather forecasting throughout the eastern United States. Part IV: Lake effect snow. *Wea. Forecasting*, **10**, 61–77, [https://doi.org/10.1175/1520-0434\(1995\)010<0061:WWFTE>2.0.CO;2](https://doi.org/10.1175/1520-0434(1995)010<0061:WWFTE>2.0.CO;2).
- Nordeng, T. E., 1987: The effect of vertical and slantwise convection on the simulation of polar lows. *Tellus*, **39A**, 354–375, <https://doi.org/10.1111/j.1600-0870.1987.tb00313.x>.
- Ogura, Y., and H.-R. Cho, 1973: Diagnostic determination of cumulus cloud populations from observed large-scale variabilities. *J. Atmos. Sci.*, **30**, 1276–1286, [https://doi.org/10.1175/1520-0469\(1973\)030<1276:DDOCCP>2.0.CO;2](https://doi.org/10.1175/1520-0469(1973)030<1276:DDOCCP>2.0.CO;2).
- Petch, J. C., A. R. Brown, and M. E. B. Gray, 2002: The impact of horizontal resolution on the simulations of convective development over land. *Quart. J. Roy. Meteor. Soc.*, **128**, 2031–2044, <https://doi.org/10.1256/003590002320603511>.
- Petterssen, S., 1936: Contribution to the theory of frontogenesis. *Geophys. Pub.*, **11**, 1–27.
- Quass, J., 2012: Evaluating the “critical relative humidity” as a measure of subgrid-scale variability of humidity in general circulation model cloud cover parameterizations using satellite data. *J. Geophys. Res.*, **117**, D09208, <https://doi.org/10.1029/2012JD017495>.
- Randall, D. A., P. Ding, and D. M. Pan, 1997: The physics and parameterization of moist atmospheric convection. *The Arakawa-Schubert Parameterization*, Vol. 505, NATO ASI Series, Springer, 281–296.
- Rasp, S., T. Selz, and G. C. Craig, 2016: Convective and slantwise trajectory ascent in convection-permitting simulations of midlatitude cyclones. *Mon. Wea. Rev.*, **144**, 3961–3976, <https://doi.org/10.1175/MWR-D-16-0112.1>.
- Roberts, N. M., and H. W. Lean, 2008: Scale-selective verification of rainfall accumulations from high-resolution forecasts of convective events. *Mon. Wea. Rev.*, **136**, 78–97, <https://doi.org/10.1175/2007MWR2123.1>.
- Rockwood, A. A., and R. A. Maddox, 1988: Mesoscale and synoptic scale interactions leading to intense convection: The case study of 7 June 1982. *Wea. Forecasting*, **3**, 51–68, [https://doi.org/10.1175/1520-0434\(1988\)003<0051:MASSIL>2.0.CO;2](https://doi.org/10.1175/1520-0434(1988)003<0051:MASSIL>2.0.CO;2).
- Rosenow, A. A., R. M. Rauber, B. R. Jewett, G. M. McFarquhar, and J. M. Keeler, 2018: Elevated potential instability in the comma head: Distribution and development. *Mon. Wea. Rev.*, **146**, 1259–1278, <https://doi.org/10.1175/MWR-D-17-0283.1>.
- Rossby, C.-G., 1932: Thermodynamics applied to air analysis. *MIT Meteor. Papers*, **1**, 31–48.
- Schultz, D. M., and P. M. Schumacher, 1999: The use and misuse of conditional symmetric instability. *Mon. Wea. Rev.*, **127**, 2709–2732, [https://doi.org/10.1175/1520-0493\(1999\)127<2709:TUAMOC>2.0.CO;2](https://doi.org/10.1175/1520-0493(1999)127<2709:TUAMOC>2.0.CO;2).
- , P. N. Schumacher, and C. A. Doswell III, 2000: The intricacies of instabilities. *Mon. Wea. Rev.*, **128**, 4143–4148, [https://doi.org/10.1175/1520-0493\(2000\)129<4143:TIOI>2.0.CO;2](https://doi.org/10.1175/1520-0493(2000)129<4143:TIOI>2.0.CO;2).
- Scinocca, J. F., and N. A. McFarlane, 2004: The variability of modeled tropical precipitation. *J. Atmos. Sci.*, **61**, 1993–2015, [https://doi.org/10.1175/1520-0469\(2004\)061<1993:TVOMTP>2.0.CO;2](https://doi.org/10.1175/1520-0469(2004)061<1993:TVOMTP>2.0.CO;2).
- Sherburn, K. D., and M. D. Parker, 2014: Climatology and ingredients of significant severe convection in high-shear, low-CAPE environments. *Wea. Forecasting*, **29**, 854–877, <https://doi.org/10.1175/WAF-D-13-00041.1>.
- , —, J. R. King, and G. M. Lackmann, 2016: Composite environments of severe and nonsevere high-shear, low-CAPE convective events. *Wea. Forecasting*, **31**, 1899–1927, <https://doi.org/10.1175/WAF-D-16-0086.1>.
- Sherwood, S. C., 2000: On moist instability. *Mon. Wea. Rev.*, **128**, 4139–4142, [https://doi.org/10.1175/1520-0493\(2000\)129<4139:OMI>2.0.CO;2](https://doi.org/10.1175/1520-0493(2000)129<4139:OMI>2.0.CO;2).
- Simpson, J., 1983: Cumulus clouds: interactions between laboratory experiments and observations as foundations for models. *Mesoscale Meteorology*, D. K. Lilly and T. Gal-Chen, Eds., Springer, 339–412.
- , and V. Wiggert, 1969: Models of precipitating cumulus towers. *Mon. Wea. Rev.*, **97**, 471–489, [https://doi.org/10.1175/1520-0493\(1969\)097<0471:MOPCT>2.3.CO;2](https://doi.org/10.1175/1520-0493(1969)097<0471:MOPCT>2.3.CO;2).
- Suhas, E., and G. J. Zhang, 2014: Evaluation of trigger functions for convective parameterization schemes using observations. *J. Climate*, **27**, 7647–7666, <https://doi.org/10.1175/JCLI-D-13-00718.1>.
- , and —, 2015: Evaluating convective parameterization closures using cloud-resolving model simulation of tropical deep convection. *J. Geophys. Res. Atmos.*, **120**, 1260–1277, <https://doi.org/10.1002/2014JD022246>.
- Sundqvist, H., 1978: A parameterization scheme for non-convective condensation including prediction of cloud water content. *Quart. J. Roy. Meteor. Soc.*, **104**, 677–690, <https://doi.org/10.1002/qj.49710444110>.
- Teixeira, J., 2001: Cloud fraction and relative humidity in a prognostic cloud fraction scheme. *Mon. Wea. Rev.*, **129**, 1750–1753, [https://doi.org/10.1175/1520-0493\(2001\)129<1750:CFARHI>2.0.CO;2](https://doi.org/10.1175/1520-0493(2001)129<1750:CFARHI>2.0.CO;2).
- Tiedtke, M., 1989: A comprehensive mass flux scheme for cumulus parameterization in large scale models. *Mon. Wea. Rev.*, **117**, 1779–1800, [https://doi.org/10.1175/1520-0493\(1989\)117<1779:ACMFSF>2.0.CO;2](https://doi.org/10.1175/1520-0493(1989)117<1779:ACMFSF>2.0.CO;2).
- Tomas, R. A., and P. J. Webster, 1997: The role of inertial instability in determining the location and strength of near-equatorial convection. *Quart. J. Roy. Meteor. Soc.*, **123**, 1445–1482, <https://doi.org/10.1002/qj.49712354202>.
- Trier, S. B., and D. B. Parsons, 1993: Evolution of environmental conditions preceding the development of a nocturnal mesoscale convective complex. *Mon. Wea. Rev.*, **121**, 1078–1098,

- [https://doi.org/10.1175/1520-0493\(1993\)121<1078:EOECPT>2.0.CO;2](https://doi.org/10.1175/1520-0493(1993)121<1078:EOECPT>2.0.CO;2).
- Uccellini, L. W., and D. R. Johnson, 1979: The coupling of upper and lower tropospheric jet streaks and implications for the development of severe convective storms. *Mon. Wea. Rev.*, **107**, 682–703, [https://doi.org/10.1175/1520-0493\(1979\)107<0682:TCOUAL>2.0.CO;2](https://doi.org/10.1175/1520-0493(1979)107<0682:TCOUAL>2.0.CO;2).
- van Den Broeke, M. S., D. M. Schultz, R. H. Johns, J. S. Evans, and J. E. Hales, 2005: Cloud-to-ground lightning production in strongly forced, low-instability convective lines associated with damaging winds. *Wea. Forecasting*, **20**, 517–530, <https://doi.org/10.1175/WAF876.1>.
- Vescio, M. D., and R. L. Thompson, 2001: Forecaster's Forum: Subjective tornado probability forecasts in severe weather watches. *Wea. Forecasting*, **16**, 192–195, [https://doi.org/10.1175/1520-0434\(2001\)016<0192:FSFSTP>2.0.CO;2](https://doi.org/10.1175/1520-0434(2001)016<0192:FSFSTP>2.0.CO;2).
- Wakimoto, R. M., 1982: Life cycle of thunderstorm gust fronts as viewed with Doppler radar and rawinsonde data. *Mon. Wea. Rev.*, **110**, 1060–1082, [https://doi.org/10.1175/1520-0493\(1982\)110<1060:TLCOTG>2.0.CO;2](https://doi.org/10.1175/1520-0493(1982)110<1060:TLCOTG>2.0.CO;2).
- Weckwerth, T. M., J. Hanesiak, J. W. Wilson, S. B. Trier, S. K. Delegia, W. A. Gallus Jr., R. D. Roberts, and X. Wang, 2019: Nocturnal convection initiation during PECAN 2015. *Bull. Amer. Meteor. Soc.*, **100**, 2223–2239, <https://doi.org/10.1175/BAMS-D-18-0299.1>.
- Weisman, M. L., W. C. Skamarock, and J. B. Klemp, 1997: The resolution dependence of explicitly modeled convective systems. *Mon. Wea. Rev.*, **125**, 527–548, [https://doi.org/10.1175/1520-0493\(1997\)125<0527:TRDOEM>2.0.CO;2](https://doi.org/10.1175/1520-0493(1997)125<0527:TRDOEM>2.0.CO;2).
- Yano, J.-I., M. Bister, and Ž. Fuchs, L. Gerard, V. T. J. Phillips, S. Barkidija, and J.-M. Piriou, 2013: Phenomenology of convection-parameterization closure. *Atmos. Chem. Phys.*, **13**, 4111–4131, <https://doi.org/10.5194/acp-13-4111-2013>.
- Zhang, G. J., and N. A. McFarlane, 1995: Sensitivity of climate simulations to the parameterization of cumulus convection in the Canadian Climate Centre general circulation model. *Atmos.–Ocean*, **33**, 407–446, <https://doi.org/10.1080/07055900.1995.9649539>.
- , and X. Wu, 2003: Convective momentum transport and perturbation pressure field from a cloud-resolving model simulation. *J. Atmos. Sci.*, **60**, 1120–1139, [https://doi.org/10.1175/1520-0469\(2003\)060<1120:CMTAPP>2.0.CO;2](https://doi.org/10.1175/1520-0469(2003)060<1120:CMTAPP>2.0.CO;2).
- Zhang, J., and Coauthors, 2016: Multi-Radar Multi-Sensor (MRMS) quantitative precipitation estimation: Initial operating capabilities. *Bull. Amer. Meteor. Soc.*, **97**, 621–638, <https://doi.org/10.1175/BAMS-D-14-00174.1>.