1. Introduction

Warm-season, progressive derecho-producing mesoscale convective systems (MCSs; e.g., Johns and Hirt 1987; Coniglio and Stensrud 2004; Guastini and Bosart 2016; Corfidi et al. 2016) continue to be a challenge to operational forecasters, despite the advances in research on these systems (Coniglio et al. 2012; Evans et al. 2014; Campbell et al. 2017, among others). Derecho-producing MCSs generally present distinctive mesoscale features such as a bowing convective line, intense cold pool, a mesohigh and a mesolow, which were associated with greater elevated thermodynamic instability and lower level of free convection in the successful member. The subsequent organization of a dominant bowing MCS was well predicted by the model that had more widespread convection in the early stages and no detrimental interaction with other simulated convective systems. Last, the inability of MPAS ensemble members to predict the MCS maintenance over western and middle Tennessee was linked to a dry bias in low levels and much lower thermodynamic instability ahead of the MCS compared to observations. This case demonstrates the challenges in operational forecasting of warm-season derecho-producing progressive MCSs, particularly when ensemble numerical weather prediction guidance solutions differ considerably.

KEYWORDS: Ensembles; Mesoscale forecasting; Numerical weather prediction/forecasting; Operational forecasting; Short-range prediction; Model errors

ABSTRACT: This study analyzes the low short-range predictability of the 3 May 2020 derecho using a 40-member convection-allowing Model for Prediction Across Scales (MPAS) ensemble. Elevated storms formed in south-central Kansas late at night and evolved into a progressive mesoscale convective system (MCS) during the morning while moving across southern Missouri and northern Arkansas, and affected western and middle Tennessee and southern Kentucky in the afternoon. The convective initiation (CI) in south-central Kansas, the organization of a dominant bow echo MCS, and the MCS maintenance over Tennessee were identified as the three main predictability issues. These issues were explored using three MPAS ensemble members, observations, and the Rapid Refresh analyses. The MPAS members were classified as successful or unsuccessful with regard to each predictability issue. CI in south-central Kansas was sensitive to the temperature and dewpoint profiles in low levels, which were associated with greater elevated thermodynamic instability and lower level of free convection in the successful member. The subsequent organization of a dominant bowing MCS was well predicted by the member that had more widespread convection in the early stages and no detrimental interaction with other simulated convective systems. Last, the inability of MPAS ensemble members to predict the MCS maintenance over western and middle Tennessee was linked to a dry bias in low levels and much lower thermodynamic instability ahead of the MCS compared to observations. This case demonstrates the challenges in operational forecasting of warm-season derecho-producing progressive MCSs, particularly when ensemble numerical weather prediction guidance solutions differ considerably.

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1. Introduction

Warm-season, progressive derecho-producing mesoscale convective systems (MCSs; e.g., Johns and Hirt 1987; Coniglio and Stensrud 2004; Guastini and Bosart 2016; Corfidi et al. 2016) continue to be a challenge to operational forecasters, despite the advances in research on these systems (Coniglio et al. 2012; Evans et al. 2014; Campbell et al. 2017, among others). Derecho-producing MCSs generally present distinctive mesoscale features such as a bowing convective line, intense cold pool, a mesohigh–wake low couplet and a descending rear inflow jet (Weisman 1992; Wakimoto et al. 2006; Evans et al. 2014). The evolution of these systems can be sensitive to the timing, location and intensity of the initial convection and its subsequent evolution (e.g., Lawson and Gallus 2016), which can make them particularly difficult to forecast. One example is the 3 May 2020 derecho that affected southeast Kansas, southern Missouri, northern Arkansas, southern Kentucky and Tennessee (Fig. 1). This study uses a convection-allowing ensemble (CAE) with the Model for Prediction Across Scales (MPAS; Skamarock et al. 2012) to evaluate some relevant physical processes associated with the low predictability in this case.

There were several important predictability issues with regard to the 3 May 2020 event. These include: the formation of the initial elevated convection north of a surface front, the transition to surface-based convection and its organization into a bow echo, and its subsequent propagation across the frontal boundary and toward the warm sector over western and middle Tennessee, which were not well predicted by convection-allowing model (CAM) guidance and operational forecasts. The National Weather Service’s Storm Prediction Center (SPC) Convective Outlook issued nearly 4 h before convection initiated in south-central Kansas did not anticipate severe storms in southeast Kansas and southwest Missouri. A higher risk of severe storms was introduced in subsequent Convective Outlooks in the area ahead of the evolving MCS, particularly after the bow echo signature was well characterized. Additionally, only 1 in 10 members of the National Centers for Environmental Prediction’s (NCEP) High Resolution Ensemble Forecast (Roberts et al. 2019, 2020; NOAA/NWS/SPC 2020a) initialized at 0000 UTC 3 May 2020 depicted a bow echo MCS in simulated reflectivity, suggesting a low probability of occurrence of this event. An investigation into the physical processes associated with the low operational predictability of this derecho is desirable.

Research has led to the development and utilization of forecasting tools such as CAMs and CAEs in the last two decades (Xue et al. 2007; Weisman et al. 2008; Schwartz et al. 2009; Romine et al. 2014; Sobash et al. 2016; Clark et al. 2018; Schwartz 2019; Johnson and Wang 2020, among others). CAMs are important tools in today’s severe thunderstorm forecasting process because of their ability to explicitly predict storm mode, intensity, motion, and evolution (Weisman et al. 2008), which are not resolved by coarser resolution models incorporating parameterized convection. However, details in the mesoscale and convective scale, such as the exact location...
of a specific storm or the number of storms in a given area, can vary substantially in different CAM solutions in response to dynamic core, initial conditions (ICs), lateral boundary conditions, and physical parameterizations employed (e.g., Stensrud et al. 2000; Schwartz et al. 2010; Lawson and Gallus 2016). In this context, CAEs offer a range of potential outcomes that can be used by forecasters to assess the relative likelihood of specific solutions, and creation of probabilistic guidance that helps to assess uncertainty.

One of the first studies using a global MPAS CAE (Schwartz 2019) used a mesh with cell spacing ∼15 km over most of the globe decreasing to ∼3 km over the conterminous United States, with 10 members initialized using the NCEP Global Ensemble Forecast System (GEFS) analyses and run for 132 h (5 days). Even though the focus was on the evaluation of medium-range (3–5 days) precipitation forecasts, the MPAS ensemble guidance proved useful for short-range (1–2 days) forecasts as well. Gallo et al. (2017) also indicate MPAS forecasts with convection-allowing resolution can be useful in forecasting severe convection. As computational capability increases and global CAMs become operationally viable, new studies using models with unstructured meshes, such as MPAS, should prove useful (Ha et al. 2017; Judt 2018).

We examine aspects of the sensitivity of MPAS CAE convective solutions to differences in ICs on the early development of convection and the subsequent organization and evolution of the 3 May 2020 derecho. In particular, we explore if differences in the ICs and in their associated physical processes led to different outcomes within the ensemble envelope. The assessment of physical processes leading to different member outcomes provides insight into factors contributing to ensemble spread, and this may improve our understanding of low-probability but potentially high-impact severe weather events. A 40-member, convection-allowing MPAS ensemble (hereafter “MPAS ensemble”) is used to investigate these sensitivities.

Section 2 describes the methodology and datasets used in this study, including the MPAS ensemble configuration. Section 3 presents the results, from the general ensemble performance in this case to comparisons of different ensemble members. Section 4 discusses the conclusions of the study.

2. Data and methodology

a. MPAS ensemble configuration

The ensemble forecasts for the 3 May 2020 event use the MPAS version 7.0 and a global mesh with variable horizontal resolution. The cell spacing is ∼60 km between points over most of the globe decreasing incrementally to ∼3 km spacing over a mesoscale domain covering the derecho region centered in southern Missouri at 37°N, 92°W (Fig. 2). The area with mesh spacing ∼3 km covers the area of convective initiation (CI), the entire track of the bow echo MCS, and the severe weather reports swath (Fig. 1). The model has 40 vertical levels extending from near the surface up to 30 km. Following Ha et al. (2017) and Schwartz (2019), a time step of 18 s is used in the MPAS integration. Convection is parameterized using the “scale-aware” Grell–Freitas scheme (Grell and Freitas 2005).
so that simulated convection within most of the area with mesh spacing < 4 km is explicitly resolved (Fowler et al. 2016; Schwartz 2019). All ensemble members use the same dynamic core, mesh configurations and physical parameterizations (Table 1).

The ICs are the only source of diversity in the MPAS ensemble and are provided by the GEFS analyses. The GEFS version valid at the time of the event (version 11.0) is a TL574 (∼34-km horizontal grid spacing at the equator) spectral model with 64 hybrid vertical levels (Zhou et al. 2017), and the MPAS preprocessing system interpolates the GEFS ICs to the MPAS mesh. GEFS has 20 perturbed members and one control member and is initialized at the four synoptic times (0000, 0600, 1200, and 1800 UTC) each day. The perturbations in the GEFS ICs are generated from the ensemble Kalman filter component of NCEP’s Global Data Assimilation System (Whitaker and Hamill 2002). In this study, we use the operational version of the GEFS available on the day before the event (NCEI 2020).

The MPAS ensemble employed here has 40 total members initialized on 2 May 2020 (the day before the derecho). Half of these members are initialized with 20 perturbed GEFS 1800 UTC 2 May 2020 analyses (members 1–20) and the other half are initialized with 20 perturbed GEFS 1200 UTC 2 May 2020 analyses (members 21–40). The 40-member MPAS ensemble using these GEFS ICs was able to generate both somewhat successful and unsuccessful forecasts of the event, which allows a comparison of the different scenarios (Trier et al. 2019). This time-lagged approach (members 21–40 initialized 6 h before) increased the spread of the ICs and forecasts compared to using GEFS ICs from a single time.

The first tests using the MPAS ensemble were initialized at 0000 UTC 3 May (the day of the derecho) using the GEFS analyses at the same time. However, spurious convection propagation from Colorado to Kansas between 0000 and 0900 UTC 3 May potentially affected new CI in south-central Kansas just after 0900 UTC 3 May. The 1200 and 1800 UTC 2 May 2020 times were subsequently selected to initialize the MPAS ensemble at earlier times in order to avoid potential influences of the MPAS “cold-start” spinup of the upstream convection during the first 6–12 h of integration (Schwartz 2019).

b. Other datasets

The Multi-Radar Multi-Sensor (MRMS; Zhang et al. 2011) data are used to access the observed composite reflectivity. Severe weather reports were obtained from the SPC’s Storm Data (NOAA/NWS/SPC 2020b) archive. The NCEP’s Rapid Refresh (RAP; Benjamin et al. 2016) analyses are used for the synoptic-scale overview of the case. The surface analysis and observation maps were obtained from the Weather Prediction

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2 The GEFS was upgraded to version 12.0 on 23 September 2020.
Center (WPC; NOAA/NWS/WPC 2021) archive and observed soundings from the University of Wyoming (2020) archive.

3. Results

a. Overview of the event and predictability issues

Figure 1 shows the radar evolution of the bow echo MCS and the SPC’s severe storm reports from 1000 to 2300 UTC 3 May 2020. Elevated thunderstorms formed just after 0900 UTC in south-central Kansas and produced severe hail just after 1000 UTC and the first severe wind report at 1140 UTC. After 1200 UTC, the storms grew upscale into a linear MCS that tracked east-southeastward over the next several hours producing significant damaging wind gusts in southeast Kansas. The convective line acquired a bow shape in southwest Missouri around 1500 UTC and soon after developed a northern line-end cyclonic mesovortex (Weisman 1993). Discrete storms ahead of and along the southern end of the line produced severe hail in southern Missouri and northern Arkansas, while the main bow echo continued to produce damaging wind. The mature bow echo and attendant mesovortex then affected Tennessee, southwest Kentucky, and far northern Alabama throughout the afternoon before it dissipated in eastern Tennessee after 2300 UTC.

The SPC’s Day 1 Convective Outlook from 0600 UTC 3 May 2020 and valid from 1200 UTC 3 May to 1200 UTC 4 May (Fig. 3a) delineated a marginal risk (5% probability) of severe hail and wind extending from the Missouri–Arkansas border to the Chesapeake Bay. Severe storms in this area were expected to form along the southward-moving cold front in the afternoon, according to the forecaster’s written text discussion (not shown). No severe risk (less than 5% probability) was delineated over southeast Kansas or southwest Missouri, indicating that the early day severe convection that occurred there after 1200 UTC was not anticipated. The SPC introduced a slight risk of severe storms (15% probability of severe hail and severe wind) in the 1300 UTC Day 1 Outlook update (Fig. 3b), after the first storms produced severe weather in southeast Kansas, and then upgraded the outlook to an

(a) Scale-aware Grell–Freitas (Grell and Freitas 2014)
(b) Thompson (Thompson et al. 2008)
(c) RRTMG (Mlawer et al. 1997; Iacono et al. 2008)
(d) MYNN (Nakanishi and Niino 2009)
(e) Noah (Chen and Dudhia 2001)
enhanced risk (30% probability of severe wind) in the area ahead of the bow echo system in the subsequent 1630 and 2000 UTC Day 1 Outlook updates (Figs. 3c,d). Therefore, the greater areal extent and coverage of the severe weather threat only became apparent after the MCS had developed. This highlights the low level of predictability that was associated with the early evolution of severe convection in southeast Kansas and southwest Missouri, and its subsequent formation into a rapidly moving bowing MCS.

RAP analyses at 0900 UTC 3 May 2020, just before elevated CI in south-central Kansas, are shown in Fig. 4. Zonal flow at 500 hPa predominated over much of the central Great Plains toward the middle Mississippi Valley ahead of an upstream weak ridge axis that extended from Colorado to Montana. A low-amplitude shortwave trough was moving eastward near the Kansas–Nebraska border (Fig. 4a). At 850 hPa (Fig. 4b), a low was centered over the Texas Panhandle and western Oklahoma with a front extending to the east-northeast across southeast Kansas, central Missouri, and areas north of the Ohio River. Warm advection in the 800–700-hPa layer occurred from central Oklahoma to southeast Kansas. The surface observations at 0900 UTC (Fig. 5) show a low centered in southwest Oklahoma and northern Texas and an attendant quasi-stationary front extending east-northeastward, south of the 850-hPa front. The elevated thunderstorms in south-central Kansas occurred well north of the surface front in association with warm advection at low levels. Later during the day, the front slowly moved south over the Central Plains and the Ohio River Valley.

The Topeka, Kansas (TOP), 1200 UTC observed sounding (Fig. 6a) shows a cool stable surface-based layer and a dry air mass extending to nearly 820 hPa. There was northerly flow near the surface and westerly flow aloft into the middle and upper troposphere, with a nearly saturated layer to nearly 600 hPa and no apparent low-level warm advection in the wind profile. The elevated instability present in this sounding is evidenced by 458 J kg⁻¹ of most unstable convective available potential energy (MUCAPE) for lifted parcels originating near 820 hPa and no surface-based CAPE (SBCAPE). The Springfield, Missouri (SGF), 1200 UTC observed sounding (Fig. 6b) sampled the air mass nearly along the surface front (Fig. 5). Westerly winds are observed beginning just above the surface, where there is only a very shallow temperature inversion. A moist layer extends up to nearly 700 hPa, above which an elevated mixed layer is present. This sounding has larger instability at the surface (SBCAPE of 795 J kg⁻¹) compared to the TOP sounding, though there is convective inhibition as well when lifting surface parcels. The MUCAPE for parcels originating at nearly 870 hPa is 2348 J kg⁻¹ with no convective inhibition. The effective bulk wind difference of 18 m s⁻¹ indicates an environment supportive of organized convection (Rasmussen and Blanchard 1998; Thompson et al. 2007).

b. MPAS ensemble performance

Figure 7 shows the smoothed neighborhood maximum ensemble probabilities (NMEPs; Schwartz and Sobash 2017) of hourly maximum 2–5-km updraft helicity (UH; Kain et al. 2010) exceeding 75 m² s⁻² and hourly maximum wind speed at the lowest model level (nearly 31 m above ground) exceeding 20 m s⁻¹ in the 24-h period from 0600 UTC 3 May to 0600 UTC 4 May 2020. The lowest model level wind is examined because diagnostic 10-m winds from CAMs tend to underestimate observed convective wind gust speeds, whereas model winds in the boundary layer above the surface appear to correspond better with observed gusts (Benjamin et al. 2020). The NMEPs are calculated using a grid having horizontal spacing of nearly 80 km (radius of influence of 40 km) and then smoothed using a two-dimensional Gaussian kernel with a 120-km smoothing length scale. The UH is commonly used as a diagnostic for rotating updrafts in CAM simulated storms.
and the threshold used here has been shown to have utility in indicating potential for severe weather (Sobash et al. 2016). The ensemble indicated higher UH smoothed NMEPs (Fig. 7a) in a corridor extending from southeastern Kansas to southern Missouri/northern Arkansas and across Kentucky into West Virginia associated with the cold front. The highest probabilities coincided well with the western half of the severe report axis (Fig. 1) from southeast Kansas into

![Fig. 5. Surface observations at 0900 UTC 3 May 2020. The temperature (red) and dewpoint temperature (green) values are in degrees Fahrenheit. The front and the dryline are drawn using the typical symbols. TOP and SGF refer to Topeka, Kansas, and Springfield, Missouri, sounding station locations, respectively. Source: https://www.wpc.ncep.noaa.gov/html/sfc-zoom.php.](image-url)

![Fig. 6. Observed 1200 UTC 3 May 2020 soundings at (a) Topeka, Kansas (TOP), and (b) Springfield, Missouri (SGF). The thin red line is the virtual temperature. Winds in the hodograph are in m s⁻¹. All CAPE calculations use virtual temperature correction.](image-url)
southeast Missouri, but the UH guidance focused north of the actual derecho track from near the Mississippi River eastward. The wind speed smoothed NMEPs (Fig. 7b) focused higher probabilities (30%) farther east than the UH probability maxima, although north of the severe reports. This suggested the evolution from discrete convection in the early stages to a forward propagating MCS later in the event, although assessment of convective mode is necessarily accomplished by examination of simulated reflectivity (see below).

The wind probabilities corresponded better to the observed report axis east of the Mississippi River compared to UH, including the 10% probability contour farther south over Tennessee. In general, the ensemble probability guidance indicated a relatively low probability of occurrence of severe wind-producing storms from the Ozarks into the Tennessee Valley, consistent with operational forecasts (Figs. 3a,b) prior to the development of the observed MCS. However, we cannot discern from a single case the reliability of the NMEP products nor their optimal interpretation as forecast guidance for derechos; this requires analysis of more cases and is beyond the scope of this study.

We subjectively classified the MPAS ensemble members into three performance categories by examining simulated reflectivity forecasts to assess the ability of each member to predict the precursor initial elevated convection, the bow echo formation and subsequent evolution. Since the focus of this study is on physical processes associated with different outcomes in the ensemble, the subjective assessment of each member’s solution is necessary to separate members that might be similar but due to very different physical processes. The mesoscale processes can be more easily isolated by analyzing only one member considered to be representative of a given category. For example, it is more difficult to comparatively explore the evolution of an MCS’s cold pool and the development of a rear inflow jet using ensemble means of each group.

**FIG. 7.** (a) 40-member MPAS ensemble smoothed NMEPs of 24-h maximum updraft helicity greater than 75 m$^2$ s$^{-2}$ (%; brown contours) and 24-h maximum updraft helicity greater 75 m$^2$ s$^{-2}$ within 40 km of a point for each member (shaded with different colors for each member). (b) As in (a), but for maximum wind speed at the lowest model level (~31 m above ground level) greater than 20 m s$^{-1}$. The gray dots indicate the grid with 80-km horizontal spacing used to calculate the NMEPs. SPC’s storm reports are shown as in Fig. 1.
The three categories are as follows: type 1) members that were unable to generate the predecessor sustained convection in south-central Kansas between 0900 and 1100 UTC and did not generate a bow echo MCS (8 members); type 2) members that correctly generated the predecessor convection and were able to predict a dominant bow echo MCS in approximately the right time and location, even though some aspects of the structure, evolution, and track were different from the observed system (6 members); and type 3) members that predicted CI in Kansas reasonably well, but the convection did not organize into a dominant bow echo (26 members). None of the members generated a dominant bow echo MCS from convection somewhere other than south-central Kansas.

Figure 8 shows the simulated composite reflectivity every 2 h from 1000 to 2000 UTC of a representative member from each of these three categories and corresponding MRMS composite reflectivity. Member 2 (type 1) was unable to produce sustained convection in south-central Kansas between 0900 and 1100 UTC (Fig. 8a) and did not generate a bow echo MCS (Figs. 8e,i,m). This member generated widespread...
convection starting around 1800 UTC along the cold front near the Mississippi river (Figs. 8q,u). Member 6 (type 2) was successful at producing a bow echo MCS. The first storms in south-central Kansas at 1000 and 1200 UTC (Figs. 8b,f) are in good agreement with the observed early evolution of convection (Figs. 8d,h). Convection grew upscale between 1400 and 1600 UTC (Figs. 8j,n), and the bow echo is well defined at 1800 UTC (Fig. 8r), similar to observations (Fig. 8t). Member 6 was unable to maintain the bow echo organization into Tennessee and Kentucky after this time (Fig. 8v). In member 18 (type 3), convection began in south-central Kansas after 0900 UTC (Figs. 8c,g) and evolved into a multicellular cluster at 1400 and 1600 UTC (Figs. 8k,o). The CI was not as widespread and intense in south-central Kansas as in member 6. Additionally, another group of intense storms formed spuriously downstream in southwest Missouri which evolved into an organized multicellular cluster between 1000 and 1400 UTC (Figs. 8c,g,k), a characteristic observed in most type 3 members (not shown). These two clusters (hereafter referred to as “MCSI” and “MCS2” as indicated in Fig. 8o) moved east-southeastward along the Missouri–Arkansas border and briefly displayed bow echo shapes, but neither of them were able to sustain a well-defined bow echo MCS (Figs. 8s,w). The spurious convection leading to MCSI in member 18 is also present to some extent in members 2 (but much weaker) and 6 (Figs. 8a,b,e,f), but it weakens in member 2 and merges with the main MCS in member 6, without affecting the main MCS considerably. Members 2, 6 and 18 (hereafter referred to as “NOCI,” “BOW,” and “NOBOW,” respectively) will be used to represent the three category types in the subsequent analyses.

Based on the three MPAS performance categories, the observed radar evolution, and the SPC Convective Outlooks, the predictability issues of this event can be divided into three chronological phases: 1) the elevated CI in south-central Kansas between 0900 and 1100 UTC, 2) the transition from discrete storms to a bow echo MCS over southeast Kansas and southwest Missouri between 1100 and 1500 UTC, and 3) the maintenance and evolution of the bow echo over Tennessee and southern Kentucky after 1800 UTC. The CI in south-central Kansas will be explored by comparing NOCI (unsuccessful) and BOW (successful) to assess if the environment was less favorable for CI in NOCI. The evolution of elevated convection to a mature bow echo will be addressed by comparing NOBOW (unsuccessful) and BOW (successful) and assess if differences in cold-pool formation and intensity helped support faster organization into a single surface-based forward-propagating bow echo system in BOW. The maintenance and evolution of the bow echo over Tennessee was not well predicted by any MPAS ensemble member, so we explore this predictability issue by comparing BOW with the RAP analyses and surface observations.

A comparison of representative ensemble members was used by Schumacher et al. (2013), Trier et al. (2015, 2019), among others, in order to analyze the physical processes leading to different outcomes in specific members. This methodology is not fully objective as ensemble sensitivity tests (e.g., Torn and Hakim 2008; Torn and Romine 2015; Berman et al. 2017), but some subjectivity is required to accurately distinguish members with different storm modes. For example, the NOBOW member would provide some utility to forecasters in terms of the areal coverage of severe storms activity, but its evolution of convective mode differed considerably from that of the observed system (Fig. 8). A limitation of using only one representative member is uncertainty in the representativeness of environmental conditions based on assessments from individual members. To evaluate the representativeness of these selected members, we also analyzed the ensemble means composed of all members in each type.

c. Convective initiation over south-central Kansas

Members NOCI and BOW are compared to evaluate why sustained convection formed over south-central Kansas only in BOW. Figure 9 shows the differences between the two members at 0900 UTC 3 May, immediately prior to CI (Figs. 8a,b). The 850-hPa geopotential height (Fig. 9a) is 1–2 dam lower in BOW over an area extending from the Texas Panhandle to northern Oklahoma, which is associated with stronger southerly flow (by 1 m s\(^{-1}\)) over southern Kansas and east-central Oklahoma relative to NOCI. As a result, the 850-hPa temperature and dewpoint (Fig. 9b,c) over south-central Kansas are 0.5–1°C higher at this level in BOW. In both members, there was warm advection between 800 and 700 hPa (not shown), but no substantial differences were apparent between the members. The MUCAPE (Fig. 9d) is also larger by 200–600 J kg\(^{-1}\) in BOW where convection initiated, though NOCI also has considerable instability (MUCAPE > 1500 J kg\(^{-1}\)) in the area. Therefore, the relatively lower 850-hPa geopotential height over southwest Oklahoma and northern Texas contributed to enhanced southerly flow in low levels, which is associated with a warmer and more moist environment and larger MUCAPE in south-central Kansas prior to CI in BOW.

Area-averaged soundings where convection initiated are shown in Fig. 10. A rectangle between latitudes 37.4°N and 38.8°N and between longitudes 98.3°W and 96.3°W is used (Figs. 10c,d). The mean, 25th and 75th percentile profiles of temperature and dewpoint are derived from the distributions comprised of all 2966 MPAS mesh points within the rectangle. The average soundings for both members are characterized by a shallow layer of relatively colder and drier air at the surface, above which the air is nearly saturated from approximately 850 to 750 hPa (Figs. 10a,b). This near-saturation layer is ~50 hPa deeper and the temperature and dewpoint are 0.5–1°C higher in BOW, in agreement with Figs. 9b,c. The winds change from northeasterly within the stable layer below 850 hPa to southwesterly above it, consistent with the surface front being to the south of the area at this time. Above 750 hPa, there is drier air in both members. The NOCI profile is ~1°C warmer at nearly 680 hPa, which is associated with greater area-averaged convective inhibition of the most unstable parcel (MUCIN). The area-averaged vertical velocity (\(\omega\); Figs. 10a,b) is calculated using only values between the 25th and 75th percentiles for each vertical level to prevent the inclusion of mesh points.
with convective cells, particularly in BOW (Fig. 10c). The average $w$ is larger between 850 and 500 hPa in BOW (>2 cm s$^{-1}$); ascent is also evident in NOCI, but the average $w$ is lower in the same layer (~1 cm s$^{-1}$). The 75th percentile of $w$ in BOW exceeds 10 cm s$^{-1}$ above 800 hPa. There is also a better vertical collocation of the area-averaged ascent with the deeper, nearly saturated layer in BOW, which increases the probability of CI and likely contributed to a more widespread activity. In NOCI, the association of comparatively lower temperature and dewpoint in the moist layer between 850 and 750 hPa and relatively warmer and drier air above 700 hPa would require greater upward motion for parcels in the moist layer to attain their level of free convection (LFC), but the ascent in this member was weaker. The $w$ profiles in Figs. 10a,b suggest mesoscale ascent was responsible for CI in this area. Area-averaged soundings between 0600 and 0900 UTC in the same area (not shown) also indicated greater ascent in BOW several hours before CI.

To investigate further the reason for failure in sustained CI in NOCI, Fig. 11 shows cross sections along the AB line indicated in Figs. 10c,d. CAPE and LFC are a function of the level from which the parcel is lifted. Convection initiated in BOW between latitudes 37.5° and 38.5°N (Fig. 8b). The cross section of CAPE (Figs. 11a,b) in both members shows that the instability where convection initiated was elevated (there was zero CAPE near the surface). The surface front was located near 37°N, but the warm and moist air with greater instability sloped well to the north of the surface front aloft. The CAPE is 200–600 J kg$^{-1}$ greater in the latitude range where convection began in BOW as compared to NOCI, as seen in Fig. 11c, which shows the difference in CAPE between the cross sections for BOW and NOCI (Fig. 11a minus Fig. 11b).

The vertical distance that parcels have to be lifted to reach their LFCs (hereafter “LFC distance”) is also shown in Fig. 11. The LFC distance is less than 600 m between 850 and 750 hPa (Figs. 11a,b) in both members, where the area-averaged soundings exhibited nearly saturated air (Figs. 10a,b). Above this layer, despite the moderate instability (CAPE > 1000 J kg$^{-1}$) the LFC distance is much greater (>1000 m) due to the presence of relatively drier air. The LFC distance is even greater (>2000 m) within the stable layer near the surface as well. The largest difference in LFC distance between

![Diagram](image-url)
The lower LFC distance for parcels between 850 and 750 hPa in BOW (Fig. 11d), in association with greater ascent in this layer (Figs. 10a,b), appear to be key factors for the differences in CI between BOW and NOCI. The 75th percentile of \( w \) in the 850–750-hPa layer in BOW is nearly 10 cm s\(^{-1}\), which is calculated using only values between the 25th and 75th percentiles for each vertical level, is shown in the bottom-left corner within the 25th and 75th percentiles. The magenta, dashed line is the most unstable parcel path derived from the mean sounding in the area. (c),(d) Composite reflectivity (dBZ) at 0900 UTC 3 May 2020 for BOW and NOCI, respectively. The rectangle marks the area used to generate the area-averaged soundings in (a) and (b) (between latitudes 37.4°N and 38.8°N and between longitudes 98.3°W and 96.3°W). The AB line marks the cross sections in Fig. 11.

**d. Evolution from elevated convection to a mature bow echo**

In this section, we compare BOW with NOBOW, in which elevated convection occurred in south-central Kansas but a dominant bow echo MCS did not organize. The main hypotheses are that: 1) the cold pool associated with convection in BOW intensified more rapidly, enabling the MCS to organize more quickly, and 2) MCS1 in NOBOW (Fig. 8a) was
Fig. 11. (a),(b) Cross sections along AB (Figs. 10c,d) of CAPE (shaded; J kg\(^{-1}\)) and LFC distance (contours every 200 m) as a function of the pressure level from which the parcel is lifted for BOW (Fig. 10c) and NOCI (Fig. 10d), respectively, at 0900 UTC 3 May 2020 (15-h forecast). (c),(d) Differences in CAPE (J kg\(^{-1}\)) and LFC distance (m), respectively, between BOW and NOCI. The yellow dashed lines represent the approximate latitude range where convection initiated in BOW (37.5°–38.5°N). The terrain height is shown at the bottom in brown.

1) COLD POOL DEVELOPMENT

The MCS in BOW had propagated farther to the east than in NOBOW by 1500 UTC, which is surmised to be a result of stronger cold pool and attendant cold-pool propagation due in part to greater precipitation generated in BOW (Fig. 12). To evaluate the cold pool evolution during the first hours of the MCS evolution, the buoyancy (\(B; \text{Weisman 1992}\)) is calculated using

\[
B = g \left( \frac{\theta - \overline{\theta}}{\overline{\theta}} - 0.61(q_v - \overline{q_v}) - q_i \right),
\]

where \(g\) is the gravitational acceleration, \(\theta\) is the potential temperature, \(q_v\) is the water vapor mixing ratio, \(q_i\) is the mixing ratio of all hydrometeors, and the bars denote environmental conditions, which are calculated by averaging the variable over the 50 × 50 km\(^2\) area ahead of the MCS.

Figure 13 shows the buoyancy in the lowest model level at 1200, 1500, and 1800 UTC for BOW and NOBOW. At 1200 UTC, there is relatively colder, more stable air to the north and warmer, unstable air to the south of where convection is ongoing. In BOW (Fig. 13a), there is relatively cold air near the surface just beneath the 40-dBZ convective cores, though the negative buoyancy at this time is small (\(B < -1 \times 10^{-1} \text{ m s}^{-2}\)). The weak and isolated convection in NOBOW, on the other hand, did not produce noticeable convectively generated negative buoyancy near the surface (Fig. 13b), which relates to the much lower precipitation accumulated over the last 3 h (Figs. 12a,b). The negative buoyancy near the surface in BOW suggests the multicell clusters at this time had already produced a surface-based cold pool. This does not necessarily confirm that convective updrafts along the leading edge of the cold pool were based at the surface at this time, since elevated MCSs have been found to generate surface cold pools (Miller et al. 2020; Hitchcock et al. 2019). However, it does suggest that BOW may have had a faster transition from elevated to surface-based convection in comparison to NOBOW.

The differences in the cold pool structure and intensity between BOW and NOBOW increase at 1500 UTC (Figs. 13c,d). The cold pool in both members is more intense beneath the 40-dBZ cores, but BOW has more negative buoyancy (\(B < -2 \times 10^{-1} \text{ m s}^{-2}\)) in the immediate wake of the leading edge of the MCS. Additionally, the buoyancy gradient across the leading edge of the MCS gust front is much larger in BOW. Both members have a mesohigh associated with the cold pool, but only in BOW there is a clear trailing wake low. At 1800 UTC (Fig. 13e), when the MCS in BOW is a mature bow echo (Fig. 8r), there is an intense cold pool (\(B < -4 \times 10^{-1} \text{ m s}^{-2}\)) near the leading convective line associated with a mesohigh with sea level pressure greater than 1016 hPa and a trailing wake low with sea level pressure less than 1006 hPa. The area with lesser negative buoyancy near the wake low is responsible for reducing instability ahead of MCS2, precluding its further intensification and organization.
wider in comparison to that at 1500 UTC (Fig. 13c). The gradient in both sea level pressure and buoyancy across the leading edge of the convective line increased further and resulted in a faster forward speed at this time. MCS2 in NOBOW presents a mesohigh and a wake low at 1800 UTC (Fig. 13f), but they are smaller in areal coverage and less intense than in BOW. It is also clear that the MCS2 is moving toward relatively cold and more stable air in the wake of MCS1 (Fig. 8o), which impacts the maintenance and/or further intensification of MCS2 (the impact of MCS1 on MCS2 will be discussed in more detail later). The different evolution of the cold pool in BOW and NOBOW is associated with greater precipitation that occurred in BOW, which likely caused more intense diabatic cooling in low levels and enhanced downward momentum transfer (Mahoney et al. 2009). Greater downward momentum transfer in BOW is also supported by a stronger rear-inflow jet (RIJ; Weisman 1992), as will be shown in the next section. Also, stronger hourly maximum wind speeds at the lowest model level were indicative of greater downward momentum transfer in BOW (not shown). These processes combined to favor faster cold-pool intensification and MCS motion (Corfidi 2003).

2) REAR INFLOW JET

Another important feature related to organized bow echo MCSs is the RIJ (e.g., Weisman 1992; Wakimoto et al. 2006; Grim et al. 2009). Figure 14 presents the evolution of the 1–3-km (above ground level) storm-relative (SR) mean wind (taken as a proxy for the RIJ height) and vorticity. The initial convection in BOW is associated with larger vorticity ($>6 \times 10^{-4} \text{ s}^{-1}$) at 1200 UTC (Figs. 14a,b) relative to NOBOW, but no organized RIJ structure (westerly SR winds) is evident. At 1500 UTC (Figs. 14c,d), the leading edge of the MCS in BOW has a maximum in 1–3-km vorticity greater than $12 \times 10^{-4} \text{ s}^{-1}$ and 1–3-km SR winds < 4 m s$^{-1}$ just to the south of the vorticity maximum, indicating the early stage of organization of a mesoscale cyclonic circulation associated with a RIJ (Wakimoto et al. 2006). The MCS in NOBOW at this time is also associated with a low-level vorticity maximum ($<10 \times 10^{-4} \text{ s}^{-1}$) and weak SR winds (4–8 m s$^{-1}$) just behind the convective line, but these features are less evident compared to those in BOW.

There is a much greater difference between BOW and NOBOW 3 hours later at 1800 UTC (Figs. 14e,f). The RIJ is evident in the MCS in BOW at this time, extending from the rear of the MCS to the leading edge, where the maximum 1–3-km SR wind speed exceeds 20 m s$^{-1}$. The maximum 1–3-km SR wind speed occurs just to the west-southwest of an intense ($>24 \times 10^{-4} \text{ s}^{-1}$) 1–3-km vorticity maximum. In NOBOW, both MCSs are associated with weak 1–3-km SR wind behind the convective line, but no rear inflow is evident. Additionally, the cyclonic vorticity along the convective lines of both MCSs in NOBOW is notably weaker ($<10 \times 10^{-4} \text{ s}^{-1}$). These factors relate to the inability of both MCSs in NOBOW to become as organized and intense as the single dominant bowing MCS in BOW.
3) COLD POOL PROPAGATION SPEED

The difference in density between cold-pool air and the warmer air in the immediate prestorm environment is proportional to the cold-pool propagation speed (e.g., Rotunno et al. 1988; Weisman 1992; Bryan and Rotunno 2008). Therefore, the faster cold-pool intensification in BOW likely played an important role for greater acceleration of the convective line during the early hours of the MCS. At 1500 UTC (Figs. 14c,d) and mainly at 1800 UTC (Figs. 14e,f) the MCS in BOW is moving faster (estimated propagation speed of 20–24 m s$^{-1}$, which is similar to the observed derecho forward motion of ~24 m s$^{-1}$) than both MCSs in NOBOW (estimated propagation speed of 16–20 m s$^{-1}$). These propagation speeds are greater than the 0–6-km mean wind in the area (13–17 m s$^{-1}$) consistent with findings by Johns and Hirt (1987), and Corfidi et al. (2016) on progressive derechos. More widespread convection in BOW in the early stages of the MCS likely contributed to greater downward transfer of horizontal momentum, favoring a faster cold pool propagation (Corfidi 2003). Campbell et al. (2017) showed that more organized MCSs, with a single bowing convective line and line-end mesovortex (their “Class 1” MCSs), tend to move faster than MCSs with lower degrees of organization.

4) THERMODYNAMIC ENVIRONMENT EFFECTS ON MCS EVOLUTION

It is evident from Fig. 13f that MCS1 negatively affected the evolution of MCS2 by cooling the near-surface air and stabilizing the environment ahead of MCS2 in NOBOW. A similar detrimental impact from convection ahead of the main MCS is present in most type 3 members as well (not shown). MCS1 started to organize at 1200 UTC in southwest Missouri.
(Fig. 8g), and then moved east-southeastward along the Missouri–Arkansas border (Figs. 8k,o,s). The convection from which MCS1 evolved at nearly 1200 UTC was spurious (Figs. 8d,h,l). Downstream convection was also present in BOW as an apparent “warm advection wing” that has been observed with some progressive derechos as described by Smith (1990) and Przybylinski (1995) (Figs. 8b,f,i). In BOW, however, this convection did not evolve into a separate MCS and was absorbed by the faster moving main bow echo MCS.

Figure 15 shows the MUCAPE at 1500 and 1800 UTC for both members. The reduction of MUCAPE in the wake of MCS1 in NOBOW at 1500 UTC (Fig. 15b) is well evident over extreme south-central Missouri. At 1800 UTC, the MCS in BOW (Fig. 15c) is moving toward an environment with MUCAPE of generally 1500–2000 J kg\(^{-1}\), whereas much lower CAPE (<1000 J kg\(^{-1}\)) is observed ahead of MCS2 in NOBOW (Fig. 15d). Therefore, as MCS1 progressed east-southeastward along the Missouri–Arkansas border it substantially reduced the instability in its wake and ahead of MCS2. This factor, in combination with a slower cold-pool intensification rate and a much weaker rear inflow jet in MCS2, led to less intense MCSs in NOBOW in contrast to an intense, dominant bow echo MCS in BOW.

e. MCS maintenance over Tennessee

Last, we explore why none of the MPAS ensemble members were able to maintain the bow echo over Tennessee. Figures 16a and 16b show the MUCAPE, 0–6-km bulk wind difference (BWD) and the derecho composite parameter (DCP; Evans and Doswell 2001) from the RAP analysis and MPAS BOW at 1800 UTC 3 May 2020, when the observed and simulated MCSs were over southeast Missouri and

FIG. 14. Storm-relative wind magnitude (shaded; m s\(^{-1}\)), vectors and vorticity (green contours every 3 \(\times\) 10\(^{-4}\) s\(^{-1}\) starting at 3 \(\times\) 10\(^{-4}\) s\(^{-1}\)) averaged in the 1–3 km above ground level layer and the 40-dBZ simulated composite reflectivity contour (red) at (a),(b) 1200; (c),(d) 1500; and (e),(f) 1800 UTC 3 May 2020 for (left) BOW and (right) NOBOW. The estimated storm motion components \(u\) and \(v\) are shown in the bottom-left corner. The reflectivity contour is slightly smoothed for better visualization.
northeast Arkansas. The DCP is a nondimensional parameter that includes MUCAPE, downdraft CAPE, 0–6-km BWD and 0–6-km mean wind in its formula. DCP values greater than 1 indicate an environment favorable for a derecho-producing MCS. The RAP analysis indicates much higher instability over western and middle Tennessee (1400–2200 J kg$^{-1}$) in comparison to BOW (400–1200 J kg$^{-1}$), but the 0–6-km BWD is similar. The lower instability in BOW is also present in other members (ensemble mean MUCAPE lower than 1000 J kg$^{-1}$ in Fig. 16b). The DCP is less than 0.5 in BOW and greater than 1 in the RAP analysis over western and middle Tennessee. DCP values greater than 3 are present in northern Tennessee and southern Kentucky based on the RAP analysis, suggesting an environment favorable for MCS maintenance and compatible with the observed evolution of the system (Fig. 1).

The 2-m dewpoint was also lower in BOW over western Tennessee (Figs. 16c,d). The 2-m dewpoint errors in the RAP analysis are generally less than 1.8°C in this area, whereas BOW dewpoints were lower than observations by more than 2.8°C at some grid points. The RAP analysis errors are expected to be smaller, since the RAP assimilates these surface observations. There were positive errors greater than 2°C in 2-m dewpoint in BOW over Kentucky (Fig. 16d) collocated with MUCAPE greater than 2000 J kg$^{-1}$ (Fig. 16b), where several spurious storms formed in the model. The simulated bow echo in BOW likely moved north of the observed track over Kentucky, while weakening over Tennessee, due to the northward displacement of the high instability axis (and higher values of DCP) relative to observations.

Point soundings in western Tennessee (star in Figs. 16a,b) derived from the RAP analysis and MPAS BOW at 1800 UTC (Fig. 17) show much drier air at surface and low levels in BOW, resulting in lower mixed-layer CAPE (207 J kg$^{-1}$ compared to 1272 J kg$^{-1}$ in the RAP analysis) and higher convective inhibition for the mixed-layer parcel (358 J kg$^{-1}$ compared to 75 J kg$^{-1}$ in the RAP analysis). Proximity soundings ahead of quasi-linear convective systems analyzed by Coniglio et al. (2007) indicate that CAPE is one of the best parameters in discriminating mature from dissipating systems. Other parameters with useful discrimination potential include the shear vector over a deep layer and the low-to-upper-level mean wind, both of which are similar in the RAP analysis and BOW (Fig. 17). Therefore, the low instability ahead of the simulated MCS was likely the main factor in limiting the MCS maintenance over western and middle Tennessee in BOW, which is consistent with DCP values being lower than 1. The soundings suggest low-level (~900–700 hPa) subsidence and transport of relatively drier air from the southwest due to stronger southwesterly low-level flow in BOW, which relate directly to the lower dewpoints at surface and low levels in the model.

4. Conclusions

A 40-member MPAS ensemble with convection-allowing resolution initialized with time-lagged GEFS ICs is used to analyze the low predictability associated with the 3 May 2020 derecho, which includes CI and the organization and track of the bow echo MCS. The only difference between each ensemble member was the ICs, with the dynamic core, physical
parameterizations and variable mesh sizes remaining constant across all members.

The MPAS ensemble UH and wind speed smoothed NMEPs indicated severe weather potential in the early stages of the 3 May 2020 event, but suggested a low probability of a dominant, progressive MCS formation and its continued rightward propagation and maintenance into the warm sector over Tennessee. The majority of the UH and wind speed tracks in the early stages were not associated with a single dominant bow echo MCS, but with discrete storms with rotating updrafts embedded within multicell clusters. When the convective mode is considered, the number of better performing members (those developing a forward-propagating bowing MCS at approximately the right place and time) was relatively low (6 out of 40), also indicating a low (but non-zero) probability from MPAS guidance for a progressive MCS to develop across the area.

The formation of the progressive bow echo system in the MPAS ensemble is related to the initial prediction of numerous elevated thunderstorms in south-central Kansas between 0900 and 1200 UTC. This convection initiates in association with weak low-level warm advection above relatively colder stable air at surface, which then consolidates and evolves into a progressive bowing MCS during the morning. By comparing environments between the better performing BOW member and the poorly performing NOCI member, which did not produce sustained convection in south-central Kansas, BOW was characterized by comparatively higher (by 0.5–1.0°C) temperature and dewpoint in low levels, which contributed to larger MUCAPE values in the region where convection initiated. Additionally, the LFC distances for parcels in the 850–750-hPa layer were lower by more than 600 m in BOW and thus more supportive of CI. Therefore, the initial convection that evolved later into a bow echo MCS was highly sensitive to the temperature and dewpoint profiles at low-mid levels in south-central Kansas, which led to differences in CAPE and LFC distance of elevated parcels and resulted in less widespread CI in NOCI.

The intensity and areal coverage of the elevated storms in south-central Kansas were also important factors for the
formation of the subsequent bow echo MCS. In NOBOW, CI in south-central Kansas occurred, but storms were fewer in number and less intense than in BOW. As a result, the precipitation rate in NOBOW associated with these elevated storms was lower, the cold-pool development was weaker and its intensification was slower to occur, and the MCS simulated in this member was unable to evolve into a mature bow echo system. The latter failure was also strongly related to the spurious development of downstream convection that organized into a lead MCS (MCS1) in NOBOW. The outflow from this erroneous convective system stabilized the environment ahead of the second MCS (MCS2) in southern Missouri, limiting its potential organization and intensification into a progressive bow echo system. While the two MCSs in NOBOW each attained a period of strong-to-severe potential as indicated by their UH fields, this incorrect evolution precluded development of a single dominant MCS in NOBOW. The MCS in BOW had an intense RIJ, a line-end mesovortex, an intense cold pool associated with a mesohigh–wake low couplet, and moved faster than the mean wind, all of which are common characteristics of derecho-producing bow echo MCSs (Johns and Hirt 1987; Johns 1993; Corfidi et al. 2016; Campbell et al. 2017). Last, the early dissipation of the MCS in BOW over western Tennessee and its propagation north of the observed track over Kentucky was associated with the northward shift of the high-CAPE axis in BOW relative to observations.

Given the high sensitivity to the formation and character of nocturnal, elevated convection in this case, more research is needed to understand this phenomenon and improve its representation in CAMs. Other warm-season progressive derecho cases, including the high-impact 10 August 2020 derecho, also evolved from nocturnal, elevated convection and were associated with low operational predictability. The transition from discrete convection to a bow echo MCS also needs to be better understood. Improvements in the operational predictability of warm-season derechos rely on better understanding of the physical processes governing bow echo MCSs but also how the sensitivity to these processes is handled by CAEs. Future studies should also investigate whether warm-season progressive derechos are associated with inherently low probability in CAEs, owing to subtle forcing for ascent and complex meso- and storm-scale processes associated with the transition from elevated-to-surface based convection and upscale growth into a bow echo system.

As is often the case in warm-season progressive derechos, the 3 May 2020 event had low operational predictability in the short range (<12 h). This study showed that small differences in the environment derived from different ICs led to vastly distinct outcomes in terms of the severe weather threat in the CAM guidance. This case also illustrates the complex decision-making and risk communication challenges faced by severe weather forecasters when utilizing CAE guidance that provides solutions ranging from a minimal severe weather threat to high-end significant severe potential.

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Data availability statement. The MPAS ensemble simulations upon which this study is based are too large to archive or to transfer. Instead, we provide in section 2 all the information needed to replicate the simulations. The model code can be found at http://mpas-dev.github.io/. MPAS namelists are available upon request. Initial and boundary condition files from the GEFS are available at http://weather.uwyo.edu/upperair/sounding.html. RAP analyses data are available at https://www.ncep.noaa.gov/data-access/model-data/model-datasets/global-ensemble-forecast-system-rap. Observed soundings are available at http://weather.uwyo.edu/upperair/sounding.html.

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