The Sensitivity of Momentum Transport and Severe Surface Winds to Environmental Moisture in Idealized Simulations of a Mesoscale Convective System

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ABSTRACT

Analysis of a pair of three-dimensional simulations of mesoscale convective systems (MCSs) reveals a significant sensitivity of convective momentum transport (CMT), MCS motion, and the generation of severe surface winds to ambient moisture. The Weather Research and Forecasting model is used to simulate an idealized MCS, which is compared with an MCS in a drier midlevel environment. The MCS in the drier environment is smaller, moves slightly faster, and exhibits increased descent and more strongly focused areas of enhanced CMT near the surface in the trailing stratiform region relative to that in the control simulation. A marked increase in the occurrence of severe surface winds is observed between the dry midlevel simulation and the control. It is shown that the enhanced downward motion associated with decreased midlevel relative humidity affects CMT fields and contributes to an increase in the number of grid-cell occurrences of severe surface winds. The role of a descending rear-inflow jet in producing strong surface winds at locations trailing the gust front is also analyzed, and is found to be associated with low-level CMT maxima, particularly in the drier midlevel simulation.

1. Motivation

a. Convective momentum transport in mesoscale convective systems

Mesoscale convective systems are responsible for a large proportion of global precipitation, as well as significant transport of heat, mass, and momentum throughout the atmosphere (e.g., Houze 1989, 2004). A common type of mesoscale convective system (MCS) is the “leading convective-line trailing-stratiform” MCS (e.g., Newton 1950; Houze et al. 1990; Parker and Johnson 2000), characterized by a leading line of intense convection followed by a region of stratiform precipitation. The trailing stratiform region (TSR) is typically 50–200 km wide and features two main flow regimes: front-to-rear flow that ascends from lower to upper levels, and rear-to-front flow that descends from middle to low levels (Fig. 1). The vertical transport of horizontal momentum by these two branches of circulation within an MCS has been examined by previous studies based on its potential importance to large-scale feedbacks, parameterization in numerical models, MCS motion, and severe surface wind generation (e.g., LeMone 1983; Grubišić and Moncrieff 2000; Geerts 2001; Mechem et al. 2006; Mahoney et al. 2009).

Analysis of a numerically simulated MCS by Mahoney et al. (2009) presented a conceptual picture of convective momentum transport (CMT), linking both the horizontal acceleration of mid- to low-level flow in the direction of storm motion as part of a pressure-gradient-driven rear-inflow jet (PGA in Fig. 1), and the downward transport of the total horizontal wind field (the CMT terms in Fig. 1). CMT can be subdivided into the downward transport of (i) the mean cloud-bearing wind \[\text{CMT}(u)\] and (ii) the storm perturbation flow \[\text{CMT}(u')\].

Air motions within an idealized MCS embedded in westerly flow and moving generally eastward can be largely described by the zonal momentum equation:
where $\mathbf{u}$ is the environmental wind field, $u_{95}$, the horizontal wind has been partitioned into system and system-relative components (i.e., $u = u_{SR} + c_x$ and $v = v_{SR} + c_y$, where $c_x$ and $c_y$ are the speeds at which the MCS moves in the east–west and north–south directions, respectively), and other terms have their usual meanings.

From Fig. 1, it is evident that the relevant terms in (1) to the CMT process are $\frac{\partial p}{\partial t}$, $\frac{\partial u}{\partial x}$, and $\frac{\partial u}{\partial z}$. The pressure-gradient acceleration term [PGA, $\frac{\partial p}{\partial t}$] is fundamental to the formation and acceleration of the midlevel rear inflow jet (RIJ), thus contributing to an increase in the perturbation flow $u'$. The latter two terms describe the actual transport of the environmental wind [CMT($\mathbf{u}$)] and the MCS perturbation wind [CMT($u'$)], as shown in Fig. 1. As the systems analyzed in this study move toward both the east and the south, subsequent assessments of perturbation momentum transport [CMT($u'$)] will also include CMT($u'$) [where CMT($u'$) is $-\frac{1}{\rho} \frac{\partial p}{\partial x}$ when (1) is expressed in the north–south coordinate system as opposed to east–west].

Additional motivation to better understand CMT across a variety of storm types and environments stems from the ambiguity surrounding the role of midlevel dry

\[ \frac{\partial u}{\partial t} + c_x \frac{\partial u}{\partial x} + c_y \frac{\partial u}{\partial y} = -(u - c_x) \frac{\partial u}{\partial x} - (v - c_y) \frac{\partial u}{\partial y} - w \frac{\partial u'}{\partial z} - \frac{\partial p}{\partial z} + f v - \frac{1}{\rho} \frac{\partial p}{\partial x}, \] (1)

FIG. 1. Schematic of the CMT process in an idealized, eastward-moving MCS. Arrows illustrate relative contributions of the PGA, and downward transport of the horizontal environmental wind [CMT($\mathbf{u}$)] and downward transport of the MCS-induced perturbation flow [CMT($u'$)] as indicated (thickness of arrows gives an approximate indication of the relative magnitude of terms).

Mahoney et al. (2009) illustrate the importance of CMT to MCS motion, and several previous studies discuss the process in general terms with respect to severe surface wind generation (e.g., Foster 1958; Johns and Doswell 1992; Weisman 1992; Geerts 2001; Kuchera and Parker 2006). Such studies have discussed the idea that CMT may be a driver of severe surface winds, but precise relationships and an understanding of the complex physical processes involved remain elusive. Mahoney et al. (2009) also discuss the importance of the strength and location of the RIJ with respect to CMT, MCS motion, and severe surface winds.

Yang and Houze (1995, hereafter YH95) investigate the sensitivity of MCS rear inflow to ice microphysics and environmental humidity, and their results reveal considerable sensitivity in these relationships. Thus, it may be reasonably hypothesized that if the role of CMT is heavily dependent on the strength and degree of descent of rear inflow, and rear inflow reveals strong sensitivity to environmental moisture and microphysical processes, then these sensitivities should be evident in both CMT magnitude and MCS motion. Indeed, YH95 do note differences in MCS ground speeds (their Table 2); however, such changes are not the focus of their experiments and are thus not analyzed in detail.
air in MCS development and severe surface wind generation. This relationship is the focus of recent work by James and Markowski (2010), in which they illustrate that the perception that dry air unambiguously favors downdraft and outflow strength is in reality more complex, and depends strongly on the storm environment as well as ice microphysical processes. A similar question is posed here: If dry air aloft indeed increases subsidence and downward mass flux, do increases in low-level CMT lead to significant dynamical changes in an MCS? Specifically, do such changes affect MCS ground speed and/or the speed and location of the system’s surface winds?

b. Severe surface winds

A recurring theme in past studies of convectively driven windstorms has been the challenge of precisely documenting the mechanisms that produce intense thunderstorm outflow at the earth’s surface. Wakimoto (2001) provides a comprehensive summary of this body of literature, and lists several phenomena that remain incompletely understood and compel additional research. Included in this list are (i) determining the cause of small-scale damage swaths (such as those associated with gust-front mesovortices), (ii) understanding processes that inhibit or augment RIJ descent to the surface, and (iii) elucidating the role of microphysical processes in the genesis of convective downdrafts.

Descending RIJs have been observed to produce damaging surface winds, particularly at the apex of an MCS bow echo (e.g., Fujita 1978; Wheatley et al. 2006). Observations show that damage can occur along a storm’s leading edge both in isolated, small-scale swaths (as in the case of mesovortices) and also in larger, more diffuse areas (e.g., Forbes and Wakimoto 1983; Johns and Hirt 1987; Wakimoto et al. 2006). Wheatley et al. (2006) and Kuchera and Parker (2006) discuss how these two factions of severe thunderstorm wind mechanisms have typically been treated as distinct problems in the literature, with most studies separating intense thunderstorm downdrafts due to precipitation or dynamic pressure forces (e.g., Fujita and Wakimoto 1981) from winds produced by more persistent, larger-scale features such as descending RIJs associated with bow echoes and MCSs (e.g., Johns and Hirt 1987; Weisman 1992, 1993). In either case, the location of interest for severe wind generation is typically at or near the leading edge of the gust front or bow-echo apex.

Rutledge et al. (1988) and Weisman (1992) discuss RIJs that descend to the surface 20–30 km rearward of the leading edge of the cold pool, and Johnson and Hamilton (1988) discuss strong winds produced in the TSR by wake lows. The former cases are analyzed as nonsevere, decaying systems, while the latter involves strong front-to-rear-directed winds responding to local pressure-gradient accelerations. While likely less common, it is conceivable that severe wind speeds may also be generated by RIJ processes in areas that extend well rearward of the leading edge of the gust front, even in mature, long-lived systems; this phenomenon is emphasized herein.

Most severe wind forecasting and nowcasting applications focus on predicting the processes that lead to intense, localized downdrafts formed along the leading edge of the gust front. These winds are often the product of the general “downburst” process—defined by Fujita (1978) as “a strong downdraft inducing an outward burst of damaging winds on or near the ground.” Many studies have further detailed this process by which convective downdrafts intersect the surface and diverge quickly outward, often producing an asymmetrical pattern of severe winds due to both shear- and pressure-gradient-induced accelerations (e.g., Fujita and Wakimoto 1981; Eilts and Doviak 1987; Vescio and Johnson 1992; Orf and Anderson 1999; Wakimoto 2001).

While much research has been devoted to downbursts and related phenomena (e.g., microbursts and macrobursts), understanding and predicting the descent of the RIJ and the potential for realizing severe surface winds in areas other than the bow-echo apex has received considerably less attention. Severe wind speeds in the TSR are examined here to assess the role that CMT may play in producing strong surface winds, particularly in “atypical” locations away from the gust front leading edge.

In light of the questions posed above, the objectives of this investigation are to (i) assess the sensitivity of the CMT process to environmental humidity, (ii) quantify and explain sensitivities in MCS motion as a result of changes in environmental humidity, and (iii) evaluate changes in frequency and intensity of strong surface wind speeds in an environment of reduced midlevel moisture.

2. Sensitivity simulation hypotheses and methods

The experimental design employed here bears some similarity to those of previous studies examining MCS sensitivity to environmental humidity and microphysical processes, such as the work of Fovell and Ogura (1988), YH95, and Grim et al. (2009). The Weather Research and Forecasting (WRF) model is used with idealized initial conditions as described in Mahoney et al. (2009). A control simulation is compared to a simulation with decreased midlevel relative humidity in order to test the sensitivity of the motion and surface winds of the simulated MCS to changes in environmental midlevel moisture and evaporation.
a. Control simulation

The initial conditions for the control simulation (CTRL) consist of a simple westerly jet stream, an environment with moderate CAPE (~2000 J kg\(^{-1}\)), and a horizontal wind field in thermal wind balance (see Mahoney et al. 2009, their Fig. 2). Convection is triggered by a 2°C warm bubble (with unchanged mixing ratio and horizontal dimensions ~100 km x 40 km) located between the surface and 3 km above ground level (AGL). The model domain uses 1-km horizontal grid spacing; other characteristics follow Mahoney et al. (2009) and are summarized in Table 1. Sensitivity simulations evaluating the effects of model grid spacing, and the parameterization of planetary boundary layer processes, cloud microphysics, and numerical diffusion were all performed to confirm the robustness of the model methodology and resulting MCS. Model simulation times are hereafter denoted by “forecast time” FHH or FHH:mm, specifying the number of hours (HH) and minutes (mm) into the simulation.

b. Dry midlevels (DRYM)

The “dry midlevels” simulation (DRYM) uses the same model details and parameterizations as the CTRL run, but the initial sounding features a reduction of moisture from the 750- to the 450-hPa level (Fig. 2) (and resulting in an increase of most-unstable, virtual-temperature-corrected CAPE by ~40 J kg\(^{-1}\) across the domain). The reduction in relative humidity follows the methodology of YH95, in which the initial environmental relative humidity at midlevels is reduced by approximately half of its original value at the driest part of the sounding, and is gradually blended back toward the initial values at the top and bottom of the layer.

Previous studies have noted differences in MCS speed due to altered microphysical processes. For example, Fovell and Ogura (1988) find that an increase in coldness and the depth of the subcloud cold pool due to increased

![Fig. 2. Comparison of initial sounding shape used to initialize the idealized MCS simulation (at 35° lat, -95° lon) for the (a) CTRL and (b) DRYM (dry midlevels) simulations. Shown are temperature (°C, solid line to right) and dewpoint (°C, solid line to left), with wind barbs in kt at right (1 kt = 0.5144 m s\(^{-1}\)).](image)
microphysical cooling resulted in a faster propagation speed; James et al. (2006) illustrate that relatively dry conditions at low and midlevels favor intense cold-air production and thus strong cold-pool development. Based on these and other previous studies, we hypothesize that MCS speed will increase in the DRYM simulation relative to CTRL due to enhanced evaporative cooling into dry air. An increase in ground speed is possible via 1) an increase of theoretical cold-pool speed resulting from colder cold-pool temperatures, 2) increased CMT via enhanced mid- to-low-level downward motion associated with stronger midlevel cooling, or 3) a combination of the above.

A counterhypothesis is that MCS speed may be unchanged or even decrease if drier midlevel air is entrained into the updraft and the system weakens overall. In nature, this could occur via entrainment, leading to decreased updraft buoyancy and weakened updraft strength. This in turn results in reduced condensate and ultimately a cold-pool intensity that is either unchanged or even weaker than in the CTRL simulation\(^1\) (e.g., James and Markowski 2010). Another possibility is described by YH95, in which the entrainment of drier air produces midlevel cooling that counteracts the vorticity tendency produced by the surface cold pool. This results in a more upright system, weakening the front-to-rear flow, and decreasing the size and intensity of the TSR. Less evaporation, melting, and sublimation in the TSR weaken the negative buoyancy and horizontal buoyancy gradients there, thereby weakening the rear inflow as well. Interestingly, the results of YH95 show that despite producing a weaker overall system, a decrease in the ground speed of the drier-midlevel MCS is not observed.

c. Additional sensitivity experiments

Similar to YH95, additional sensitivity simulations were performed to test the impacts of reduced (enhanced) evaporation and the removal (enhancement) of melting and sublimation on MCS motion and severe surface wind generation. While the additional microphysical process sensitivity simulations also revealed considerable differences from CTRL, the DRYM simulation and the impacts of changes in environmental moisture alone will remain the focus of this manuscript, with the remainder of the sensitivity experiments summarized in Mahoney (2009) and to be further analyzed in a separate paper.

\(^1\) Bryan et al. (2003) demonstrates that 1-km grid spacing may be too coarse to adequately represent updraft entrainment processes; this effect may be understated in the simulations discussed here.

3. Summary of sensitivity simulation results

a. CTRL

The MCS produced by the CTRL simulation is characterized by a quasi-linear, bowing shape (Fig. 3) with an intense leading convective line followed by a modest region (~100 km wide) of lighter precipitation. The system largely adheres to the “leading convective trailing-stratiform” MCS prototype (e.g., Newton 1950; Houze et al. 1990), and exhibits many of the dynamical features that have been found in past observational and modeling MCS studies. For example, vertical cross sections taken along the leading edge of the bow echo (and averaged every 40 km in an along-line direction to facilitate comparisons) show a robust midlevel mesolow, a clear depiction of a descending RIJ, and a steady surface-based cold pool that drives the system’s forward motion (Figs. 4a and 4c).

b. DRYM

1) MCS SPEED

The DRYM simulation exhibits a smaller system with reduced trailing stratiform precipitation, and a more eastward system motion relative to CTRL (Fig. 5). Calculations of MCS motion reveal a simulation-average ground speed for the DRYM MCS that is quite similar to that in CTRL; during the course of the simulations there are noticeable differences, but the average speed in DRYM is only ~1 m s\(^{-1}\) faster than that of the CTRL MCS when averaged over the duration of the simulations. A detailed analysis of MCS speed for each simulation was performed following the methodology of Mahoney et al. (2009), in which theoretical cold-pool speeds \(c\) are compared with momentum budget results in order to quantify the role of cold-pool intensity and CMT in determining MCS motion.

In the interest of focusing this manuscript on the severe surface wind speed aspect (where the largest sensitivity appears to be manifest), the details of these comparisons are omitted here, but the main results are summarized: calculations of surface cold-pool intensity and theoretical cold-pool speed show that the initial period of MCS acceleration in DRYM is actually marked by slight decreases in both of these quantities. This decrease in cold-pool intensity but increase in MCS speed also occurs during a maximum of CMT over the TSR, as well as an intensification of low-level winds. Thus, despite the subtle quantitative difference in observed ground speed, it is suggested that CMT during this stage of MCS development is likely an important contributing process to system motion during a period not completely determined by cold-pool dynamics.
That a more systematic and marked difference in MCS ground speed over the course of the simulation is not found between the simulations is at first somewhat surprising within the context of previous studies that suggest strong sensitivity to the processes affected here (e.g., Fovell and Ogura 1988; Braun and Houze 1995; YH95; Torres Brizuela and Nicolini 2008). However, as hypothesized above, the potential for system weakening due to midlevel drying indeed appears to have occurred here (e.g., weaker updrafts, pressure perturbations, and cold pool strengths in a spatially smaller system located farther to the north; see Figs. 4 and 5); this is also similar to the results found in the dry simulation of YH95.

Comparing the ground speeds, theoretical density current speeds, and momentum budgets of CTRL and DRYM illustrates that it is difficult, if not impossible, to truly separate the contribution of “cold-pool dynamics” (i.e., density current propagation) from “CMT forced” MCS motion. The two fields are fundamentally linked by both of the rear-to-front and front-to-rear storm-scale flow branches, and the connection between the RIJ and surface cold pool is itself a primary driver of MCS motion. As most past studies have employed idealized models or focused exclusively on the role of various microphysical processes with respect to a certain feature (e.g., rear inflow, TSR extent, heating profiles, etc.), it is possible that MCS motion and/or CMT simply does not vary as strongly as a result of these changes in this particular environment or simulation setup. One can envisage certain environments where a more pronounced effect might be realized (e.g., weaker background flow, a drier RIJ environment, an enhanced environmental low-level jet, higher CAPE, etc.); this spectrum of possibilities is beyond the scope of this investigation but presents avenues of potential future work.

2) THERMODYNAMIC AND KINEMATIC COMPARISONS

Comparing vertical cross sections along the leading edge of the bow echo and averaged as above shows a slightly deeper \((\Delta z \approx 0.5-1 \text{ km})\) but less intense surface cold pool \((\Delta T' \approx 1^\circ \text{C})\) and a slightly weaker \((\Delta p' \approx 1-2 \text{ hPa})\) midlevel mesolow in DRYM (Fig. 4). The downward component of motion in the main region of RIJ descent is slightly greater in DRYM (Figs. 4d and 6c,d), which is consistent with decreased buoyancy there as well (Figs. 6a and 6b). Contoured frequency by altitude diagrams (CFADs; Yuter and Houze 1995) also
offer insights into differences in the TSR of CTRL and DRYM. Examination of the stratiform area\(^2\) beyond the convective line at F09 (when each system is in its mature stage) shows an increase in downward motion in DRYM (Figs. 7c–e), consistent with the above analysis of along-line-averaged vertical cross sections. The signal found in the microphysical cooling field over the stratiform area is mixed; larger magnitudes of midlevel (4–6 km AGL) cooling are found in the DRYM simulation (the same layer in which drier initial air is maximized), as well as near the surface (0–1 km AGL). However, areas of moderate-to-strong midlevel cooling are also found in CTRL between 0 and 4 km (Figs. 8a–c), likely attributable to increased stratiform precipitation in CTRL relative to DRYM within this layer (Figs. 3 and 5). Slightly stronger midlevel rear inflow in the 3–6-km layer is also noted in DRYM, as are enhanced near-surface wind speeds (Figs. 8d–f). Vertical momentum flux and flux convergence values of both the mean and perturbation wind are enhanced in DRYM at mid- to low levels. These differences are well illustrated by CFADs comparing CMT processes (Fig. 9), in which the downward momentum transport of both the environmental wind and the perturbation wind are both larger in DRYM.

As a comparison to the DRYM simulation, a moist midlevels simulation (MOIM) was also performed, with an increase in relative humidity by 25% centered at the 600-hPa level. Results show an increase in convective intensity, as well as enhanced MCS speed and TSR extent (Fig. 10). Reflecting the increase in convective intensity

\(^2\) Stratiform grid cells were separated from convective grid cells according to the following definitions: (i) convective grid cell, \(w \approx 2 \text{ m s}^{-1}\) at 2 km AGL or composite reflectivity >30 dBZ, using a 5-km radius around each “convective” point; and (ii) stratiform grid cell, all those cells that are not convective but still have composite reflectivity >15 dBZ.

**Fig. 4.** (a) East–west along-line-averaged (in 40-km segments) cross section along line CTRL_A-CTRL_B, as in Fig. 3c at F09. Cold pool (shaded as shown beginning at \(T' = -2^\circ\text{C}\)), solid (dashed) contours represent negative (positive) pressure perturbations in hPa, and simulated reflectivity (dBZ, dotted contour, 20 dBZ); (b) as in (a), but for the DRYM simulation taken along DRYM_A-DRYM_B, as in Fig. 5c; (c) as in (a), but here black contours display the magnitude of the total perturbation wind \([u^2 + v^2]^{1/2}, \text{m s}^{-1}\) and black arrows show the ground-relative perturbation flow in the \(x-z\) plane scaled as shown by the reference vector in the bottom-left corner, simulated reflectivity (dBZ, dotted contour, 20 dBZ); and (d) as in (c), but for the DRYM simulation taken along DRYM_A-DRYM_B as in Fig. 5c. Horizontal distance of cross sections is 175 km.
is an enhancement relative to CTRL of strong surface winds at the leading edge of the gust front. To focus on the role of the dry air at midlevels, the MOIM simulation is not analyzed further, but comparing system organization and motion alone offers some insight into what a larger spectrum of moisture sensitivity experiments might reveal.

4. Surface wind speed comparisons

Though the focus in this section is on strong surface winds in the TSR, differences between simulations also emerge in the surface wind field along the leading edge of the gust front. Additional research has revealed that mesovortices (e.g., Weisman and Trapp 2003) found along the gust fronts of each simulation (not shown) are highly correlated with severe surface winds, a process discussed in several recent mesovortex studies (e.g., Trapp and Weisman 2003; Wakimoto et al. 2006; Atkins and St. Laurent 2009). This specific process, the differences found between simulations, and possible CMT-based forecast implications are not discussed here but can be found in Mahoney (2009). Here, we use the same simulations as above to examine severe winds in the TSR within the framework of CMT and discuss potential implications for severe wind forecasting.

a. Severe surface wind production

The National Weather Service (NWS) defines severe convective winds to be those measured at 10 m AGL in excess of 50 kt (25.7 m s\(^{-1}\)).\(^3\) An estimate of severe surface wind speed occurrence for each simulation is obtained by counting the total number of grid cells with 10-m wind speed values >25.7 m s\(^{-1}\) and normalizing this number by the storm area (defined as the number of grid cells in which simulated composite reflectivity exceeds 30 dB\(Z\)). The area-normalized severe wind occurrence shows that the DRYM MCS generates more than 4 times the normalized severe surface wind coverage produced by that in CTRL (when

3 The Glossary of Meteorology (Glickman 2000) notes that the duration of a gust is usually less than 20 s; by contrast, “sustained winds” are defined as an average over a 2-min period. The model time step used in each simulation is 12 s, and model output is only produced every 5 min. Therefore, the discussion of “severe surface winds” presented here is not intended to replicate observations of severe surface winds had these systems occurred in nature. Instead, grid-cell values of 10-m wind speed are used for relative comparisons between model simulations only; despite these approximations and the fact that other surface wind magnitude thresholds are also used in the following analysis, the “severe” terminology remains reserved for the 50-kt (25.7 m s\(^{-1}\)) NWS criterion noted.
storm size is not taken into account, DRYM still produces 2.5 times the number of severe winds occurrences as CTRL. Severe surface winds between the CTRL and DRYM simulations differ in spatial distribution, intensity, and generation mechanism (Fig. 11) as discussed below.

b. CMT and severe surface winds in the TSR

1) TSR SURFACE WINDS IN CTRL AND DRYM

Severe surface winds in both the CTRL and DRYM simulations are found along the leading edge of the gust front and near gust front mesovortices; strong surface winds are generally expected in these areas. However, occurrences of severe or nearly severe surface winds are also generated in locations that trail the gust front by ~50 km, areas in which severe winds are not typically anticipated (e.g., Figs. 11a and 11b). These areas of strong surface winds in the TSR also tend to display more persistence in time (generally on the order of 30 min to 1 h or longer), existing as continuous entities as opposed to isolated, short-lived microbursts. What specific process explains these areas of strong surface winds in the TSR? To varying degrees, both simulations produce severe winds across broad areas that trail the system’s leading edge, though the occurrences are more common in DRYM (Fig. 11b). Thus, the MCS in DRYM produces a greater areal coverage of severe surface winds relative to that in CTRL, and also reveals an increase in severe wind occurrences in locations rearward of the leading edge of the gust front.

The magnitude and spatial distribution of surface winds in CTRL and DRYM simulations are compared at F06 and F09, respectively (Figs. 11e and 11f), to illustrate the storm-relative locations of the strongest winds at the time of peak wind production (Fig. 11c). At these times shown, the areas of strongest winds are found along both the leading edge of the system, and rearward from it, with the DRYM simulation revealing larger areas of stronger winds (>32 m s⁻¹). The tendency for a smaller system to produce larger areas of extreme winds is noted by Wheatley et al. (2006), and the potential correlation...
between storm size and areas of strong winds associated with gust front mesovortices (e.g., Trapp and Weisman 2003) are areas of potential future research.

Vertical cross sections taken through the leading edge of the gust front at F06 for CTRL and F0730 for DRYM illustrate two examples of TSR severe wind occurrences that each appear as couplets, with one area of severe winds along the gust front leading edge, and one trailing by up to ~50 km (Fig. 12). These two severe wind episodes were selected as representative of a larger set (e.g., Fig. 11) of similar double-maxima occurrences throughout the simulations.

Cross sections reveal that the strong surface winds found 25–50 km to the rear of the gust front appear to be generated via a different process than those closer to the leading edge. Examination of areas of descent in each simulation indicates differing intensities and spatial concentrations of downdrafts (Figs. 13a and 13c); time series of these vertical cross sections demonstrate the temporal persistence of these varied downdraft types (not shown).

Descent behind the leading edge of the gust front is characterized by low-level (0–3 km AGL) concentrated, intense downdrafts (points A/B and D/E in Figs. 13a and 13c), trailed by a more diffuse area of somewhat weaker and slightly more elevated (1–5 km AGL) downdrafts. The trailing areas of descent in the CTRL simulation have magnitudes generally ~0.5–2 m s\(^{-1}\), but those in DRYM are substantially stronger (downdraft magnitude up to 7 m s\(^{-1}\)); points C/F in Figs. 13a and 13c, respectively). At the times shown, the surface winds that exceed the severe threshold are actually those associated with the weaker descent toward the TSR rather than stronger, more concentrated downdrafts at the leading edge of the cold pool. This suggests that strong downdrafts themselves may not always be necessary in order to produce strong surface winds in the TSR, as the strong surface winds realized here likely result more from the accelerated horizontal momentum of the RIJ being transported to the surface.

2) CMT AND SEVERE SURFACE WIND GENERATION

As an overall enhancement in downward motion is expected due to increased evaporative cooling in the TSR in the DRYM simulation, it is not necessarily surprising that DRYM produces more severe winds overall
(both total and per storm area), and that this increase also extends to TSR winds. However, the degree of difference and the physical processes used to explain such differences implicate CMT as a driving factor. It is thus proposed that the analysis of severe surface winds via CMT may be relevant to better understanding severe wind generation. To evaluate this mechanism, momentum budgets for CTRL and DRYM are performed following the analysis of Mahoney et al. (2009), in addition to a systematic evaluation of series of horizontal and vertical cross sections, examined both instantaneously and via temporal and spatial averaging.

While the winds at the leading edge of the gust front are accompanied by concentrated, intense downdrafts close to the leading edge [fitting the classic downburst definition offered by Fujita (1978); see Figs. 13a and 13c], strong surface winds located rearward appear to result mostly from the downward advection of the accelerated horizontal momentum in the RIJ (i.e., CMT[$u' + v'$]; see Figs. 13b and 13d). The role of momentum transport in the production of RIJ-driven severe surface winds is well illustrated by both CMT[$\overline{u}$] and CMT[$u' + v'$]. The transport of the storm-induced rear inflow is illustrated by the maximum of CMT[$u' + v'$] that persists just above the area of strong and severe surface winds trailing the cold pool’s leading edge (Figs. 13b and 13d). The magnitude of this feature is quite large, generally exceeding 20–40 m s$^{-1}$ h$^{-1}$ at 1–2 km AGL in the CTRL simulation. While CMT[$\overline{u}$] in CTRL is generally 2–4 times weaker than CMT[$u' + v'$], the process does maximize (~10–15 m s$^{-1}$ h$^{-1}$) just above the area of severe surface winds as well, demonstrating the contribution from the downward transport of the ambient flow to severe wind generation at the ground. The connection between the environmental (background) wind and severe surface winds has been discussed by Foster (1958), Johns and Doswell (1992), Evans and Doswell (2001), Geerts (2001), and Kuchera and Parker (2006), while CMT of the perturbation wind has received less attention. Comparison of CMT maxima between CTRL and DRYM reveals that the transport of horizontal momentum is stronger by up to 100% in DRYM relative to CTRL; this is true for both CMT[$\overline{u}$], and CMT[$u' + v'$] (Figs. 9 and 13).

The CFADs presented in the previous section also provide valuable insights into the differences in low-level wind speeds. As expected, within the layer that relative humidity is decreased in DRYM (~4–6 km AGL), an increase in larger magnitudes of microphysical cooling is evident. Enhanced downward motion in DRYM is also observed, particularly at low levels (Fig. 7). The increase in downward motion implies systematic increases in CMT over this region as well, assuming the horizontal momentum fields remain comparable. Figures 8d–f shows
subtle enhancement of the near-surface wind magnitude (~0–1 km AGL) in DRYM relative to CTRL; this is consistent with a secondary maximum of microphysical cooling observed there, as well as enhanced RIJ descent in the DRYM simulation. The integrated effects of these differences beget the significant increase in both \( CMT[\vec{u'} + \vec{v'}] \) and \( CMT(\bar{u}) \) in DRYM relative to CTRL over this region (Fig. 9).

Figure 14 provides a comparison of schematics conceptualizing the processes described here. Enhanced descent across the middle to low levels in the DRYM simulation preferentially provides an additional source of downward motion to focus and lower the strong perturbation flow from aloft. Thus, when the RIJ descends rearward of the gust front leading edge (and across a smaller horizontal distance as in DRYM), CMT maximizes in a more localized area; given an otherwise favorable MCS and surrounding environment, this process may contribute to an additional area of strong winds in locations behind the gust front. Dry midlevel environments are not the only situations in which these areas of trailing strong winds may be found. Indeed, such occurrences are found in the CTRL simulation as well, and past work demonstrates that RIJs may descend in MCSs across a range of environments for a variety of reasons.

Weisman (1992) in particular discusses the role of CAPE–shear balance in RIJ descent, and suggests that system “optimality” (Rotunno et al. 1988) is a primary consideration in this process. Weisman (1993) further highlights the dependence of both severe surface winds and RIJ descent in idealized bow-echo systems on magnitudes of low-level shear and CAPE ahead of the system. Fully idealized studies (e.g., Weisman 1992, 1993) provide an important complement to a study such as this one, where (i) changes in humidity have introduced slight changes in CAPE and (ii) the system itself forms in different geographical locations and thus experiences slight differences in CAPE and wind shear due to thermodynamic gradients in the initial environment [while our initial state (e.g., inclusion of the Coriolis force, a midlatitude jet stream, and thermal wind balance) more fully represents a real-world midlatitude environment]. Therefore, while many considerations exist regarding what ultimately forces an RIJ to descend in a particular storm, the comparison of the CTRL and DRYM simulations here highlights a lesser-considered mechanism for severe surface winds that appears to be augmented by microphysically enhanced subsidence. The connection of severe wind type and frequency to the background environment implies potential for improving future forecast tools or indices.

While these results show that increased CMT via microphysical cooling-enhanced downward motion in the TSR may increase TSR severe wind speeds, the relationship is complex and requires further study. For example,
the stability of the surface cold pool\(^4\) and the effects of dry midlevels on the intensity of the overall system are important. Here, the combination of enhanced TSR descent, increased low-level flow from CMT, and a slightly weaker and horizontally smaller surface cold pool (potentially facilitating mixing of momentum to the surface) appears to account for the two- to fourfold increase in severe surface winds in the DRYM simulation, particularly those occurring in or toward the TSR. That RIJ descent accounts for these instances of severe surface winds suggests a possible addition to existing common conceptual forecast models of RIJ-driven surface winds in which descended rear inflow is generally considered to be (i) part of a “derecho” type of event (e.g., Johns and Hirt 1987), (ii) associated with gust front mesovortices (e.g., Trapp and Weisman 2003), or (iii) the result of a decaying system producing more diffuse and widespread surface wind swaths of less severe intensity (e.g., Weisman 1992).

A large number of previous studies as well as most current surface wind gust forecast indices acknowledge the potential for evaporative cooling to produce or strengthen downdrafts that may penetrate the surface and lead to wind damage (e.g., Wakimoto 2001; Kuchera and Parker 2006; Cohen et al. 2007). While the results of these simple sensitivity experiments certainly corroborate that basic principle, the analysis of DRYM and CTRL in the preceding sections suggests that the sensitivity to evaporative potential may be strongly linked to increased CMT in the descending RIJ, as well as to thermodynamics (e.g., downburst production by cooling due to microphysical processes).

5. Summary and forecast implications

a. Summary

A comparison of a pair of three-dimensional, idealized MCS simulations demonstrates the impacts of ambient

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\(^4\) Findings and discussions by Weisman (1992), Bryan and Weisman (2006), Kuchera and Parker (2006), Parker (2008), and James and Markowski (2010) have shown that the relationship between surface cold-pool stability and severe surface winds remains somewhat ambiguous.
humidity on CMT, MCS motion, and the generation of severe surface winds. Drier air at midlevels slightly increases the average MCS speed by $1 \text{ m s}^{-1}$ despite producing a smaller and somewhat weaker system. While the simulation with drier midlevels was at first hypothesized to demonstrate a significant increase in ground speed, a decrease in storm intensity and an altered storm track relative to CTRL preclude a perfect separation analysis of processes contributing to storm motion. While the initial hypothesis proved too simplistic for these particular simulations, it is possible that MCS motion may demonstrate increased sensitivity to midlevel drying in other environments (e.g., an environment with increased or spatially homogeneous CAPE). In the pair of simulations reported upon here, drier air at midlevels is found to increase downward motion within the TSR and both enhance and focus areas of CMT associated with RIJ descent. Analyses of lower-tropospheric wind speeds illustrate the role of CMT in generating strong or severe surface winds. It is shown

![Figure 11](image.png)

**Fig. 11.** (a) CTRL cold pool ($T = -2^\circ\text{C}$, light and dark blue for times as labeled, areas of grid cell 10-m wind magnitude $>25.7 \text{ m s}^{-1}$ (red, shaded). (b) As in (a), but for DRYM. (c) Number of grid points with 10-m wind magnitude $>25.7 \text{ m s}^{-1}$ normalized by storm area (number of grid cells with simulated composite reflectivity $>30 \text{ dBZ}$) vs time (h) along the x axis. (d) Total number of grid points with 10-m wind magnitude $>25.7 \text{ m s}^{-1}$ normalized by the storm area for the entire simulation; simulations as labeled. (e) CTRL simulation of the 10-m winds (m s$^{-1}$, shaded as indicated at right). (f) As in (e), but for DRYM at F09.
that the additional downward motion associated with decreased midlevel relative humidity strengthens and focuses CMT and contributes to an increase in the number of grid-cell occurrences of severe surface winds in the TSR region.

Mahoney et al. (2009) demonstrate that CMT is a significant driver of the MCS low-level momentum field, and that this process impacts both MCS motion and the low-level wind field. An important finding in the current study is that significant changes in the surface wind field

Fig. 12. (a) CTRL cold-pool boundary ($T' = -2^\circ$C, contours); areas of grid-cell 10-m wind magnitude $>25.7$ m s$^{-1}$ (shaded) for times F0600-F0640, where colors denote different times as labeled in top right. (b) As in (a), but for DRYM from F0700 to F0740 as labeled in the top right. Dotted lines denote cross sections shown in Fig. 10.

Fig. 13. (a) Total wind magnitude (m s$^{-1}$, shaded as indicated), cold pool ($T' = -2^\circ$C, blue dotted contour), $w < 0$ (solid black contours, 0.5, $-1$, $-2$, $-3$, $-4$, and $-5$ m s$^{-1}$), and simulated reflectivity (dBZ, red dashed contour, 20 dBZ) at F06 along line CTRL_C-CTRL_D in Fig. 12a. Here, A, B, and C denote specific downdraft locations discussed in the text. (b) As in (a), but here the contours show $[\text{CMT} (\mathbf{u'} + \mathbf{v})]$ (m s$^{-1}$ h$^{-1}$, solid contours starting 1 m s$^{-1}$ h$^{-1}$), CMT(\theta) (m s$^{-1}$ h$^{-1}$, dashed contours starting at 1 m s$^{-1}$ h$^{-1}$), and simulated reflectivity (dBZ, dashed contour, 20 dBZ). (c) As in (a), but for DRYM at F0735 along DRYM_C-DRYM_D in Fig. 12b. (d) As in (b), but for DRYM at F0735.
can be evident in response to decreased midlevel humidity, in both the number and spatial distribution of severe wind occurrences, as well as the distinct processes (e.g., CMT) by which such surface winds are generated. This underscores that, to the extent that severe surface winds are due to enhanced CMT in regions of accelerated storm-scale flow, optimum conditions for severe surface wind production are related to thermodynamic as well as dynamical processes (and forecast tools may be optimized by incorporating both). A more complete exploration of the parameter space that maximizes low-level CMT is required in order to elucidate conditions that also maximize CMT-driven severe surface winds.

b. Forecast implications and avenues for future work

Wakimoto (2001) emphasizes that forecasting thunderstorm outflow wind speeds remains a critical but unresolved problem. While the contribution of CMT does not fully explain such a complex issue, this process is important to severe surface wind generation. That a descending RIJ can cause strong surface winds is not novel, but most studies limit the discussion of the potential for RIJ-driven surface winds to diffuse areas of less intense winds, concentrated at the gust front leading edge or the apex of a bow echo. The distinction of the two areas of strong surface winds and the different generation mechanisms found here warrant further study, and may be relevant to forecasting or nowcasting the potential for, and intensity and duration of, severe surface winds. Of particular forecasting significance is that the enhanced gust front winds found in MOIM demonstrate that we do not yet fully understand which additional environmental characteristics control whether dry air diminishes or enhances severe surface winds. Future investigation of this topic will be highly worthwhile.

Studies such as those of Johns and Doswell (1992), Weisman (1992), YH95, Evans and Doswell (2001), and Geerts (2001) have commented on the complexity of observed RIJ dynamics and their related impacts on convective wind gust generation. Geerts (2001) in particular examines the forecast utility of incorporating a simplified representation of the CMT of ambient winds [i.e., CMT(\(\overline{u}\))] into an existing thermodynamically based surface wind gust prediction index. While the overall results show an improvement in the index with CMT over the traditional thermodynamically driven index, many of the strong wind events remained underpredicted. These results corroborate those of Mahoney et al. (2009), who show the transport of ambient momentum to be of secondary significance relative to that of the storm-induced perturbation flow. In the wind events analyzed here as well, CMT(\(\overline{u}\)) is smaller than CMT of the perturbation flow (by at least a factor of 2), underscoring that a more complete integration of the suite of CMT physical processes (i.e., pressure gradient acceleration of midlevel winds and the vertical advection of both the ambient and storm-induced winds) into the forecast framework may be critical toward improving the underprediction of the strongest wind events. However, current operational model forecasts of ambient wind speeds (\(\overline{u}\)) are better resolved than a storm’s perturbation wind field (\(u'\), \(v'\)); if trying to incorporate this process into a forecast technique, this [i.e., representing (\(u'\), \(v'\))] becomes a significant consideration.

Extending beyond the initial CTRL versus DRYM comparison presented here, observational evidence of these areas of trailing winds in the TSR must be sought. Second, a better understanding of the effects of dry midlevels in MCS simulations (e.g., reduction of updraft buoyancy, changes in updraft tilt, reduction of hydrometeor concentrations, and thus cold-pool intensity) is
desirable, and will necessitate a larger suite of simulations to accomplish. The ultimate demonstration of forecast utility will also require further investigation of observed and simulated MCSs across a greater variety of environments (e.g., James and Markowski 2010). Yet another question piqued by these results is the sensitivity of MCS size to environmental moisture (and the forecast implications of corresponding changes in areal coverage of severe wind potential).

Finally, the sensitivity results discussed herein are largely dependent upon model representations of the TSR, as the processes inherent in this storm region are known to have considerable impacts on the both downward motion throughout an MCS, as well as the strength and location of a the RIJ. However, many studies note the underrepresentation of the TSR in numerically simulated MCSs, and even with the inclusion of sophisticated ice microphysics parameterizations, models remain challenged to successfully represent both the transition zone and TSR extent in MCSs (e.g., Gallus and Pfeifer 2008; Morrison et al. 2009). Therefore, models may be insufficient to describe the total influence of CMT, particularly in light of recent observational studies illustrating the initiation of downward motion at the far rear of the TSR (e.g., Smith et al. 2009; Grim et al. 2009). To address these weaknesses, high-resolution observational data should be used alongside numerical simulations of observed and idealized cases to better understand the phenomena and critical storm-scale processes examined in this work.

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REFERENCES


 Torres Brizuela, M., and M. Nicolini, 2008: Sensitivity of main downdraft features to vertical wind shear and ice-phase in a microburst-producing thunderstorm. Atmosfera, 21, 57–82.