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Prediction of convective morphology in near-cloud permitting WRF model simulations

by

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Program of Study Committee:
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Convective morphology was analyzed with 3-km WRF-ARW simulations for 37 events during the warm season from 2006 to 2010. Ten classifications were used to identify convective modes displayed in each event. An objective scoring method, based on normalized time and the type of mode exhibited, was developed to measure the accuracy of the modeled morphologies when compared to radar observations. Trends in the simulated evolution were discussed, as well as common discrepancies between the model and observed events. Environmental conditions before convective initiation were obtained from RUC analyses, and statistically significant associations between the parameters and the model accuracy scores were found. Finally, four cases highlighting the main morphological issues in the simulations were investigated further.

Overall, the simulations entailed more cellular modes and fewer linear modes than the observed systems. Bow echoes and linear systems with trailing stratiform rain regions were especially difficult for the model to forecast. Of the 21 cases with an observed bow echo, only 8 featured a simulated bow echo. Twelve cases included a missed squall line with trailing stratiform rain. Cellular modes were simpler to forecast, as 75% of the forecast comparisons with an observed cellular mode also featured a modeled cellular mode. The simulations usually portrayed convective evolution more accurately when the initial synoptic environment included strong deep-layer shear and cool potential temperatures at the level of maximum theta-E. Major timing errors with convective initiation or dissipation in the simulations usually occurred when initial 0-6 km shear was very low, surface potential temperatures were cool, and when potential temperatures quickly warmed with height.

The case studies showed that differences in wind shear and cold pool strength and development between the simulations and observations were responsible for many convective mode discrepancies. Strong boundary-normal shear, a warm cold pool, and very dry air within the rear inflow jet inhibited the formation of a trailing stratiform rain region in the simulation for the first case study. Weak deep-layer shear and the lack of a fully-developed cold pool in the model did not allow the simulated line to bow out in the second case study. Weaker forcing near the surface in the simulation compared to the observations
allowed for clustered cells instead of a broken line in the third case study. Mid-layer shear not oriented along the boundary did not provide for proper cold pool merging for linear development in the WRF output for the fourth case study. Sensitivity tests were also conducted for microphysical schemes and initial and lateral boundary conditions for the first and second case studies. None of the microphysics schemes produced the observed convective mode, but GFS initial conditions in the first case study produced a trailing stratiform region.
CHAPTER 1. GENERAL INTRODUCTION

1. Background

The prediction of deep moist convection is undoubtedly an important aspect of meteorology. Thunderstorms have a profound effect on the climate, the economy, and the population as a whole. For most of the Great Plains and Midwest, at least half of the annual precipitation is derived from thunderstorms (Changnon 2001a). Thus, crops and the hydrologic cycle in the center portion of the United States are highly dependent on convective precipitation. Along with the benefits of thunderstorms, prominent negative effects include the loss of lives and property. Thunderstorm-caused damage (e.g. tornado, hail, wind, lightning, and heavy rain) totaled $87 billion in the United States from 1949 to 1998 (Changnon 2001b). Tornadoes and severe thunderstorm winds have accounted for a combined 825 deaths from 1998 to 2007 within the continental United States (Schoen and Ashley 2011). Therefore, it is imperative that forecasts for convective events be accurate.

One issue in forecasting thunderstorms is knowing where and when convection initiates. How convection occurs is generally understood, but applying the concepts to accurately portray the timing and location of convective initiation with numerical weather prediction models (NWP) is still troublesome (Browning et al. 2007). In the convective initiation project IHOP (International H2O Project), the 10-km resolution RUC (Rapid Update Cycle) model predicted 44% of the events within 250 km and approximately three hours of the observed convection (Wilson and Roberts 2006).

The other issue regarding forecasting convection is knowing how the thunderstorms will evolve after initiation. Thunderstorms can take many forms throughout their evolution; each mode may influence the most likely type of severe weather. Individual cells, either arranged in a broken line or a cluster, often merge into solid lines (Bluestein and Jain 1985). Tornadoes and large hail are the main threats from cellular convection, whereas wind is the biggest threat from linear systems (Fujita 1978; Moller et al. 1994; Gallus et al. 2008). Similar to the initiation problem, representing atmospheric processes responsible for convective mode change within NWP models continues to be a challenge. Many factors need to be taken into account, such as wind shear orientation with respect to a surface
boundary, the relationship between low-level wind shear and the cold pool, and moisture availability and its effect on producing bowing segments (Rotunno et al. 1988; James et al. 2006; Dial et al. 2010).

The heart of the computer modeling problem lies most likely with grid resolution and the use of parameterization schemes. Coarse resolution without the use of a convective parameterization scheme (CPS) augments the chance of not producing convection (Kain et al. 2008). No particular scheme works best in every situation, though, as it has been suggested that different schemes may forecast precipitation more accurately (Wang and Seaman 1997; Gallus and Segal 2001). Recently, finer resolution NWP models have become more accessible, and the use of CPSs has come into question. In fact, multiple studies have found that models with fine-resolution grid spacing that treat convection explicitly predict rainfall more accurately than coarse-resolution models using a CPS (Done et al. 2004; Kain et al. 2006; Clark et al. 2009; Schwartz et al. 2010).

The primary goal of the present study is to investigate convective evolution with the Weather Research and Forecasting (WRF) model using radar observations and RUC (Rapid Update Cycle) analysis data for comparison. The WRF model used in the research is at a convective-allowing resolution (3-km grid spacing) that should not need a CPS. This study attempts to build upon previous observational and modeling studies’ results for determining factors of convective mode. Because initial environments could ultimately affect overall evolution, environmental parameters present at the time of convective initiation are considered in conjunction with synoptic and mesoscale features later in the evolution. Trends in simulated evolution are compared to observed trends, and selected case studies show in detail the most persistent issues with the WRF model’s evolution. Microphysics schemes have an effect on the precipitation produced, so the case studies also feature sensitivity testing to other microphysical schemes.

2. Thesis Organization

This thesis follows the journal paper format. Chapter 1 includes the general introduction to the thesis. Chapter 2 contains a literature review regarding relevant convective modes discussed in the present study. Chapter 3 is the paper that will be
submitted to *Weather and Forecasting*. Chapter 4 includes additional results from two extra case studies. Chapter 5 comprises general conclusions from the journal paper and the additional results. The final two sections entail acknowledgments and references.
CHAPTER 2. LITERATURE REVIEW

For over a century, researchers have classified deep moist convection into various types according to distinct characteristics. Frederick Starr (1887), professor at Coe College, noticed three prevalent types of thunderstorms through spotter reports: well-defined storms, heat storms, and squalls. The first category referred to convection associated with low pressure areas, usually located in the southeast quadrant. The heat storms, which accompanied hot weather, did not occur until the afternoon and evening hours and were isolated in nature. A squall referred to a windstorm that reached a “narrow, long, extended belt of land” at one time (Starr 1887).

The squall line was among the earliest recognized forms of convection, originating from French sailors describing lines with gusty winds (Talman 1907). French scientist Durand-Gréville noted that these squalls were associated with compacted isobars (Talman 1907). Wladimir Köppen (1879, 1882) discussed these systems in his publications regarding convective windstorms, referring to them as “thunderstorm squalls.” A long-lived, widespread wind event was termed “derecho” by Hinrichs (1888) in a way to differentiate from tornadoes (Johns and Hirt 1987). Fujita and Wakimoto (1981) redefined the derecho as including “a family of downburst clusters,” and Johns and Hirt (1987) assigned spatial and temporal criteria for the derecho: 1) a swath of wind gust reports of 26 m s\(^{-1}\) or greater with an axis length of least 400 km, 2) the reports must be located in a continuous and chronological path, 3) at least three wind reports of 33 ms\(^{-1}\) or greater separated by at least 64 km, and 4) no more than three hours between successive wind reports. It has been shown that derechos are as dangerous as some tornadoes, as they accounted for more deaths in the United States between 1986 and 2003 than F-0 and F-1 tornadoes combined (Ashley and Mote 2005).

Rotunno et al. (1988) suggested that squall line maintenance and strength depended on the orientation of the updraft; the more an updraft is tilted, the weaker the system. If the vorticity from the cold pool overpowers that of the shear, the updraft tilts rearward over the cold pool, and the convection ingests cooler, drier air. If the cold pool is absent, the updraft will tilt downshear, and evaporation from precipitation ahead of the squall creates an outflow.
In order for the updraft to be completely vertical, negative vorticity from the cold pool must balance the positive vorticity from low-level shear. Low-level shear was found to be the key to squall line longevity (Rotunno et al. 1988).

D. T. Williams (1948) documented precipitation patterns for seven squall lines, and noted that light and moderate rain followed the intense rain from the squall line for up to 45 minutes. In two of the cases, light rain preceded the convective line (Williams 1948). In a vertical cross-section of squall line structure and dynamics, Newton (1950) also indicated an expanse of lighter rain directly behind the main updraft and convection. This area of lighter, non-convective rainfall would later be referred to as stratiform rain. Characteristics and dynamics of the stratiform rain region were discussed in Smull and Houze (1985) using radar and satellite observations. They noted that front-to-rear flow within the convective system was critical in the formation of stratiform rain, and melting ice particles from the anvil produced the bright band appearance on radar (Smull and Houze 1985).

Houze et al. (1990) further defined the trailing stratiform rain region as large in size (at least 10 000 km²), displaying a concavity at the rear edge, and having a local maximum in reflectivity that is separated from the convective line. The authors also defined squall lines as either symmetric or asymmetric, depending on the location of the strongest cells within the convective line and the location of the stratiform rain region relative to the line. Symmetric “leading-line/trailing-stratiform” cases exhibited cell growth all along the line, and the centroid of the stratiform rain region was located behind that of the line. Asymmetric cases showed stronger cells on the southern or western edges of the line, and the centroid of the stratiform rain region is well to the north or east of the centroid of the line (Houze et al. 1990). The authors found that more reports of tornadoes and hail tended to be associated with asymmetric systems and in the southernmost cells, as this environment featured greater along-line shear. That environment was also typical for supercell development (Houze et al. 1990).

Not all squall lines show trailing stratiform rain regions, however. Schiesser et al. (1995) created classifications for lines with leading stratiform rain and lines with no stratiform rain at all. Parker and Johnson (2000) proposed nomenclature for linear MCSs with parallel stratiform rain, where the stratiform region moves parallel to the line and very
little precipitation surrounds the convective line itself. Deep and strong flow parallel to the line was present in cases with parallel stratiform rain, and those systems often transitioned to lines with trailing stratiform rain. Through radar analysis, Schumacher and Johnson (2005) identified two additional types of linear systems when studying extreme rain events. Training line/adjoining stratiform systems exhibited a line of cells, usually zonally oriented, training along a boundary with a stratiform rain region to the north. Back-building/quasi-stationary systems featured new cells consistently forming upstream while decaying cells moved downstream. They also noted that stratiform rain was produced in some cases, with a similar appearance as the line with parallel stratiform rain classification by Parker and Johnson (2000).

Bluestein and Jain (1985) created a classification scheme for the development of squall lines by using four initial convective modes: broken line, back building, broken areal, and embedded areal. Broken line formation began as a discrete cells arranged in a line with new cells forming between old cells, eventually merging into a solid line. This formation usually occurred along a cold front with a narrow band of forcing. Back building consisted of new cells initiating upstream of old cells, a process documented to last up to six hours. This situation also occurred in an environment very favorable for supercells. Broken areal began as cells unarranged in a line transitioning into a squall line, most likely from outflow boundary interaction. Embedded areal consisted of stronger cells forming a line within a broader area of lighter precipitation.

Schiesser et al. (1995) classified systems, deemed cell complexes (CC), according to the orientation of convective cells within a group or line. Cell complexes comprised isolated, group, broken line, and continuous line systems. Isolated CC was a single cell with reflectivity of at least 55 dBZ within a region of 40 dBZ reflectivity. The group CC contained multiple cells, not in a linear pattern, connected by at least 25 dBZ reflectivity. Broken line CC had multiple cells of at least 40 dBZ, arranged in a line, separated by an area of 25 dBZ, while the continuous line CC had an area of 40 dBZ reflectivity surrounding the cells. The lines must have also attained a 3:1 length-width ratio. Schiesser et al. (1995) found 3-10 km shear was substantially larger for isolated CC compared to other categories.
A special type of squall line first identified by Nolen (1959) was the “line echo wave pattern,” in which a portion of the line accelerated forward while adjacent portions maintained the original propagation speed or even slowed down. The concept was refined by Fujita (1978), where he termed the system a “bow echo.” Fujita (1978) described the morphology of the bow echo as having three main stages: 1) large/strong/tall echo, 2) bow echo, and 3) comma echo. The second phase, in which the squall line was most intense, featured a “spearhead” at the apex of the bow, a cyclonically rotating head, and an anticyclonic tail (Fujita 1978). He theorized that a strong rear-inflow jet was responsible for the bowing structure and the strong winds. Bowing segments were also noted in derechos by Johns and Hirt (1987), who found that they were the focal points of the strongest downbursts. The rear-inflow notch (RIN), where the rear flank downdraft surges toward the bow, also created damaging winds and downbursts (Przybylinksi 1995).

Weisman (1993) explained the formation of the bow echo in four stages. The updraft initially tilts downshear, but after a cold pool is generated, the updraft becomes erect. Once the vorticity from the cold pool overpowers that of the shear, the updraft tilts rearward. The interaction between negative vorticity from the front-to-rear flow in the anvil and positive vorticity from the back edge of the cold pool generates the rear-inflow jet. The rear-inflow jet, in addition to the shear, balances the cold pool circulation, which results in a less-tilted updraft and a stronger system.

Regardless of the orientation of cells within the squall line and location of stratiform rain, all squall lines can be grouped into a single classification called mesoscale convective systems (MCSs), a term created by Maddox (1980). MCS describes convection that has a length scale of 250 to 2500 km and a time scale of greater than six hours. Squall lines fit into the linear type of MCSs, while mesoscale convective complexes (MCCs) are a circular type of MCS. MCCs have their own set of criteria in addition to the general MCS criteria: 1) a cloud shield of at least 100,000 km² of infrared temperatures colder than -32°C, 2) a cloud shield of greater than 50,000 km² colder than -52°C, and 3) eccentricity of 0.7 or greater at time of greatest maturity (Maddox 1980).
CHAPTER 3. PREDICTION OF CONVECTIVE MORPHOLOGY IN NEAR-CLOUD PERMITTING WRF MODEL SIMULATIONS

A paper to be submitted to Weather and Forecasting

Darren V. Snively and William A. Gallus, Jr.

3.1 Abstract

The ability of the Weather Research and Forecasting (WRF) model to forecast convective morphology evolution is examined for 37 convective systems. The simulations included Thompson microphysics and 3-km grid spacings. Ten convective mode classifications were used. An objective score was developed to determine the accuracy of the simulated morphologies considering a normalized duration of each mode simulated and its agreement with observations. Rapid Update Cycle (RUC) analyses were used to find possible relationships between larger-scale pre-initiation conditions and simulated morphology accuracy. Two case studies selected as representative of the most common simulated morphology deficiencies were examined in more detail.

The modeled convective modes exactly matched observations in 31% of the comparisons, and shared the same classification group (i.e. cellular or linear) in 56% of the comparisons. The model simulated cellular systems relatively well but struggled more with linear systems, particularly bow echoes and squall lines having trailing stratiform rain regions. Morphological evolution was generally better simulated in environments with enhanced deep-layer shear and cool potential temperatures at the level of maximum theta-E. Weaker deep-layer shear, cool potential temperatures at the surface, and quickly warming potential temperatures with height increased the likelihood of timing errors. The first case study showed that a warmer cold pool, much larger line-normal shear, and excessive mid-level drying were present in the model run that failed to develop a trailing stratiform region. The second case study showed that weak shear and a lack of a well-developed cold pool may have played a role in the lack of bowing.


3.2 Introduction

Forecasting convection remains a challenge for meteorologists. Most past efforts toward improving forecasting of convection have focused on quantitative precipitation forecasting (QPF) (Olson et al. 1995; Wang and Seaman 1997; Gallus 1999; Alhamed et al. 2002; Gallus and Bresch 2006), finding that in models with grid spacings coarse enough to require use of a convective scheme, QPF is profoundly sensitive to the convective scheme used (Gallus and Segal 2001; Jankov et al. 2007). Clark et al. (2009) noted that convection-allowing models (e.g. 4-km horizontal grid spacing) forecasted the timing and location of precipitation better than models using convective parameterizations, particularly when mesoscale convective systems (MCSs) were involved. Improved computational resources are allowing such fine grid spacings to be used increasingly often.

The use of convection-allowing grid spacings results in simulated systems having some fine-scale structures similar to those observed with radar (Kain et al. 2006; Kain et al. 2008). Work is only just beginning to examine how well models simulate morphology (Fowle and Roebber 2003; Done et al. 2004; Weisman et al. 2008; Schumann and Roebber 2010). Done et al. (2004) showed the Weather Research and Forecasting (WRF) model failed to develop stratiform rain regions on multiple occasions, and Fowle and Roebber (2003) explained how forecasts of mode were not as accurate when substantial large-scale forcing features were absent. Weisman et al. (2008) showed that higher resolution models were valuable to forecasters for predicting convective mode, and Schumann and Roebber (2010) studied how upper-level forcing and potential vorticity patterns promoted multicellular convection. Additional work is needed to understand how well models depict evolution of convective morphology. A situation that is experienced frequently is upscale growth from single cell systems into multicell systems (Jirak and Cotton 2007).

Convective mode classification can help in understanding the behavior of observed and simulated weather systems and potentially in the forecasting of several hazards. Gallus et al. (2008), for instance, found that significantly severe hail (2”+ in diameter) was most common from cellular storms, while the main hazard from linear systems was wind. Some prior studies have classified convection into three broad groups: cellular, linear, and mixed
systems (Dial and Racy 2010; Grams et al. 2012). Other studies that focused primarily on one general system type, such as linear, divided the systems according to the location of a stratiform rain region (Bluestein and Jain 1985; Parker and Johnson 2000; French and Parker 2008). Additional studies have split both cellular and linear convection groups into more detailed categories based on the orientation of the individual cells and stratiform region location (Gallus et al. 2008; Duda and Gallus 2010).

One convective mode can evolve into another by a variety of factors. French and Parker (2008) showed that large-scale environmental parameters (shear, instability, etc.) alone do not differentiate between convective modes, but localized differences of these parameters in combination with synoptic features help determine mode evolution. James et al. (2006) concluded bow echoes tend to form in not overly dry or moist environments at the low levels and when the cold pool overwhelms the shear in a small area instead of along the entire line. The orientation of wind flow and wind shear with respect to surface boundaries has been shown to be influential factors in the formation and maintenance of linear systems. In the absence of a synoptic-scale forcing mechanism, such as a surface front, multicellular convection is more likely with higher wind shear (Schumann and Roebber 2010). Rapid evolution from cellular convection into a linear system can occur if mid-level flow is approximately parallel to a surface boundary due to merging cold pools and precipitation regions of the cells (Dial and Racy 2004). When the magnitudes of the vorticity created by low-level shear and the cold pool are nearly equal, storm updrafts tend not to tilt, and the longest-lived squall lines are observed (Weisman et al. 1988). Rear-inflow jets within linear systems, depending whether they are near the surface or are elevated, produce additional circulation within the cold pool (Weisman 1992). Weisman et al. (1988) and Weisman (1992) also showed that the circulation or shear affects the tilting of the updrafts, and in turn, the extent of the stratiform rain region. How fast a cellular system becomes linear is positively correlated with the amount of deep-layer forcing, even more than the orientation of the wind vector in the cloud layer (Dial et al. 2010). Schumann and Roebber (2010) agree that increased forcing tends to favor multicellular convection over individual cells because of widespread atmospheric destabilization.

The present study seeks to understand the predictability of convective evolution,
while making use of the aforementioned findings. Wind shear and orientation, moisture, and synoptic forcing are compared between the WRF model simulations and observations. Two detailed case studies are performed to highlight some of the most common discrepancies seen in the WRF simulations compared to observations. Section 3 explains the convective mode classification scheme and the scoring method used to determine the accuracy of the simulations. Section 4 discusses the results of the convective mode comparisons and the accuracy rating analysis. Section 5 contains the case studies, and section 6 discusses the results and the overall conclusions.

3.3 Methodology

3.3.1 Convective mode classification

Morphology classification was performed for 37 events occurring during the warm season from 2006 to 2010 primarily in the United States Great Plains and the Midwest. The dates of the events can be seen in Figs. 1. WRF version 3.1.1 using the Advanced Research WRF (ARW) dynamics core (Skamarock et al. 2008) was used to simulate the events. The model runs used the Thompson microphysics (Thompson et al. 2008), Mellor-Yamada-Janjic (MYJ) planetary boundary layer (Mellor and Yamada 1982; Janjic 2002), Monin-Obukhov surface layer (Janjic 2002), Noah land surface (Ek et al. 2003), Rapid Radiative Transfer Model (RRTM) shortwave radiation (Mlawer et al. 1997), and Dudhia longwave radiation (Dudhia 1989) schemes. Initial and lateral boundary conditions for the simulations were supplied by North American Model (NAM) 12 km analyses, and the WRF was run with 3-km horizontal grid spacing. The WRF-default of 40 vertical levels was used. For the majority of the events, the model was initialized at 1200 UTC and integrated for 24 hours. Simulated composite reflectivity output in 15 minute increments was used to determine morphology and its evolution.

Convective modes were identified throughout each event’s evolution using nine classifications from Gallus et al. (2008), along with an additional classification developed for the present research. Three modes comprised the cellular group: individual cells (IC), clustered cells (CC), and broken line (BL). Five modes represented linear systems: those with no stratiform rain region (NS), trailing stratiform (TS), parallel stratiform (PS), leading
stratiform (LS), and bow echo (BE). Another mode was nonlinear (NL). The assignment of the new classification, mixed-complex (MC), was reserved for situations that exhibited characteristics of two or more of the aforementioned convective modes. Convective initiation was defined as the first instance of 40 dBZ reflectivity, and the system had to maintain 40 dBZ reflectivity in order to retain a classification. The minimum length of the convective portion for the linear systems was set at 75 km. Stratiform regions were defined as regions of at least 30 dBZ reflectivity at least twice as wide as the nearby convective lines. The 30 dBZ criterion for stratiform rain is similar to that used by Hilgendorf and Johnson (1998). The event must have shown characteristics of a particular convective mode for a minimum of two hours to receive that mode’s classification. The system must also have initiated within 300 km of the observed system in order to be classified. The mosaic radar archive from the University Corporation of Atmospheric Research (UCAR) website [http://locust.mmm.ucar.edu] was used to analyze convective evolution of the observed systems. The same criteria for classification were also applied to the simulated events.

3.3.2 Scoring method for WRF accuracy

Events were given a score based on how well the simulated convective mode matched the radar observations. Time was normalized with convective initiation being set to zero, dissipation set to one, and the duration of each convective mode being set to its respective portion of the event’s lifetime. For example, if the observed system’s duration was 12 hours, and IC was observed for the first three hours, NS for the following three hours, and MC for the last six hours, the normalized time scale would be computed as IC during 0 to 0.25, NS from 0.25 to 0.50, and MC from 0.50 to 1. If the system did not dissipate by the end of the model run (usually 1200 UTC), the end of the model run was defined as 1.0. If the system moved out of the domain, the time at which convective mode could no longer be classified was defined as 1.0. Two time normalizations were performed for each event: one based on radar observations and the other on the simulation. If the simulated event’s initiation or dissipation occurred more than three hours different from the observed, a penalty was introduced through an adjustment to the WRF’s time scale, thus reducing the maximum possible score. If the simulated system initiated earlier or dissipated later than the observed
system, the fraction of the lifespan outside the grace period was not considered for comparison to the observed system, thus earning a score of 0 (effectively a comparison of “no system” to the simulated mode). If the simulated system initiated later or dissipated earlier than the observed system, the time between the grace period and the dissipation/initiation was added to the overall simulated timespan and not considered for comparison so that it again was scored as 0.

The two time scales were then merged by partitioning the event’s time scale whenever a convective mode changed in either the observations or the simulation. If the simulated system displayed a broken line (BL) for the first six hours, clustered cells (CC) for the following three hours, and mixed-complex (MC) for the last three hours, its time scale would be computed as BL from 0 to 0.50, CC from 0.50 to 0.75, and MC from 0.75 to 1. When this time scale is combined with the observed time scale from the earlier example, the first convective mode change would occur at 0.25, so from 0 to 0.25, a comparison is made between the observed IC and the WRF simulated BL. The next change occurred at 0.50 (in both the observations and the simulation), so from 0.25 to 0.50 the comparison is between the observed NS and the simulated BL. This method was followed until normalized time 1.0.

Scores were then computed using both general group matches and more detailed morphology matches. For a group match, the WRF-simulated general morphology (cellular, linear, or nonlinear) had to match that observed, even if the specific morphologies differed (e.g., IC versus BL). A detailed match occurred when the specific morphologies (the 10 types) matched exactly. The portion of the event’s lifespan where matches were identified was summed to calculate the total accuracy score. Group matches were awarded a 0.5 score and detailed matches a 1.0 score during the time they were observed. For example, if half the event showed group matches and the other half showed detailed matches, the total score would be 0.75.

3.3.3 Initial environment and case study criteria

Hourly RUC (Rapid Update Cycle) analyses from the National Climatic Data Center [http://nomads.ncdc.noaa.gov/data/] and the Atmospheric Radiation Measurement Program archives [ftp://ftp.archive.arm.gov] were used to determine observed environmental
conditions occurring at the time and general location of convective initiation for the 37 cases. Observations were taken at the centroid of the observed system approximately one hour before convective initiation to obtain surface-based and mixed layer CAPE, potential temperature near the surface (using the 0-30 hPa above ground level average) and at the level of maximum equivalent potential temperature, and 0-3 km and 0-6 km bulk shear.

Environmental conditions later during convective evolution were studied in more detail through two case studies representing the most frequently observed morphology errors. The WRF model output was regridded from 3-km to 20-km horizontal resolution and from sigma levels to standard pressure levels to match the RUC analysis for direct comparison of parameters. A program from the National Centers for Environmental Prediction (NCEP), the Unified Post-Processor (UPP), was used for bilinear interpolation of the model variables. The first case examined, 23-24 May 2006, featured an observed TS event with a simulation of NS, and the model also failed to develop any stratiform rain during its entire lifespan. The second case, 26-27 May 2006, featured an observed bow echo with the WRF unable to produce bowing segments within the linear system. Instead, the WRF simulated isolated cells eventually forming a squall line system without stratiform rain. At the end of its evolution, the simulated line evolved into a non-linear system.

3.4 Results

The frequency of each classification for observed and simulated systems is shown in Table 1. Overall, radar observations contained six more convective modes than the model (115 and 109, respectively). The WRF produced the same number of BL (24), PS (1), and LS (0) systems and nearly the same number of IC, NL, and MC systems as was observed. The largest differences in counts occurred with BE events (13 fewer simulated than observed), TS (6 fewer), NS systems (5 more), and CC events (9 more). In general, the WRF forecasted too many cellular systems (48 versus 37) and too few linear systems (44 versus 58). The WRF model simulated a linear mode at one point in each system’s evolution in 26 of the 34 events that featured an observed linear system. Of the 21 events that included an observed bow echo, the WRF captured that mode at some point in the systems’ evolution in only eight events. The model also had trouble simulating TS systems, only showing TS in 7
of 19 observed TS events. Lack of TS rain regions in the 37 WRF simulations is consistent with the findings from Done et al. (2004). There were also six instances of the model failing to produce a BL event.

### 3.4.1 Match accuracy

Detailed and group matches are outlined in Table 2, and convective mode comparisons are shown in Table 3. Of the 185 comparisons, only 58 (31%) consisted of a detailed match, with 104 (56%) resulting in a group match (Table 2). The model was the most accurate in simulating cellular systems, with approximately half of the comparisons for these three classifications yielding detailed matches. BL was the most matched mode of all classifications with 17 of its 34 comparisons resulting in a detailed match (Table 2). The majority of the BL comparisons that did not match tended to be cases when the model instead showed NS (8 occurrences) or CC modes (6 occurrences) (Table 3). The former discrepancy was usually due to a model timing error, and the latter discrepancy usually was present during the initial mode. When accounting for group matches, the accuracy increased to over 80% for IC and CC events and 70% for BL events (Table 2).

Linear systems were not simulated as well by the WRF model, with only 16% of bow echoes and 24% of trailing stratiform systems being correctly simulated (7 detailed matches each) (Table 2). Common model errors for BE events included simulation of NS (10 occurrences) or TS events (8 occurrences) (Table 3). When a TS system was observed, the model simulated a NS event eight times and a BL event five times. Nearly 60% of BE and TS comparisons resulted in group matches, suggesting the model did simulate a linear system when BE or TS events were observed. However, in only 42% of observed NS events did the model indicate some linear mode (Table 2). Besides the eight detailed matches for NS events, there were seven events where the model showed BL and four showing IC. This suggests the model may have more difficulty producing linear systems when no stratiform rain or bowing exists in the observed system (Table 3).

Approximately one-third of the observed non-linear systems were simulated by the model (10 matches). For the remaining NL events, the model modes were evenly split between cellular and linear modes. Simulations of mixed-complex systems were especially
poor, with only one out of the six observed events correctly forecasted.

The normalized timescale scoring method resulted in an average score of 0.49 for the 37 cases, with 15 cases receiving a score of at least 0.50. The best score was a perfect 1.0 for one case of a TS existing the entire time. Of note, the next highest score occurred for a case evolving through four different modes. The worse score was 0.03 for a case with a major delay in initiation and a simulation of IC the entire time while a linear system was observed. In nine cases, the model initiated or dissipated convection more than three hours apart from the observed times, and those cases had an average score of 0.32. The average accuracy scores for the cases where the WRF model completely failed to simulate an observed BE or TS event were 0.42 and 0.44, respectively. The WRF correctly simulated the first convective mode in 13 of the 15 cases with a score of 0.50 or greater and 12 of the 22 cases with a score less than 0.50.

3.4.2 Relationship of mode accuracy to environmental conditions

The Wilcoxon rank-sum test was applied to the cases’ mode accuracy scores and observed pre-storm environmental parameters to determine statistical significance of differences in the large-scale conditions present for cases simulated accurately and those not. The test showed cases with scores under 0.50 tended to have significantly warmer potential temperatures at the level of maximum theta-E and significantly lower 0-6 km bulk shear values than cases with scores over 0.50 with at least 90% confidence (Table 4). The average deep-layer shear for the cases where the model performed relatively well was 22.4 m s\(^{-1}\) compared to 17.8 m s\(^{-1}\) for cases simulated more poorly. Assuming the amount of shear reflects the large-scale forcing, this result implies mode might be better predicted for cases with stronger larger-scale forcing than weaker forcing, a result found to be true for QPF in convection-parameterized model runs (Jankov and Gallus 2004). The average potential temperature at the level of maximum theta-E was 3.3 K warmer in the low-scoring cases than the better scoring cases. The maximum equivalent potential temperature was on average 48 hPa above the surface for the cases with scores greater than 0.50, which was nearly 30 hPa lower than the other cases. This result implies that forecasting of mode is worse when a
stable layer near the ground is deeper, as would be the case for elevated convection. Although not statistically significant, the average difference between surface-based CAPE and mixed-layer CAPE was nearly 450 J kg\(^{-1}\) larger for the cases with scores under 0.50 versus cases with scores over 0.50. No relationship was found between accuracy scores and surface-based CAPE or 0-3 km bulk shear.

Comparisons were also made between cases that had a major timing error and cases that stayed within the three hour initiation or dissipation thresholds. Cases that violated the timing criterion had significantly lower (99% confidence) 0-6 km bulk shear values compared to cases with no timing issues (Table 4). The badly-timed events had an average shear value of 12.9 m s\(^{-1}\) versus 21.9 m s\(^{-1}\) for the well-timed cases. The rank-sum test also showed moderate evidence that cases with major timing errors had lower potential temperatures at the surface and steeper potential temperature lapse rates. Here, the potential temperature lapse rate is defined as the increase of potential temperature with decreasing pressure (-\(\partial \theta / \partial p\)) from the surface to the level of maximum theta-E. The cases with a timing error had on average a 3.6 K lower temperature near the surface and a 2.5 K Pa\(^{-1}\) steeper lapse rate than the cases with no timing error, indicating greater stability near the surface prior to convective initiation raises the likelihood of major timing errors within the model. Cases that violated the timing criterion had an average mixed-layer CAPE value nearly 1000 J kg\(^{-1}\) higher than the cases with no error (95% confidence). The average difference between surface-based CAPE and mixed-layer CAPE was much higher for the cases with large timing errors, a result also seen with the low-scoring cases. However, the increase in statistically significant results for the timing issue cases may be due to a small sample size for these cases.

These results suggest the model predicts convective evolution more accurately when deep layer shear is relatively high. The large difference between mixed-layer and surface-based CAPE and between potential temperature at the surface and aloft for the low-scoring cases show a stable layer present near the ground, in these events with convection likely being elevated. The higher level of maximum equivalent potential temperature in the bad cases supports that result. The model also appears to be more prone to delayed initiation or early dissipation when elevated convection exists.
3.5 Case Studies

Two case studies were performed for events representing the two most common types of errors observed, to investigate in more detail the potential causes of the differences in convective evolution between the WRF simulations and observations. Synoptic conditions and mesoscale environments were explored in detail when a convective mode changed either in the observations or simulations. In addition, because it has been shown that precipitation accumulation among ensemble members can be a function of the microphysical scheme used, e.g., Jankov et al. (2005) and Schwartz et al. (2010) and thus morphology might vary, microphysical sensitivity tests were also performed. For example, Gallus and Pfeifer (2008) noted that reflectivity was underestimated in the convective portion of a squall line when the Lin scheme was used, and the Thompson and WSM6 schemes overestimated reflectivity within the stratiform anvil. Morrison et al. (2009) found notable differences in trailing stratiform rain production between the one-moment and two-moment microphysical schemes. In these tests, the Thompson et al. (2008) microphysics scheme was replaced by the WSM6 (Hong and Lim 2006), Lin et al. (1983), and Morrison et al. (2005) schemes. These microphysics schemes are widely used with research pertaining to the WRF model. WSM6 and Lin are single-moment schemes, meaning mixing ratios of the hydrometeors are considered, while Morrison is a double-moment scheme, in which the number of concentration of hydrometeors is also considered. Because forecasts are also sensitive to initial and lateral boundary conditions (Jankov et al. 2007; Weisman et al. 2008), an additional test was performed for each case study where in the control run, GFS data were used for initial and lateral boundary conditions instead of NAM.

3.5.1 Case study 1: 23 May 2006 – 24 May 2006

For the 23-24 May 2006 squall line system having a trailing stratiform region, the WRF model simulated a line without any stratiform rain. It was shown earlier that this error occurred in 42% of all TS events. The model received an accuracy score of 0.62 for this event, mainly because it did capture a linear mode (Table 5), albeit without stratiform rain. RUC analyses at 1800 UTC showed little difference between surface-based CAPE and mixed-layer CAPE (131 J kg\(^{-1}\)), and the maximum equivalent potential temperature was only
47 hPa above the surface, which suggests this event began as surface-based convection. Low-level instability (lapse rate of 2.7 K Pa⁻¹) and appreciable deep-layer shear (20.7 ms⁻¹) were present, with the values being consistent with the good-scoring cases (Tables 4).

An occluding cold front stretched from a low pressure center in eastern Montana southward through Kansas and was the triggering mechanism for this event. The front was occluded from central Nebraska northward to the low pressure area, while no occlusion occurred south of Nebraska. The upper-level low at 500 hPa associated with the system deepened during the period, with heights falling from 5700 m at 0000 UTC on 24 May to 5640 m at 1200 UTC (not shown).

A broken line of convection initiated at 1900 UTC on 23 May in north-central Nebraska and south-central South Dakota ahead of the frontal system (Fig. 2a). The model predicted convective initiation well (Fig. 2b) by simulating the correct mode less than 100 km to the east just one hour later than the observed event (2000 UTC). Bulk shear in the 0-6 km level was comparable between the WRF output and the RUC analysis. A nose of higher readings (25 to 30 ms⁻¹) was located behind the front in western Nebraska and oriented southwest to northeast, while near-storm and pre-storm environment shears were approximately 15 to 20 m s⁻¹ (Fig. 3).

At 2330 UTC, both the observed and simulated systems evolved to a line with no stratiform rain region (Fig. 2). The RUC analysis showed an increase in deep-layer shear in central Nebraska that the WRF model did not predict, but the shear vectors were mainly parallel to the boundary for both systems, which helped consolidate the cells into a line (Fig. 3). The model also depicted correctly placement of low moisture content behind the front, keeping precipitable water values under one inch in the western half of Nebraska and showing negative moisture advection at 850 hPa directly behind the system in South Dakota (not shown).

However, at 0330 UTC on 24 May, the observed system developed a trailing stratiform rain region, while the WRF simulation continued to depict a line with no stratiform rain (Fig. 2). The highest values of 0-6 km bulk shear in the RUC analysis were located in Kansas, and most of the near-storm environment shear in eastern Nebraska was 15 to 20 m s⁻¹ (Fig. 3e). The WRF showed a shear maximum of 30 m s⁻¹ along the line, but values
quickly dropped off behind the line to 15 ms$^{-1}$ (Fig. 5f). Negative moisture advection at 850 hPa lagged well behind the system in the RUC analysis, but the negative advection was modeled farther east by the WRF output (Fig. 4). This pattern was also evident in the 850 hPa relative humidity in the WRF model, with values as low as 30% in eastern Nebraska behind the convective line (Fig. 4d). At the surface, the RUC analysis showed a fully-developed cold pool with potential temperatures perturbations of approximately 6 K, but the WRF model depicted a less pronounced cold pool with potential temperature perturbations of approximately 3 K. A divergent pattern can clearly be seen in the 10 m winds for much of the area with cooler potential temperature values in the RUC analysis (Fig. 5a). However, in the WRF model output, the diverging surface winds are confined to a much smaller area (Fig. 5b).

A vertical cross-section of the component of storm relative wind normal to the line was taken with the original 3-km WRF output to examine the cold pool-wind shear interaction (Fig. 6a). The erect updraft with little tilt was consistent with Rotunno et al. (1988) when a vorticity balance between the cold pool and low-level shear was present. The front-to-rear flow at anvil level suggests hydrometeors should be transported to the rear of the system (Smull and Houze 1985), and drier air should be advected in from the rear-inflow jet. The jet showed a maximum storm relative flow of approximately 15 m s$^{-1}$ near 500 hPa, consistent with “strong rear inflow” cases defined in Smull and Houze (1987). The layer of positive storm relative flow was 5 km thick directly behind the main updraft, approximately 2 km thicker than the mean “strong rear inflow” profile for lines with trailing stratiform precipitation in Smull and Houze (1987). A small area of rising motion was present behind the convective line where the trailing stratiform region would normally be located, but it intersected an area of 20% relative humidity in the rear-inflow jet (not shown). A simulated atmospheric sounding taken at the point of maximum front-to-rear flow showed a very dry layer in the mid-levels of the atmosphere (Fig. 6b). Any hydrometeors falling into this region would most likely evaporate before reaching the surface. Evaporation likely played a role in the lack of trailing stratiform rain region in the WRF model.

Dial et al. (2010) differentiated between systems that developed trailing stratiform rain regions within three hours of initiation and those that do not by looking at the 2-6 km
shear normal to the boundary. That method was followed similarly in this study, but by using a point where stratiform rain was observed on radar. The normal component of the 2-6 km shear relative to the boundary was computed ahead of the main frontogenetical zone at roughly 50 km increments. The WRF output had an average of 5.2 m s⁻¹ of shear normal to the boundary, while RUC analyses showed an average of 1.2 m s⁻¹. Dial et al. (2010) found that systems with trailing stratiform rain regions had smaller shear values, and thus this result is consistent with the lack of a trailing stratiform region in the WRF model.

By 0900 UTC on 24 May, the observed system transitioned to a nonlinear system, while the WRF simulated a line with no stratiform rain (Fig. 2g, h). At this time, the RUC analysis showed a broad region of 15 to 20 m s⁻¹ of deep-layer shear with a local maximum of 20 to 25 m s⁻¹ over central Iowa (Fig. 3g). The WRF output depicted a maximum of 25 to 30 m s⁻¹ of shear along the leading edge of the line and smaller values extending west toward the Missouri River (Fig. 3h). The broader area of higher shear in the WRF model output than observed may have provided enough support to keep an organized linear system simulated.

Because differences in convective mode in a simulation compared to observations could be the result of deficiencies within the microphysical scheme used and not necessarily due to differences in environmental conditions, three additional microphysics schemes were used in sensitivity tests. The sensitivity tests in the present case study resulted in little change in reflectivity and convective mode for the first several hours compared to the original simulation. All three additional model runs simulated BL at approximately 2000 UTC evolving to NS by 2300 UTC. However, the test simulations did produce a system with equal width and length of high reflectivity values (>60 dBZ) near 0600 UTC, prompting an NL classification, however, a result different from the control and matching observations at these later times. All simulations failed to produce a trailing stratiform rain region suggesting this problem was either not primarily related to a deficiency of the schemes, or that the same deficiency is present in all four schemes tested.

Because forecasts are sensitive to initial and lateral boundary conditions (Jankov et al. 2007; Weisman et al. 2008), one test was performed using GFS data instead of NAM for initialization and lateral boundary conditions. The Thompson et al. (2008) microphysics scheme was used for direct comparison to the original simulation with NAM initial
conditions. The GFS-initialized simulation did indeed produce a TS region for the northern half of the line at 0000 UTC, and maintained the stratiform region for several hours. Analyses at model initialization (1200 UTC on 23 May) showed that 0-6 km and 2-6 km shear and surface-based CAPE and CIN were similar between the RUC and WRF run using GFS data. The WRF run using NAM data underestimated 0-6 km bulk shear and CAPE and overestimated 2-6 km shear in the region of the convective initiation. Upon transition to TS in the GFS-initialized simulation, the 2-6 km line-normal shear ahead of the main frontogenetical forcing for the northern portion of the line was half the magnitude in the NAM-initialized WRF simulation. The line of intense reflectivity (50+ dBZ) was also more continuous than in the NAM-initialized run, reflecting the more unstable environment present at model initialization. The stronger convection in the GFS-initialized simulation may have altered the flow in the near-storm environment, weakening the line-normal component of the mid-layer shear, so as to better allow development of a TS system.

3.5.2 Case study 2: 26 May 2006 – 27 May 2006

In the 26-27 May 2006 case, the WRF model initiated individual cells that grew into a linear system without a stratiform rain region, but the radar showed a bow echo for much of the system’s life. The modeled line also transitioned to a nonlinear system late in its lifespan while the radar continued to display a bow echo. The WRF model received a score of 0.24 for this event due to errors outlined in Table 6. Convection initiated with nearly 3500 J kg\(^{-1}\) of surface-based CAPE and 3900 J kg\(^{-1}\) of mixed-layer CAPE. Deep-layer shear was below the 37-case average at 18 ms\(^{-1}\), and there was very little change in potential temperature (2 K) from the surface to the level of maximum equivalent potential temperature, which was 90 hPa above ground level.

A surface low pressure located in southeastern Colorado was moving northeast into the central Great Plains, and a lee trough extended from the low pressure south into Texas and New Mexico. A warm front traversing Kansas and Missouri slowly moved north across the region and was dissipating. Rawinsondes from 0000 UTC on 27 May, around the time of the convective initiation of interest, indicated a 500 hPa trough over the Pacific Northwest and a ridge centered over the western Great Lakes, placing the study area in southwest flow.
A remnant MCS was located across north-central Kansas and south-central Nebraska at the beginning of the study period, and it traveled southeast before dissipating near Topeka, Kansas around 1900 UTC. The simulation also showed the ongoing MCS, but it allowed the convection to stay intact well into Missouri.

The first cell that would quickly become a cluster was observed on radar at 2230 UTC in northwest Kansas, roughly 200 km northeast of the surface low (Fig. 7a). Radar also showed some convective cells in eastern Colorado, but that convection was not associated with the upscale evolution into the bow echo. The RUC analyses showed a weak band of frontogenesis (not shown) associated with the warm front along the Nebraska-Kansas border extending to the convection. A stronger band of frontogenetical forcing was present near the low pressure in southeast Colorado, and moisture at 700 hPa (not shown) was being advected northward toward the observed cells. Cellular convection was underway two hours earlier in the simulation along the Front Range of Colorado, and the ongoing MCS was located in east-central Kansas still attaining reflectivity values of 65 dBZ (Fig. 7b). The simulation depicted approximately 50% stronger frontogenesis (not shown) with the warm front, but the cells that initiated in Colorado were not impacted by that forcing. Surface frontogenesis was also strong near the ongoing MCS. The simulated convection in Colorado initiated in an environment with cooler dew point temperatures (10ºC) than the observed convection in northwest Kansas (20ºC).

By 0300 UTC on 27 May, the observed CC had grown upscale into a line and had quickly formed a bowing segment (Fig. 7c). The WRF model, however, maintained individual cells into northwest Kansas and southwest Nebraska. A few of the simulated cells merged, while others remained discrete during the time frame. Because any merging lasted less than two hours, the classification of the system remained IC. The modeled system moved east-northeast into southern Nebraska, consistent with a southwest 2-6 km mean wind. This trajectory helped steer the cell into an area with negligible forcing and frontogenesis (Fig. 8a). The primary band of frontogenesis associated with the warm front was also located too far south in the simulation. The RUC analysis showed the frontogenesis along the Nebraska-Kansas border and stretching into the Kansas City area (Fig. 8b). The RUC 2-6 km mean flow also had a larger westerly component than what was indicated in the WRF
output, which allowed the observed convection to ride along the warm front and to develop into a bow echo.

By 0500 UTC, convection had quickly developed in eastern Nebraska north of the ongoing bow echo. The new cells merged into the bow echo approximately 90 minutes later to expand the bow echo from Sioux City, Iowa, to Salina, Kansas. TS rain was also observed with the bow echo. The simulation quickly developed new cells in eastern Nebraska at 0600 UTC, and they merged into a line within an hour; thus, the system was classified as NS (Fig. 7d). No significant features conducive for convective initiation in the immediate area of the newly formed line were found at the surface. The nearest outflow boundaries were 200 km to the west, and the WRF output showed no discernible variation in surface temperature or dew point in eastern Nebraska. Simulated atmospheric soundings near the convection confirmed the origin of the thunderstorm parcels was near 750 hPa, approximately 200 hPa above ground level (Fig. 9). Extensive drying was evident in the layer from the surface to 750 hPa in addition to a small inversion directly above the surface. Despite the stable conditions near the surface, most unstable CAPE values reached 1300 J kg\(^{-1}\). Higher potential temperatures were advected into the region from central and western Nebraska by 20 m s\(^{-1}\) 700 hPa flow. Much of western Iowa was under much weaker 700 hPa flow (5 to 10 m s\(^{-1}\)). A strip of higher 0-6 km bulk shear values existed over much of southern Iowa and extreme eastern Nebraska, with values of 15 m s\(^{-1}\) (not shown). Positive moisture advection in combination with speed convergence and sufficient elevated instability over eastern Nebraska was adequate for rapid convective initiation. These results agreed with the 0600 UTC RUC analyses well. However, the simulation failed to produce a bowing segment within the line. A cross section of the storm relative zonal wind revealed the lack of both front-to-rear flow and a rear-inflow jet in this system (not shown). The more isolated nature of the initial cells may have affected the system’s ability to produce stratiform rain and a bowing segment when the event became more linear. The observed CC system had more cell interaction through hydrometeor fallout and a better chance to form a single unbroken cold pool. A cold pool was not present at 0600 UTC in the simulation, based on surface potential temperature, air temperature, dew point, and wind analyses. The cold pool is essential in forming a bow echo by overwhelming the low-level shear and generating the rear-inflow jet.
By 0830 UTC, the observed BE was located in southwest Iowa and northwest Missouri, and it retained a TS rain region (Fig. 7e). Instead of transitioning to BE, the simulated system lost its linearity and evolved to an NL event (Fig. 7f). The environment in which the system was moving was not favorable for bow echoes, let alone linear convection, as the shear and CAPE criteria specified by Weisman (1993) and deemed necessary by James et al. (2006) were not met in the simulation. WRF point soundings revealed most unstable CAPE values under 1500 J kg\(^{-1}\) and lowest 3 km shear under 10 m s\(^{-1}\) in the area east of the convection that was not influenced by precipitation. Weisman (1993) suggested that at least 2000 J kg\(^{-1}\) of CAPE and 20 ms\(^{-1}\) of low-level shear produced bow echoes. The environment also lacked deep-layer shear with much of central Iowa measuring less than 15 m s\(^{-1}\), and the 2-6 km mean wind measured less than 10 m s\(^{-1}\) in central Iowa, indicating very weak steering flow.

Sensitivity tests in place of the Thompson again were conducted by substituting the original scheme for three other schemes. The sensitivity tests showed more variation in convective mode and distribution of precipitation than was present in the first case. The Lin scheme simulated individual cells at initiation, similar to the original run, but the cells dissipated three hours before the NS formed. An NS system formed at 0700 UTC in western Iowa and merged an hour later with an ongoing NL event farther north. The WSM6 scheme produced CC at initiation, which matched the observed system, but it was located in Colorado instead of Kansas. Those cells grew into an NL event at 0200 UTC and joined with the brief line in eastern Nebraska at 0630 UTC. The Morrison scheme produced the most convection of all four schemes. One event began as IC in Colorado at 2030 UTC and transformed into NL at 0200 UTC. A separate NL system initiated in Nebraska at 0000 UTC, and a cluster of cells formed in Kansas at the same time. A line near Omaha developed at 0600 UTC and quickly joined with the NL system that began as IC at 0700 UTC, and the system was classified NL. Overall, all four microphysical schemes failed to produce the observed BE.

As was done in the first case study, to better determine if the shortcomings in the 26 May 2006 case were related to the initialization or boundary condition data, a sensitivity test
was executed with GFS output substituted for NAM. Even less convection was produced when GFS data were used, but the convective modes primarily remained the same. IC initiated and grew to NS later in the evolution without becoming a BE event. Convective initiation was later in the GFS-initialized run compared to that using NAM (2200 UTC versus 2030 UTC), and the transition to NS occurred at 1000 UTC on 27 May compared to 0600 UTC for the NAM-initialized run. No NL event was observed with the GFS-initialized run, possibly due to the late transition to NS. The GFS-initialized simulation featured much less frontogenetical forcing with the warm front and ongoing convection compared to the NAM-initialized run and the RUC analyses, in addition to smaller precipitable water values in the foothills of the Rocky Mountains.

3.6 Conclusions

The present study investigated simulated convective morphologies for 37 warm-season events. The classifications used to designate the convective modes for the events included three cellular, five linear, and a nonlinear mode used by Gallus et al. (2008) in addition to a mixed-complex mode added for the present study. Overall, the model produced more cellular modes than observed, specifically CC, and too few linear modes, especially BE and TS.

A method using normalized timescales was devised to gauge the model’s accuracy in predicting convective mode with respect to radar observations. A “match” occurred if the simulated mode was exactly the same as the observed mode at the same normalized time in the evolution, or if the two modes were in the same classification group. Of the 185 mode comparisons made, 58 were exact matches, and 104 were group matches. The model was least accurate in matching observed BE and TS events, and most accurate matching cellular modes. The total time in the event’s evolution in which a match was observed defined the event score. The average accuracy score for the 37 cases was 0.49, with 15 cases scoring at least 0.50. The model was also penalized if convective initiation or dissipation occurred more than three hours different from that observed. Nine cases violated the timing criterion, and their average accuracy score was 0.32.

Statistical significance testing showed that observed stronger 0-6 km bulk shear and
cooler potential temperatures aloft before convective initiation tended to result in higher accuracy scores for the model. Meanwhile, weaker deep-layer shear, cooler surface potential temperatures, greater potential temperature lapse rates, and large differences between surface-based and mixed-layer CAPE were associated with cases having major timing errors in the simulations. The temperature-related parameters imply elevated convective situations were more difficult to simulate accurately.

Two case studies were performed to investigate in detail the poor simulation of TS and BE events. The first case study, 23 May 2006, featured an observed TS system that traversed the northern Great Plains throughout the evening and overnight hours. The simulation successfully predicted the first two convective modes (BL and NS), but failed to develop the observed stratiform rain. It was shown that the amount deep-layer shear was a major factor in the simulation’s failure, as well as excessive drying in the mid layers of the atmosphere. Very dry air was being funneled into the area directly behind the convective line in the simulation. Shear values normal to the line in the simulation were consistent with NS events from Dial et al. (2010), where greater shear does not properly transport hydrometeors upshear. A vertical cross section of storm relative zonal flow in the simulated system showed a nearly vertical updraft and a rear-inflow jet approximately twice as deep as the average strong-shear TS system shown in Smull and Houze (1987).

In the second event, 26 May 2006, a cluster of cells evolved into a bow echo, while the simulation portrayed individual cells briefly forming a line with no stratiform before transitioning into a nonlinear system. In the simulation, the convection initiated in a drier environment compared to that of the observations because of its location. The isolated nature of the cells did not produce a proper cold pool needed for mature linear development, and mid-level flow steered the convection away from frontogenetical forcing supplied by a warm front located south of the area. Speed convergence and moisture advection at 700 hPa eventually allowed a line to develop. However, weak shear, little CAPE in the near-storm environment, the lack of a developed cold pool, and weak front-to-rear flow within the system brought a quick end to the event’s linearity.

Microphysical sensitivity tests for the case studies showed that problems remained no matter what scheme was used. Although the choice of microphysical scheme did have some
effect on the morphologies simulated, no scheme in these cases was able to produce a system matching that observed. Initial condition sensitivity tests showed some variation with the convective modes, but more importantly, different data did allow the WRF to simulate the observed mode in the first case study. Poor prediction of CAPE and shear in the run using NAM data for model initialization were among the causes of the lack of stratiform rain development in the 23 May 2006 event. Mode varied less between the simulations in the second case study, but substantial differences in the amount of convection were present. Less frontogenetical forcing at the surface and smaller precipitable water values in the GFS data inhibited much of the cellular convection the NAM-initialized run produced.

The findings from the present study suggest a link between some observed environmental parameters before initiation and the accuracy of the model’s depiction of convective evolution. While these factors, such as shear and instability on the synoptic scale, are not solely responsible for the convective mode evolution, they can influence storm-scale processes that determine mode changes. It is unclear how much of the differences in these larger scale processes in the WRF runs are a result of the initial and lateral boundary conditions, and how much of the differences are due to model errors within the WRF. Future work should examine the role of these errors in more detail. In this sample of cases, the most common problems affecting morphological evolution were the formation of TS rain regions and bow echoes. Additional case studies could address other notable issues in the simulations, such as the WRF simulating CC events when BL were observed, which occurred in six cases, and the lack of matching MC events, as only one of six comparisons with an observed MC system was a match.

3.7 Acknowledgments

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3.8 References


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3.9 Figures

Figure 1. Simulated (top bar) and observed (bottom bar) morphologies for each case having a) objective scores of at least 0.5, b) scores less than 0.5, and c) major errors in timing (initiation or dissipation timing differences exceeding 3 hours). Dashed borders in c) represent fraction of timespan outside a three hour grace period for initiation and dissipation.
Shading for convective modes is shown at end of c). Dashed boxes in c) indicate the model either initiated more than three hours early or dissipated more than three hours later than observed event. Dash-dotted boxes indicate the model either initiated more than three hours late or dissipated more than three hours earlier than the observed event.

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Figure 2. Reflectivity (dBZ) during convective mode transitions for the 23-24 May 2006 event. Observed reflectivity a) at 1900 UTC, c) at 2330 UTC, e) at 0330 UTC, g) at 0900 UTC. Simulated reflectivity b) at 2000 UTC, d) at 2330 UTC, f) at 0330 UTC, h) at 0900 UTC. The first green hue correlates to 30 dBZ, and the first yellow hue correlates to 40 dBZ. Contours are listed every 5 dBZ. Observed reflectivity from UCAR imagery archive.
Figure 3. Bulk shear (m s$^{-1}$) from 0-6 km during convective mode transitions for the 23-24 May 2006 event. RUC analyses of 0-6 km bulk shear a) at 1900 UTC, c) at 0000 UTC, e) at 0400 UTC, g) at 0900 UTC. WRF output of 0-6 km bulk shear b) at 2000 UTC, d) at 2330 UTC, f) at 0330 UTC, h) at 0900 UTC. Contour interval is 5 m s$^{-1}$ with blue shades magnitudes of 15 m s$^{-1}$ and greater.
Figure 4. RUC analyses a) at 0400 UTC and WRF output b) at 0330 UTC of 850 hPa moisture advection ($10^9 \text{ g kg}^{-1} \text{s}^{-1}$) on 24 May 2006. RUC analysis c) at 0400 UTC and WRF output d) at 0330 UTC of 850 hPa relative humidity (%).
Figure 5. RUC analyses a) at 0400 UTC and WRF output b) at 0330 UTC of surface potential temperature (K) and 10 m wind (m s\(^{-1}\)) on 24 May 2006.
Figure 6. WRF output at 0330 UTC on 24 May 2006 for a) line-normal storm relative flow (m s\(^{-1}\)), and b) atmospheric sounding at the point of maximum front to rear flow.
Figure 7. Reflectivity (dBZ) during convective mode transitions for the 26-27 May 2006 event. Observed reflectivity a) at 2230 UTC, c) at 0300 UTC, e) at 0830 UTC. Simulated reflectivity b) at 2030 UTC, d) at 0600 UTC, f) at 0830 UTC. The first green hue correlates to 30 dBZ, and the first yellow hue correlates to 40 dBZ (see color bar). Observed reflectivity from UCAR imagery archive.
Figure 8. Surface frontogenesis (x $10^9$ K m$^{-1}$ s$^{-1}$) at 0300 UTC on 27 May 2006 as indicated by the a) WRF output, and b) RUC analysis.
Figure 9. Skew-T log-P diagram from the WRF output at 0600 UTC on 27 May near the NS event.
### 3.10 Tables

Table 1. Frequency of convective modes for the observed and simulated events.

<table>
<thead>
<tr>
<th></th>
<th>IC</th>
<th>CC</th>
<th>BL</th>
<th>BE</th>
<th>TS</th>
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<th>PS</th>
<th>LS</th>
<th>NL</th>
<th>MC</th>
<th>Total</th>
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<td>24</td>
<td>21</td>
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<td>17</td>
<td>1</td>
<td>0</td>
<td>17</td>
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</tr>
<tr>
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<td>8</td>
<td>16</td>
<td>24</td>
<td>8</td>
<td>13</td>
<td>22</td>
<td>1</td>
<td>0</td>
<td>16</td>
<td>1</td>
<td>109</td>
</tr>
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<td>Difference</td>
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<td>9</td>
<td>0</td>
<td>-13</td>
<td>-6</td>
<td>5</td>
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<td>-1</td>
<td>-2</td>
<td>-6</td>
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Table 2. Number of detailed and group matches for each mode comparison.

<table>
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<tr>
<th>Observed mode</th>
<th>Number of occurrences</th>
<th>Detailed matches</th>
<th>Group matches</th>
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<tr>
<td>IC</td>
<td>7</td>
<td>3 43%</td>
<td>6 86%</td>
</tr>
<tr>
<td>CC</td>
<td>11</td>
<td>5 45%</td>
<td>9 82%</td>
</tr>
<tr>
<td>BL</td>
<td>34</td>
<td>17 50%</td>
<td>24 71%</td>
</tr>
<tr>
<td>BE</td>
<td>44</td>
<td>7 16%</td>
<td>26 59%</td>
</tr>
<tr>
<td>TS</td>
<td>29</td>
<td>7 24%</td>
<td>17 59%</td>
</tr>
<tr>
<td>PS</td>
<td>1</td>
<td>0 0%</td>
<td>1 100%</td>
</tr>
<tr>
<td>NS</td>
<td>24</td>
<td>8 33%</td>
<td>10 42%</td>
</tr>
<tr>
<td>LS</td>
<td>0</td>
<td></td>
<td></td>
</tr>
<tr>
<td>NL</td>
<td>29</td>
<td>10 34%</td>
<td></td>
</tr>
<tr>
<td>MC</td>
<td>6</td>
<td>1 17%</td>
<td></td>
</tr>
<tr>
<td>Total</td>
<td>185</td>
<td>58 31%</td>
<td>104 56%</td>
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Table 3. Distribution of simulated convective mode (top of table) as a function of observed mode (left side). The grey shadings represent detailed matches.

<table>
<thead>
<tr>
<th>Modeled Modes</th>
<th>IC</th>
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<th>BL</th>
<th>BE</th>
<th>TS</th>
<th>PS</th>
<th>NS</th>
<th>LS</th>
<th>NL</th>
<th>MC</th>
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<tbody>
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<td>Observed Modes</td>
<td>IC 3 2 1</td>
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<td></td>
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<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>CC 2 5 0</td>
<td>TS 1 7 1 8</td>
<td>4</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>BL 1 6 17</td>
<td>PS 1 1 8</td>
<td>1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>BE 2 4 6</td>
<td>NS 4 2 7</td>
<td>1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td></td>
<td>TS 1 2 5</td>
<td>LS 3 3 3</td>
<td>10</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>PS 1 1 1</td>
<td>MC 1 1 1</td>
<td>1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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</tbody>
</table>
Table 4. Averages of several environmental parameters for all 37 cases, cases with accuracy scores of at least 0.50, cases with scores less than 0.50, cases without a major timing error, and cases with a timing error. “SBCAPE” and “MLCAPE” represent surface-based and mixed-layer CAPE values, and “CAPE diff.” represents the difference between the two CAPE values. The bulk shear values are in the 0-3 km and 0-6 km above ground level layers. “Theta sfc.” and “Theta max.” represent potential temperature near the surface (0-30 hPa above ground level average) and at the level of maximum equivalent potential temperature. “Theta diff.” represents the difference between the two “Theta” values. “Press. diff.” represents the pressure difference between the surface and the level of maximum equivalent potential temperature. “Theta lapse” represents the change of potential temperature per unit pressure in the layer specified by “Press. diff.”. Grey shading represents statistically significant differences either between cases with scores greater than 0.5 versus less than 0.5, or cases with a timing error versus no timing error at the 90% confidence level or greater.

<table>
<thead>
<tr>
<th></th>
<th>All cases</th>
<th>Score &gt; 0.5</th>
<th>Score &lt; 0.5</th>
<th>No error</th>
<th>Timing error</th>
</tr>
</thead>
<tbody>
<tr>
<td>SBCAPE (J kg⁻¹)</td>
<td>1860</td>
<td>1936</td>
<td>1802</td>
<td>1783</td>
<td>2100</td>
</tr>
<tr>
<td>MLCAPE (J kg⁻¹)</td>
<td>2422</td>
<td>2251</td>
<td>2553</td>
<td>2192</td>
<td>3139</td>
</tr>
<tr>
<td>CAPE diff. (J kg⁻¹)</td>
<td>562</td>
<td>315</td>
<td>751</td>
<td>409</td>
<td>1039</td>
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<tr>
<td>0-3 km Shear (ms⁻¹)</td>
<td>11.7</td>
<td>11.9</td>
<td>11.6</td>
<td>11.9</td>
<td>11.3</td>
</tr>
<tr>
<td>0-6 km Shear (ms⁻¹)</td>
<td>19.8</td>
<td>22.4</td>
<td>17.8</td>
<td>21.9</td>
<td>12.9</td>
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<tr>
<td>Theta sfc. (K)</td>
<td>303.5</td>
<td>302.6</td>
<td>304.2</td>
<td>304.4</td>
<td>300.8</td>
</tr>
<tr>
<td>Theta max. (K)</td>
<td>307.1</td>
<td>305.2</td>
<td>308.5</td>
<td>307.2</td>
<td>306.7</td>
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<tr>
<td>Theta diff. (K)</td>
<td>3.6</td>
<td>2.6</td>
<td>4.3</td>
<td>2.8</td>
<td>5.9</td>
</tr>
<tr>
<td>Press. diff. (hPa)</td>
<td>65</td>
<td>48</td>
<td>77</td>
<td>62</td>
<td>73</td>
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<tr>
<td>Theta lapse (K hPa⁻¹)</td>
<td>5.8</td>
<td>5.7</td>
<td>5.8</td>
<td>5.2</td>
<td>7.7</td>
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</table>
Table 5. The normalized timescale and mode morphology for the 23-24 May 2006 case. Initiation is represented as “0”, and the end of the period is represented as “1” on the normalized timescale. The “time of mode change” is the point in the event’s timespan that a convective mode changed in either the observations or model.

<table>
<thead>
<tr>
<th>Time of mode change</th>
<th>Observed mode</th>
<th>Simulated mode</th>
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<tbody>
<tr>
<td>Initiation</td>
<td>BL</td>
<td>BL</td>
</tr>
<tr>
<td>0.22</td>
<td>BL</td>
<td>NS</td>
</tr>
<tr>
<td>0.26</td>
<td>NS</td>
<td>NS</td>
</tr>
<tr>
<td>0.5</td>
<td>TS</td>
<td>NS</td>
</tr>
<tr>
<td>0.82</td>
<td>NL</td>
<td>NS</td>
</tr>
<tr>
<td>End of period</td>
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</tbody>
</table>
Table 6. The normalized timescale and mode morphology for the 26-27 May 2006 case.

<table>
<thead>
<tr>
<th>Time of mode change</th>
<th>Observed mode</th>
<th>Simulated mode</th>
</tr>
</thead>
<tbody>
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<td>0 CC</td>
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<td>0.31</td>
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<tr>
<td>0.61</td>
<td>BE</td>
<td>NS</td>
</tr>
<tr>
<td>0.77</td>
<td>BE</td>
<td>NL</td>
</tr>
<tr>
<td>End of period</td>
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CHAPTER 4. ADDITIONAL RESULTS

4.1 Case studies

Two additional case studies were completed to examine other common tendencies with convective evolution in the simulations. The third case, 10-11 June 2010, showcased an observed BL versus simulated CC event at initiation and a missed BE later in the evolution. The model simulated CC instead of the observed BL six times throughout the study. In the fourth case study, 30-31 August 2010, the model maintained a BL event throughout its lifetime, whereas the observed system transitioned to a linear system shortly after initiation. Observed NS and TS systems versus modeled BL events comparisons were common with 12 combined occurrences.

4.1.1 Case study 3: 10 June 2010 – 11 June 2010

The 10 June 2010 case was similar to the second case study in that the WRF model failed to create a bow echo. The simulation did, however, develop a TS region (Table 7). Also, CC initiated in the simulation instead of a BL as observed. From the RUC analyses, the environment before convective initiation featured a large difference between surface-based and mixed-layer CAPE (975 J kg\(^{-1}\)), high 0-6 km bulk shear (27.4 m s\(^{-1}\)), a lapse rate of 2.8 K hPa\(^{-1}\), and a warm potential temperature at the point of maximum theta-E (314 K). The accuracy score for the 10 June 2010 event was 0.45, which was just below the average score of 0.49 for all cases.

A low pressure center located in north-central South Dakota and its associated cold front that extended southwest into Nebraska and Colorado were the focus for convective initiation. A warm front also extended southeast from the low pressure into western Iowa and northern Missouri. Another low pressure center in southwest Kansas featured a dryline traversing the Oklahoma and Texas panhandles. At 500 hPa, a jet streak with winds of approximately 30 m s\(^{-1}\) was rounding the base of the trough over the Intermountain West and stretched from Utah into western South Dakota.

At 2100 UTC, convection initiated on radar in southeast Wyoming and northeast
Colorado as BL (Fig. 10a), while convective initiation in the simulation occurred at 2000 UTC in the same location, but as CC (Fig. 10b). The simulation predicted the location of the cold front well from surface pressure and wind vector analyses, but the thermal gradient was not as sharp. The WRF output showed the band of forcing in east-central Nebraska instead of far western Nebraska, which was shown by the RUC analyses (Fig. 11). Some frontogenetical forcing was present in the simulation, but it was in a circular pattern near the convection and did not stretch along the front. Because similar moisture and shear profiles were present near the front for the RUC analyses and the simulation, it seems that the strength and expanse of the forcing were the main factors in distinguishing between the observed BL and simulated CC events.

The observed system evolved to NS at 0430 UTC (Fig. 10c), and the WRF-simulated convection grew upscale into an NS event at 0330 UTC (Fig. 10d). The band of forcing expanded and strengthened in the simulation, which promoted more convection to develop along the cold front. The deep-layer shear vectors were virtually parallel to the boundary with magnitudes of 15 to 25 m s\(^{-1}\) ahead of the line. The sufficient shear provided the line with proper hydrometeor transport among the cells and the merging of cold pools.

At 0600 UTC, the line in the simulation transitioned to NL (Fig. 10f). The large-scale environment ahead of the line was not favorable for linear maintenance. The sufficient CAPE (>3000 J kg\(^{-1}\) in a portion of northeast Nebraska) was offset by weak 0-6 km bulk shear with magnitudes less than 15 m s\(^{-1}\). The RUC analyses showed shear values of at least 20 ms\(^{-1}\) in the same location (Fig. 12), which helped the observed line become a bow echo at 0730 UTC.

The observed system continued to be classified as BE through the remainder of the study period, while the simulation transitioned the NL event to TS at 0930 UTC (Fig. 10h). Deep-layer shear increased to 15 to 20 m s\(^{-1}\) in the region ahead of the system, and the difference between the RUC analyses and WRF output in eastern Nebraska and western Iowa was smaller than what was shown at 0600 UTC. However, the magnitude in the WRF output still did not meet the criterion for long-lived bow echoes of 20 m s\(^{-1}\) from Weisman (1993) in most areas. The RUC analyses met the criterion, confirming the observed BE event. Also, no local minima in surface potential temperature were found within the broad cold pool.
region in northeast Nebraska, which suggests that the cold pool was relatively uniform throughout the line. A locally stronger cold pool helps surge a section of the line outward which creates a bowing structure, but if the whole line becomes tilted rearward, the bowing is inhibited (James et al. 2006).

Some features with the simulated TS system were different than those in the failed TS system from the first case study. A vertical cross section of the storm-relative flow displayed an updraft slightly more tilted upshear, which promoted sufficient hydrometeor transport to the rear of the system (Fig. 13a). The cross section also showcased a shallower and lower rear inflow jet than in the NS event from the first case study. The jet also descended more toward the surface, and it was much moister. A skew-T log-P diagram from the point of maximum front-to-rear flow exhibited a virtually saturated column from 600 hPa upward (Fig. 13b). The wind barbs from the diagram showed little speed or directional change in the mid and low levels of the atmosphere (Fig. 13b). The line-normal 2-6 km shear was computed ahead of the convection, and the average component was 3.9 m s$^{-1}$, which was 1.2 m s$^{-1}$ weaker than the simulated NS system from the first case study.

The model instead showed CC and TS instead of BL and BE, with the lack of frontogenetical forcing along the front and deep-layer shear comprising the main factors for discrepancy. Some findings from the simulated TS system did verify the failed TS event in the first case study, though, as line-normal shear, the tilt of the updraft, and the moisture content of the rear inflow jet all promoted TS development.

4.1.2 Case study 4: 30 August 2010 – 31 August 2010

The 30 August 2010 event was selected for the simulation’s failure to develop a linear system altogether during its evolution. The observed system merged its cells into a line and eventually grew a TS rain region, whereas the simulated system maintained a BL throughout the entire lifespan. The event received an accuracy score of only 0.25 (Table 8). This event initiated with virtually the same amount of surface-based and mixed-layer CAPE (1044 and 1055 J kg$^{-1}$), sufficient deep-layer shear at 21.4 m s$^{-1}$, and a level of maximum theta-E only 20 hPa above the surface, allowing for a change in potential temperature of 1.5 K. These factors suggest surface-based convection.
The synoptic conditions for the 30 August 2010 case were similar to those of the third case study with a low pressure area over the northern Great Plains and the presence of cold and warm fronts. For the current case study, the low pressure center moved from western South Dakota into southern Manitoba and deepened considerably from 1002 hPa to 994 hPa in 24 hours. The cold front extended southward into northern Colorado, while the warm front stretched eastward into Minnesota and western Ontario. At 500 hPa, a jet streak exited the base of the trough and reached speeds of 35 m s\(^{-1}\) over the Black Hills of South Dakota. Speeds of 50 m s\(^{-1}\) were attained at 300 hPa over the same area, indicating divergence.

A broken line of cells formed at 2030 UTC in the central portion of the Dakotas, and the simulation initiated a broken line 30 minutes later in South Dakota and Nebraska (Fig. 14). Placement and intensity of the surface low were forecasted accurately in the simulation, but the shear was underestimated along and ahead of the cold front. A pocket of 25 to 30 m s\(^{-1}\) deep-layer shear was present in central South Dakota and Nebraska in the RUC data, but the simulation showed mainly 15 to 20 m s\(^{-1}\). The zonal component of the 2-6 km average wind was computed ahead of the main frontogenetical forcing and convection at 2100 UTC. This procedure was similar to the method used by Dial et al. (2010) that discriminates between systems that retain individual cells versus systems that become linear three hours after initiation. The zonal component was chosen as a proxy for boundary-normal flow in the present case because the line was primarily in the north-south direction. Slow 2-6 km average flow normal to a boundary (<10.5 ms\(^{-1}\)) supported more rapid growth into a linear system (Dial et al. 2010). Overall, the simulation had higher zonal component wind speeds near the convection than the RUC analyses, and the nose of 10 m s\(^{-1}\) flow was simulated too far north (Fig. 15). The average boundary-normal flow in the simulation was 10.2 m s\(^{-1}\), which was slightly lower than the “threshold” as stated in Dial et al. (2010), but it was higher than the average speed in the RUC analyses (7.9 m s\(^{-1}\)).

At 0000 UTC, the observed system transitioned to NS (Fig. 14c), but the simulated system maintained BL (Fig. 14d). Surface potential temperatures in the WRF output indicated separate smaller, disconnected cold pools with the system, consistent with a BL event (Fig. 16a). Contrary to the modeled system, the RUC analyses depicted a broader area of cooler surface air temperatures and potential temperatures and a more divergent wind
pattern, which was indicative of a cold pool from a solid line of convection (Fig. 16b). Frontogenetical forcing was in a disconnected linear pattern in the WRF output, whereas the observed maximum frontogenesis occurred in a solid line from Fargo, North Dakota, to central Nebraska. The deep-layer shear near the convection had a larger westerly component in the simulation than what was observed. The lack of boundary-parallel 0-6 km shear inhibited the merging of cold pools by not transporting the hydrometeors toward other cells. The simulation again underestimated deep-layer shear ahead of the system by 5 to 10 m s\(^{-1}\) for most of eastern South Dakota and Nebraska.

The observed event formed a TS rain region at 0230 UTC (Fig. 14e), but the modeled system failed to develop into a line (Fig. 14f). RUC surface potential temperature analyses at 0300 UTC exhibited a single cold pool associated with the line, but WRF output continued to show separate cold pools. The simulation also continued to lack in deep-layer shear by 5 to 10 m s\(^{-1}\) in southwest Minnesota and northwest Iowa.

In general, the WRF underestimated deep-layer shear near the convection throughout the study period, and the mid-level flow’s orientation normal to the boundary was too strong. The 2-6 km average wind mostly agreed with findings from Dial et al. (2010), where systems that have large boundary-normal components tended to stay discrete three hours after convective initiation. The large zonal deep-layer shear component also contributed to the failing of merging cold pools. The lack of a single, strong cold pool inhibited the formation of a line of storms, thus maintaining the BL classification.
4.2 Figures

Figure 10. Reflectivity during convective mode transitions for the 10-11 June 2010 event. Observed reflectivity a) at 2100 UTC, c) at 0430 UTC, e) at 0730 UTC, g) at 0930 UTC. Simulated reflectivity b) at 2000 UTC, d) at 0330 UTC, f) at 0600 UTC, h) at 0930 UTC. The first green hue correlates to 30 dBZ, and the first yellow hue correlates to 40 dBZ. Contours are listed every 5 dBZ. Observed reflectivity from UCAR imagery archive.
Figure 11. Surface frontogenesis ($x \times 10^9 \text{K m}^{-1} \text{s}^{-1}$) in the a) RUC analyses at 2100 UTC, and b) WRF output at 2000 UTC. Surface potential temperature (K) and 10m wind vectors (m s$^{-1}$) in the c) RUC analyses at 2100 UTC, and d) simulation at 2000 UTC.
Figure 12. Bulk shear (m s$^{-1}$) from 0-6 km at 0600 UTC in the a) simulation, and b) RUC analyses.
Figure 13. WRF output at 0930 UTC on 11 June 2010 for a) line-normal storm relative flow (m s$^{-1}$), and b) skew-T, log-P diagram at the point of maximum front to rear flow.
Figure 14. Reflectivity during convective mode transitions for the 30-31 August 2010 event. Observed reflectivity a) at 2030 UTC, c) at 0000 UTC, e) at 0230 UTC. Simulated reflectivity b) at 2100 UTC, d) at 0000 UTC, f) at 0230 UTC. The first green hue correlates to 30 dBZ, and the first yellow hue correlates to 40 dBZ. Contours are listed every 5 dBZ. Observed reflectivity from UCAR imagery archive.
Figure 15. Zonal component of the 2-6 km average wind (m s$^{-1}$) at 2100 UTC for a) the simulation, and b) the RUC analyses. The black dots correlate to the main frontogenetical forcing and convection.
Figure 16. Surface potential temperature (K) and 10m wind (m s$^{-1}$) at 0000 UTC on 31 August 2010 for a) the WRF output, and b) the RUC analyses.
4.3 Tables

Table 7. The normalized timescale and mode morphology for the 10-11 June 2010 case.

<table>
<thead>
<tr>
<th>Initiation</th>
<th>Time of mode change</th>
<th>Observed mode</th>
<th>Simulated mode</th>
</tr>
</thead>
<tbody>
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</tr>
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</tr>
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<td>0.84</td>
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<td>End of period</td>
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<td></td>
</tr>
</tbody>
</table>
Table 8. The normalized timescale and mode morphology for the 30-31 August 2010 case.

<table>
<thead>
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<th>Observed mode</th>
<th>Simulated mode</th>
</tr>
</thead>
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</table>
CHAPTER 5. GENERAL CONCLUSIONS

The goal of the present study was to analyze convective morphology in WRF simulations and to examine discrepancies between the simulated and observed convective events. Convective modes from 37 warm-season cases were classified, and trends with simulated morphology were noted. A scoring system was developed to determine the accuracy of the WRF simulations. Associations between the morphology accuracy scores and large-scale environmental parameters before convective initiation were established. Four case studies were then performed for events that featured common convective mode differences between the simulated and observed events.

The classification scheme from Gallus et al. (2008) and Duda and Gallus (2010) was used to identify all the convective modes displayed by the events. The scheme consisted of three cellular modes, five linear modes, and a non-linear mode (NL). The cellular modes comprised individual cells (IC), clustered cells (CC), and broken line (BL). The linear modes were based on the orientation of the stratiform rain region (if any) and whether the line bowed. The classifications included bow echo (BE), systems with no stratiform rain (NS), parallel stratiform rain (PS), leading stratiform rain (LS), and trailing stratiform rain (TS). A mixed-complex mode (MC), which was added for the present study, was used for situations that shared characteristics of two or more of the aforementioned modes. Throughout the 37 events, the simulations featured more cellular modes, particularly CC, and fewer linear modes, especially BE and TS, than the observed events. The simulation failed to develop a bow echo in 13 cases and a squall line with trailing stratiform rain in 12 cases.

The scoring system for WRF accuracy was based on a normalized scale of the event lifespan and the duration of each mode displayed by the model and radar. A scale showing the chronological order and duration of each mode was created for the observed and simulated event in every case, and the two scales were compared. If the convective mode was the same for the two scales at the same normalized time in each event, a “detailed match” occurred, and a point was given. If the convective modes shared a classification group (i.e., cellular or linear), a half point was given. A total of 185 mode comparisons were made, with 58 resulting in detailed matches and 104 in group matches. Overall, the model was most
accurate with matching cellular modes and least accurate with matching linear modes. Only 55% of comparisons with an observed linear mode resulted in a group match, versus 75% for cellular mode comparisons. The model exactly matched 48% of cellular modes and 22% of linear modes. The average accuracy score for all events was 0.49, and only 15 of the 37 cases received a score of at least 0.50. A penalty was introduced if the simulated convective initiation or dissipation occurred more than three hours apart from the observed. The average score for the nine cases that did not meet the criterion was 0.32.

RUC analyses were used to obtain environmental parameters one hour before initiation at the centroid of the system in order to gather insight of the synoptic environment. The parameters computed included surface-based and mixed-layer CAPE, 0-3 km and 0-6 km bulk shear, and potential temperature at the surface and at the level of maximum theta-E. Statistical significance testing suggested that stronger 0-6 km bulk shear and cooler potential temperatures at the level of maximum theta-E before convective initiation increased the simulation’s accuracy score for convective evolution. The testing also concluded that very weak deep-layer shear, cooler potential temperatures at the surface, steep potential temperature lapse rates (from the surface to the level of maximum theta-E), and large differences between surface-based and mixed-layer CAPE increased the possibility of major timing errors within the simulation.

The first case study, 23-24 May 2006, was chosen for the simulation’s failure to produce a TS rain region. A squall line developed ahead of a cold front in the northern Great Plains and upper Midwest, and a TS rain region eventually formed. The model accurately predicted the modes BL and NS, but the simulated system never transitioned to TS like the observed system. Strong 2-6 km boundary-normal shear, a very dry, deep rear inflow jet, and a warmer cold pool led to the failure of the stratiform rain. Microphysics sensitivity tests showed similar results, but exchanging NAM initial conditions for GFS initial conditions allowed the model’s system to evolve into TS. Large-scale deep-layer shear and instability profiles in the GFS data upon model initiation were more similar to the observed data, which appeared to have a profound effect on convective mode development later in the evolution.

The second case study, 26-27 May 2006, was chosen because the simulated system never formed the observed bow echo. The WRF output also initiated CC instead of IC.
warm front stretched across the region was the focal point for the observed bow echo
development. The isolated nature of the cells upon initiation did not allow for proper cold
pool merging needed for upscale growth into a linear system, and the steering flow at mid-
levels directed the convection away from the forcing. An elevated linear system managed to
develop toward the end of the study period due to enhanced speed convergence and moisture
advection, but weak shear and little instability resulted in a transition from NS to NL. No
bow echo was formed during the sensitivity tests, but CC was produced with two of the three
other microphysics schemes.

The third case, 10-11 June 2010, featured another failed bow echo, but the simulated
system did form a TS event. The initial mode was also incorrect, as CC was formed instead
of BL. A lack of surface forcing along the cold front inhibited the BL development in the
simulation. Weak shear and a weak cold pool were unfavorable for transition from TS to BE
in the model, but analyses did show differences between the TS in this case study and the
failed TS from the first case study. Weak mid-level boundary-normal shear sent
hydrometeors upshear, and a moister, shallower rear-inflow jet reduced excess evaporation,
ultimately leading to TS development.

The fourth case, 30-31 August 2010, was chosen for the failure of the simulated event
to become linear altogether during its lifespan. The observed system transitioned from BL to
NS to TS, but the modeled event remained BL throughout the study period. Strong mid-level
flow normal to the boundary inhibited upscale growth from IC into a linear system shortly
after initiation. Deep-layer shear was not oriented with the boundary after the first few hours,
which prevented merging of the cold pools.

Synoptic and mesoscale factors of the simulations’ problems with convective
morphology have been identified, but still much work remains to alleviate convective
evolution issues in future NWP modeling. Convective evolution is not only dependent in a
few environmental parameters, such as shear and instability, but also the model configuration.
Questions arise regarding the full impact of initial and lateral boundary condition data and
deficiencies within the WRF model on convective morphology and forecasts of certain
convective modes. Therefore, additional research is needed to examine initial condition and
parameterization effects on convective evolution. In addition, detailed examination of other
mode differences mentioned in this study would be beneficial. Improvement in the forecasting of convective evolution would reap benefits for the population as a whole, and all avenues should be taken to reach that goal.
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