WRF forecast skill of the Great Plains low level jet and its correlation to forecast skill of mesoscale convective system precipitation

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WRF forecast skill of the Great Plains low level jet and its correlation to forecast skill of mesoscale convective system precipitation

by

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A thesis submitted to the graduate faculty in partial fulfillment of the requirements for the degree of

MASTER OF SCIENCE

Major: Meteorology

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Iowa State University
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One of the primary mechanisms for supporting summer nocturnal precipitation across the central United States is the Great Plains low-level Jet (LLJ). Mesoscale Convective Systems (MCSs) are organized storm complexes that can be supported from the upward vertical motion supplied at the terminus of the LLJ, which bring beneficial rains to farmers. As such, a need for forecasting these storm complexes exists. Correlating forecast skills of the LLJ and MCS precipitation in high spatial resolution modeling was the main goal of this research. STAGE IV data was used as observations for MCS precipitation and the 00-hr 13 km RUC analysis was employed for evaluation of the LLJ. The 4 km WRF was used for high resolution forecast simulations, with 2 microphysics and 3 planetary boundary layer schemes selected for a sensitivity study to see which model run best simulated reality. It was found that the forecast skill of the potential temperature and directional components of the geostrophic and ageostrophic winds within the LLJ correlated well with MCS precipitation, especially early during LLJ evolution. Since the 20 real cases sampled consisted of three LLJ types (synoptic, inertial oscillation and transition), forecast skill in other parameters such as deep layer and low level shear, convergence, frontogenesis and stability parameters were compared to MCS forecast skill to see if consistent signals outside of the LLJ influenced MCS evolution in forecasts. No correlations were found among these additional parameters. Given the variety of synoptic setups present, the lack of forecast skill correlations between several variables and MCSs resulted as different synoptic or mesoscale mechanisms played varying roles if importance in different cases.
CHAPTER 1: THESIS OVERVIEW

1.1 Introduction

Nocturnal convection is a primary source of precipitation across the US Great Plains during the warm season months (Jirak and Cotton 2007; Coniglio et al. 2010; French and Parker 2010). Agriculture is heavily dependent on the resultant rainfall (which usually occurs in organized thunderstorm clusters known as Mesoscale Convective Systems or MCSs), making nocturnal convection an important phenomenon to understand and predict. High spatial resolution modeling may thus be needed to improve the accuracy in simulating the initiation, evolution and sustenance of MCSs, which would help forecasters better predict night time flash flooding, severe winds/hail and tornado potential (Stensrud and Fritsch 1993). Such forecasts may allow farmers to make better short term decisions that involve fertilizer runoff and soil hypoxia (Schaffer et al. 2011). In addition, emergency managers may also use improved forecasts of MCSs to better prepare for the threats that night time flash flooding and severe weather present (Jirak and Cotton 2007).

In order to improve MCS forecasts, proper simulations of the Great Plains nocturnal Low-Level Jet (LLJ) are necessary given that the LLJ is believed to have a substantial impact on MCS development and maintenance (Mitchell et al. 1995; French and Parker 2010). In general, the classic Great Plains LLJ is defined as a strong current of southerly winds located approximately between 150-2000 m above ground level (Bonner 1968; Mitchell et al. 1995; Whiteman et al. 1997; Song et al. 2005) and is considered to be one
of the most widely studied phenomena in atmospheric sciences (Bonner 1968; Parish and Oolman 2010).

In addition to varying definitions of the LLJ, the lack of observations of the LLJ presents significant challenges to the research and forecasting communities. At the present time, radiosondes are launched twice (0000 and 1200 UTC) each day at select locations across the Great Plains and Midwest states. In addition, two high powered (915 MHz) vertical wind radar profiler are operational in the Southern Plains, which collect wind data at 60m intervals, hourly. These limited observations are the meteorology community’s current source of in-situ data for the LLJ, which have been employed in several LLJ studies (Izumi and Barad 1963; Bonner 1968; Frisch et al. 1992; Mitchell et al. 1995; Whiteman et al. 1997). Unfortunately, these resources are not adequate for identifying the spatial evolution (three dimensional LLJ structure) and temporal evolution of the LLJ, especially since the LLJ peak is known to occur around 0600-0900 UTC (Whiteman et al. 1997).

Given the lack of a high density three dimensional network of in-situ meteorological observations, understanding the details of LLJ behavior would prove to be a daunting task. Without the ability to fully understand and predict LLJs to an acceptable degree of accuracy, difficulty may follow in properly simulating nocturnal MCS activity in atmospheric models given MCS dependency on LLJ evolution. As such, it would be beneficial to employ an appropriate substitute to in situ observations (validated by previous research) and if improvements in LLJ simulations are achieved (compared to the
substitute to in situ data), then better simulations of MCSs may naturally follow and better precipitation forecasts could be made.

1.2 Research Questions

This study focused on the relationship between nocturnal MCS activity and the evolution of the Great Plains LLJ. Specifically, it was hypothesized that a better simulation of the nocturnal LLJ at a high three dimensional spatial resolution would result in better simulated MCSs. A necessity for this project was to employ a substitute to in situ observations. For this study, the substitute was gridded output of wind, moisture and temperature data from the 00-hr 13 km Rapid Update Cycle (RUC) analysis. The first question to address was whether the RUC analysis was an appropriate substitute for in situ observations. Past research by Thompson et al. (2003), Hane et al. (2008), Schumacher and Johnson (2009), Coniglio et al. (2010) has shown that using the 13 km RUC analysis in an operational and research setting can be successful for mesoscale analyses, including the region of the LLJ. STAGEIV data were used for observed precipitation to validate model rainfall in the simulated MCSs. STAGEIV data are gridded and computed from hourly surface observations combined with derived satellite and composite radar QPE as noted in Lin (2005), so a direct comparison from model output can be made if both the STAGEIV data and model output are placed on the same grid.
Simulations were conducted by running the Weather Research and Forecasting Model (WRF) with the Advanced Research WRF core (ARW), at 4 km horizontal grid spacing, with 50 defined eta levels in the vertical (with 28 of those levels concentrated to approximately 5 hPa grid spacing below 850 hPa). Given that the LLJ is influenced at least in part by planetary boundary layer (PBL) characteristics (Blackadar 1957; Izumi and Barad 1963; Bonner 1968; Bonner and Paegle 1970; Parish and Oolman 2010), the WRF was run using three different PBL schemes (Mellor Yamada Janjic (MYJ), Mellor Yamada Nakanishi Niino (MYNN 2.5) and Yonsei University (YSU)) to see how PBL and resultant LLJ evolution differed among multiple local and non-local mixing PBL schemes. In addition, the WRF was run using two separate microphysics (MP) schemes. The Thompson scheme is a double moment ice predictive scheme that uses time split fall terms (Thompson et al. 2008) to determine the evolution of hydrometeors with time, while the WSM6 scheme (a single moment scheme defined in Hong et al. (2006) that was designed for high resolution modeling) centered its ice hydrometeor generation calculations around graupel processes. Using two different schemes (single vs. double moment) helps to answer the question regarding which hydrometeor computation process would be more appropriate in simulating MCS core and precipitation shield induced rainfall.

Having the 4 km WRF-ARW output, and 13 km RUC analysis gridded over the same areal domain means that direct statistical analyses could be conducted in order to determine a measure of forecast skill of both the LLJ and simulated MCSs. Mean absolute errors would be calculated on LLJ variables such as atmospheric water vapor mixing ratios, potential temperature and the magnitude and directions of the total,
geostrophic and ageostrophic components. For determining skill in MCSs, STAGEIV data and WRF data may be directly compared at six hourly intervals, with Equitable Threat Scores computed. From here, the LLJ MAEs may be plotted against ETSs associated with MCS induced precipitation to see if there is a correlation in forecast skill between the LLJ and MCSs. Assuming one specific MP/PBL WRF parameterization proved to be outstanding in simulating the LLJ and associated MCS(s) in comparison to other MP/PBL combinations (having overall higher ETS precipitation scores and lower MAEs for LLJ variables), forecasters could decide which parameterization to employ when developing and running high resolution models that involved forecasting LLJs and associated MCS activity.

1.3 Thesis Outline

A journal based thesis was submitted for this research. The first chapter highlights the general background of the thesis and its structure. The second chapter is a literature review that discusses previous works on LLJs, MCSs and modeling features implemented in the present research. Chapter three is the journal paper to be submitted to Weather and Forecasting. Chapter 4 will provide additional materials and results to the research conducted and chapter five will highlight the conclusions of the research conducted along with ideas for future works and critical reflections of the research performed.
CHAPTER 2: LITERATURE REVIEW

2.1 Mechanisms of LLJ Development

Given that the LLJ plays a key role in MCS development (Mitchell et al. 1995; French and Parker 2010), understanding the nature of LLJ evolution is vital. Blackadar (1957) was one of the first to deeply investigate the nocturnal wind maximum at low levels of the troposphere and diagnose dynamic behavior. He concluded that upon the development of a nocturnal inversion, turbulent mixing rapidly diminished at the top of the PBL, allowing for only the horizontal pressure gradient forces and the coriolis force to act on a given air parcel atop the inversion layer. Given the force imbalances with turbulent mixing and friction now negligible, introducing the complex number $i$ to the $u$ and $v$ geostrophic wind equations (1 and 2) allowed Blackadar (1957) to derive a prognostic mathematical solution (3-4) that when integrated (5) yielded the LLJ wind profile, where the peak magnitude of the LLJ could be calculated along with the time it occurs. This relationship is known as the inertial oscillation

\[
\frac{\partial}{\partial t} (u - u_g) = f (v - v_g) \quad (1)
\]

\[
\frac{\partial}{\partial t} (v - v_g) = -f (u - u_g) \quad (2)
\]

\[
W = (u - u_g) + i(v - v_g) \quad (3)
\]

\[
\frac{\partial W}{\partial t} = -ifW \quad (4)
\]

\[
W = W_0 e^{-ift} \quad (5)
\]
where \( u, u_g, v, v_g \), and \( f \) are the total and geostrophic components of wind in the x and y directions, respectively, \( f \) is the coriolis parameter, \( W \) is the acceleration vector and \( W_0 \) is the acceleration vector deviation at initial time \( t \) around sunset.

Wexler (1961) discovered that in addition to the inertial oscillation, the deflection of air parcels to the right due to the Rocky Mountains (from an east to west pressure gradient induced flow) that when combined with the inertial oscillation, induced the development of a narrow but strong southerly current of air over the Great Plains states. The importance of terrain influence from the Rockies would later be highlighted by many scientists, including Bonner and Paegle (1970) who found that differences in diurnal terrain heating had two major impacts on LLJ development. The first impact is seen through the development of a derivation of the \( u \) and \( v \) geostrophic wind components (6 and 7, respectively) that included both the sloping of terrain and daytime heating (8).

\[
U_g = -\frac{g}{f} (1 + S^*) \frac{\partial D}{\partial y} + S^* \frac{g}{f} \frac{\partial z}{\partial y} \tag{6}
\]

\[
V_g = -\frac{g}{f} (1 + S^*) \frac{\partial D}{\partial x} - S^* \frac{g}{f} \frac{\partial z}{\partial x} \tag{7}
\]

\[
S^* = \frac{T^* - T_p}{T_p} \tag{8}
\]

For this derivation, \( g \) is gravity, \( f \) is the coriolis parameter, \( D \) is the difference in terrain height and pressure altitude, \( S^* \) is the specific virtual temperature anomaly defined by \( T^* \) (the virtual temperature) and \( T_p \), the temperature altitude. During the day, the \( S^* \) increased to an extent where a maximum in \( V_g \) was noted across the plains, allowing for a strong background flow to exist within the PBL that would act on frictional decoupling.
later in the evening and contribute to LLJ intensity. Similarly, shear in the geostrophic flow (induced by diurnal heating and sloping terrain) with height yielded a thermal wind relationship which also contributed to a daytime geostrophic maximum, ultimately culminating in a circular oscillation in the total wind, which allowed for a southerly maximum to develop around midnight. This research was further solidified via modeling studies. First, Pan et al. (2004) performed regional model sensitivity simulations, where the Rocky Mountains were removed in the fifth generation Pennsylvania State University-National Center for Atmospheric Research Mesoscale Model (MM5) for real time cases in 1993, during a period where several strong LLJs were observed. As a result, the summer time Bermuda High extended farther to the west with a weaker southerly flow and lack of appreciable LLJs, giving credit to the contribution of slope induced nocturnal thermal gradients to LLJ activity. Lastly, Parish and Oolman (2010), who investigated LLJ development in the 12 km North American Model (NAM), found that maximum geostrophic southerly winds occurred in the Great Plains during the late afternoon due to the heating of sloped terrain. These provided the strong background flow for the inertial oscillation to act upon for LLJ development, thus backing the work of Bonner and Paegle (1970).

Large scale synoptic influences also play a significant role in LLJ development (both diurnally and nocturnally). Uccellini and Johnson (1979) determined that synoptic scenarios where strong cyclonic flow and an upper level jet were present proved to be favorable for intense LLJs to develop. It was found that strong divergence in the exit region of an upper level (300-200 hPa) jet streak promoted large scale upward motion, with low level convergence due to mass continuity, promoting strong southerly low level
flow in a process known as jet coupling. The works of Chen and Kpaeyeh (1993) also solidified the findings of Uccellini and Johnson (1979).

2.2 LLJ Climatology

Great Plains southerly LLJ climatology is important to understand given that the LLJ is a major source of atmospheric moisture and supports large scale upward vertical motion, thus rendering the environment conducive for widespread convection, including MCS activity during the warm season (Chen and Kpaeyeh 1993; Augustine and Caracena 1994; Mitchell et al. 1995; Higgins et al. 1997; Jirak and Cotton 2007; Schumacher and Johnson 2009; Coniglio et al. 2010).

Hoecker (1963) examined the frequency and occurrence of all LLJs using a pilot balloon network which spanned from the Texas Panhandle to central Arkansas, while Bonner (1968) conducted similar research across the lower 48 contiguous states using rawinsonde data. Both scientists were among the first to determine that the classic southerly Great Plains LLJ was the most common LLJ to occur. Seasonally, these types of LLJs were mostly present during the summer due to both the LLJ mechanisms mentioned previously, and because of prevailing southerly flow rotating anticyclonically around the western edge of the Bermuda high. Chen and Kpaeyeh (1993) backed this notion using a composite analysis of 850 hPa wind and moisture data for 11 warm seasons. 850 hPa and 200 hPa air mass and atmospheric water vapor flows were evaluated via a stream function analysis based on pure rotational flow and a divergent circulation calculation. Results showed that a large high pressure anticyclone at 850 hPa
dominated the eastern United States, promoting strong southerly flow of moist air from the Gulf of Mexico (proving that the LLJ is an important mechanism in moisture transport). This was further shown in the divergent fields, where large scale mass convergence was noted in the central United States at 850 hPa. Large scale cyclonic flow and divergence was noted over the Great Plains at 200 hPa, also hinting at the fact that many LLJs are likely synoptically influenced by jet coupling (discussed earlier). Data were lacking spatially and temporally for many of these scientists (and many others performing LLJ climatology studies, including Mitchell at al. 1995), casting doubt on the fact that all LLJs were being observed and suggesting that their peak magnitude was underestimated, with the height of the peak wind potentially overestimated. Using the National Centers for Environmental Prediction-Data Assimilation Office (NCEP-DAO) multi-year reanalysis, Higgins et al. (1997) was able to conclude that an enhanced transport of moisture from the Gulf of Mexico across the Great Plains States typically commenced from May to August, where 25% more in excess precipitation fell nocturnally vs, diurnally. This suggested that the southerly Great Plains LLJ was mainly a summer and overall warm season occurrence and that there was a direct link between the LLJ and nocturnal warm season convective activity. Studies by Whiteman et al. (1997) and Song et al. (2005) used a high vertical resolution profiler in Lamont, Oklahoma to observe LLJs and determined that many jets were in fact undetected due to their maximums occurring closer to the ground than what other observation networks were able to detect as seen in Mitchell at al. (1995). Still, Song et al. (2005) was able to show that a maximum in southerly Great Plains LLJs occurred during the summer months, backing the claim that many jets are induced in part by the Bermuda High.
2.3 LLJ Classification

LLJs have often been classified by their intensity and the ambient synoptic environment they are embedded in. Bonner (1968) was one of the first to introduce criteria for rating LLJ strength by introducing three categories for LLJ depth and intensity. Criterion 1 required that the magnitude of the wind must be at or greater than 12 ms\(^{-1}\) and must decrease to a minimum by 6 ms\(^{-1}\) at or below 3000 m above ground level (AGL). Criterion 2 required a peak wind of at least 16 ms\(^{-1}\) with a minimum decrease by 8 ms\(^{-1}\) at or below 3000 m AGL, and criterion 3 peak winds were at least 20 ms\(^{-1}\) with a decrease of at least 10 ms\(^{-1}\) at or below 3000 m AGL. While Kumjian et al. (2006) found that many LLJs can surpass the criterion 3 thresholds, the addition of 3 higher categories (i.e. magnitude thresholds of 24 ms\(^{-1}\), 28 ms\(^{-1}\) and 32 ms\(^{-1}\)) revealed that the LLJ case distribution was approximately normal, with a vast majority of cases falling under the criterion 3 threshold (a finding noted in the present research). In addition, Mitchell (1995) found that a majority of the strongest jets (criterion 3) were six times more likely to occur at/around midnight vs. daytime, indicating that the rawinsonde network would most likely underestimate the strength of the LLJ or perhaps miss it completely. These LLJs would occur between 0600-0900 UTC as this was when the inertial oscillation had the greatest influence on LLJ behavior. It was also found by Mitchell et al. (1995) that criterion 3 LLJs were most impacted by boundary layer processes and the inertial oscillation more so than criterion 1 and 2 LLJs. Whiteman et al. (1997) (while employing the Bonner (1968) criteria for LLJ magnitude classification) found that many LLJs were predominantly in the criterion 1 and 2 categories, with more jets observed overall in the winter. Still, two caveats existed within the research of
Whiteman et al. (1997). First, LLJs were not divided in categories of northerly vs. southerly jets. In addition, they used rawinsonde and low powered (400 MHz) profilers which only recorded data at or above 500 m AGL, like Mitchell et al. (1995). As such, many LLJs were likely missed, as was determined by Song et al. (2005), who employed a high powered (915 MHz) profiler that was able to record data at or just below 100 m. Their findings concluded that a majority of criterion 2 and 3 LLJs had a southerly component and occurred during the warm season.

In addition to LLJ intensity, work has been done to classify LLJs by their synoptic environment. Unpublished work by Dr. Tsing-Chang Chen and his team at Iowa State University classified LLJs via a streamline analysis with wind magnitudes of the 900 hPa and 200 hPa fields along with overlaid 900 hPa and 200 hPa divergent fields. Cases where cyclonic flow and a jet streak at 200 hPa with strong southerly flow at 900 hPa were present, along with coupled 900 hPa convergence and 200 hPa divergence were classified as type cyclonic (type C) LLJs. These were considered to be strongly synoptically forced jets. Cases where 200 hPa weak, anticyclonic flow correlated with strong southerly 900 hPa flow and little to no 900 hPa convergence and 200 hPa divergence coupling were classified as type anticyclonic (type A) and were suspected to be caused mostly by the inertial oscillation as well as terrain sloping and heating processes. Lastly, cases where synoptic influences may have begun (or finished) playing a role in some LLJ dynamics (or cases that could not be solidly classified as type A or C) were considered type transition (type T) LLJs. This last category was rather subjective in nature.
2.4 LLJ Peak Altitude

Bonner (1968) was one of the first scientists to attempt diagnosing the height of the peak LLJ wind magnitude above ground level. Using a pilot rawinsonde network (consisting of eight stations across the Great Plains that took measurements at the surface, 150 m, 300 m, and measurements from 500-3000 m at 500 m intervals), Bonner was able to conclude that the majority of jets at 0600 UTC peaked (on average) at a height of 785 m AGL. Mitchell et al. (1995) used a data set with higher horizontal spatial resolution (the Wind Profiler Demonstration Network) which consisted of 31 stations across the Great Plains. The downside was that the lowest level of data given was 500m (which increased at 250 m intervals to 19 km). As such, the average height for the peak wind of the LLJ in this study was approximately 1000 m AGL. As mentioned earlier, LLJs are often contained within the lowest few hundred meters AGL and as such, the peaks of many LLJs were likely missed. Taking this into account, Whiteman et al. (1997) employed a Long Range Navigation (LORAN-C) cross-chain wind finding technique involving Vaisala Rs80-15L radiosondes, which was able to frequently achieve its lowest measurement between 50-100 m AGL for cases over the span of two years. As a result, their average LLJ peak wind altitude was 596 m AGL. Using the 915 MHz vertical profilers near Lamont, Oklahoma, Whitewater, Kansas and Beaumont, Kansas, Song et al. (2005) investigated six years of profiler data, identifying the LLJ wind maxima to occur between 200-400 m AGL, with the strongest jets located at or above 300 m AGL.
2.5 LLJ Terminus and its Relation to Nocturnal Convection and MCSs

The terminus of the LLJ refers to the jet exit region, where large scale low level convergence is evident. Bonner (1966) was one of the first to investigate in detail the convergence associated with the LLJ using hourly precipitation charts, surface weather observations, and upper air data from the Air Force and Navy stations. Through the cases he evaluated, Bonner (1966) concluded that precipitation did not occur in the core of the LLJ where the greatest moisture and instability resided. Rather, precipitation developed and sustained itself at the exit region of the LLJ (as was determined from cross section analyses) where convergence was maximized and vertical velocity magnitudes were greatest. He noted though, that this might not be the reason for nocturnal convective initiation, but rather sustenance of convection already in progress. Uccellini and Johnson (1979) noted that for the more strongly synoptic driven setups where an upper level jet stream was present, the moisture flux provided at the terminus of the LLJ enhanced instability in the region of greatest upper level support, allowing for the development of vigorous convection and MCSs capable of severe weather. Augustine and Caracena (1994) found that well developed and long lived nocturnal MCSs often propagated across regions that experienced a relative maximum in surface geostrophic winds and 850 hPa frontogenesis, while shorter lived MCSs had no fronto-genetic support. As mentioned earlier, LLJs would often manifest themselves in the regions where stronger surface and low level geostrophic winds existed, so Caracena (1994) discovered a correlation that exists between strong MCS development and diurnal geostrophic winds (manifesting into a LLJ by evening), where the 850 hPa frontogenesis acted as the terminus for the LLJ. In addition to convergence, frontal overrunning and the isentropic lift of moisture as well as
moisture transport over a stable surface layer at the terminus of the LLJ can also be a source of lift, which may foster the maintenance of both MCSs and Mesoscale Convective Complexes (MCCs) as was noted in Cotton et al. (1989).

Convective initiation (particularly instances of elevated nocturnal convection) is difficult to predict with accuracy. Given that moisture tends to pool at the terminus of the LLJ and that convergence (as explained before) plays a significant role in MCS initiation and sustainment, it would seem appropriate to investigate a variable that includes both moisture and convergence. Horizontal moisture flux convergence (MFC) was explored by Banacos and Shultz (2005) as a diagnostic variable for convective triggers. It was determined that MFC was useful in determining locations of mesoscale boundaries that could aid in convective initiation, but mass convergence was a more appropriate tool to use for an ingredients based methodology given that mass convergence is triggered by several additional factors such as jet coupling and frontogenesis on the synoptic scale. MFC was defined as:

\[ MFC = -\nabla \cdot (qV_h) = -(V_h \cdot \nabla q) - (q \nabla \cdot V_h) \]  

(9)

where \( \nabla \cdot (qV_h) \) represents the total MFC, \( V_h \cdot \nabla q \) represents the advection component, and \( q \nabla \cdot V_h \) is the convergent component of the MFC. A scale analysis completed by Banacos and Shultz (2005) revealed that MFC and its individual components varied greatly based on the temporal and spatial scale to which it was applied. For the spatial and temporal scale of fronts investigated in Banacos and Shultz (2005) (which is explored in the present research), the convergent component dominated, with a magnitude of \( 10^{-3} \text{ g kg}^{-1}\text{s}^{-1} \).
While the LLJ terminus is considered to be an ideal environment for MCS sustenance, some trends in MCS initiation relative to the LLJ terminus have been reported. Kumjian (2006) found that the majority of the 45 MCSs studied initiated in the left exit region of the LLJ. This was believed to be due to a secondary circulation which aided in enhanced lift for convective initiation. After initiation, MCSs propagated across the terminus of the LLJ, where strong theta-e transport and mass convergence were present. These same findings have been further supported by Jirak and Cotton (2007) as well as Schumacher and Johnson (2009).

Lastly, Coniglio et al. (2010) found that the key to MCS sustenance and decay was the veering of the ageostrophic wind. When the ageostrophic component of the total wind was no longer parallel to the direction of MCS motion, MCSs began to decay.

2.6 Other factors involved in MCS development and maintenance

While the LLJ is a key feature in aiding the sustenance of MCS activity, there are many other factors that also influence the development and maintenance of an MCS which require adequate attention. While large scale synoptic forcing can initiate and maintain MCSs as has been discussed in Uccellini and Johnson (1979), identifying triggers for convective initiation with MCSs that form in environments with less synoptic forcing (such as the type A LLJ environments discussed earlier) can be challenging (Schumacher and Johnson 2008). This is especially the case where no solid low level boundaries can be defined, hinting to the idea that no obvious mechanism for convective initiation exists for a forecaster to focus on (Schumacher and Johnson 2009).
Maddox and Doswell (1982), Cotton et al. (1989), and Jirak and Cotton (2007) found that MCSs often initiated in a region where 700 hPa warm air advection (WAA) occurred, suggesting that the ascent associated with the WAA helped trigger and sustain well developed MCSs. In addition to the LLJ, positive 700 hPa temperature advection was considered to be one of the most beneficial ingredients to MCS development and evolution.

MCSs can in fact be initiated by a combination of atmospheric features, with ambiguity often an issue for identifying causes for MCS initiation. Jirak and Cotton (2007), identified nearly twenty initiating factors, including orographic forcing mechanisms, cold fronts, warm fronts, drylines, stationary boundaries, and troughs (along with many others), but for nearly twenty percent of cases, an initiating factor could not be identified.

Strong low level shear has often been considered to be an important ingredient to well-developed, long lived MCSs. Coniglio et al. (2010) noted that weaker shear above 3 km contributed to the lack of updraft organization (as seen in supercells) which would result in deeper cold pools, fostering an environment for cold pool mergers and resultant convection to grow upscale into MCSs, particularly when stronger low level shear was contained in the lower levels (keeping the cold pools from undercutting convection and causing it to diminish). French and Parker (2010) conducted a numerical study by simulating an MCS in the Bryan cloud model (CM1) with 250 m grid spacing and found that increasing low level shear in the 0-1 km layer was the key in balancing convective cold pools and associated outflows while enhancing lift at the leading edge of convection,
promoting MCS longevity. It was argued that evaluating 0-1 km shear was effective compared to 0-3km shear given that the LLJ is contained entirely within this layer, where winds above the LLJ peak would decrease and contribute to negative shear. The suggestion that strong 0-1 km low level counter-balanced the surge of a cold pool from linear convection and promoted enhanced lift along the leading edge of convection has been backed by numerous scientists through extensive research (Rotunno et al. 1988; Weisman 1992; Weisman and Rotunno 2004, and Jirak and Cotton 2007).

Lastly, Snively and Gallus (2014) determined that the orientation of the 0-6 km shear vector played a considerable role in the upscale growth of convection simulated through high resolution modeling. The 0-6 km shear vectors aligned parallel to a lower level boundary (i.e. a surface or 850 hPa front) often encouraged cold pool mergers and subsequent upscale growth of convection, leading to the development of MCS activity with stratiform rain. In one case, it was found that the 850 hPa moisture advection was well simulated in the 4 km WRF, but the 0-6 km shear vector was not aligned parallel to the low level convergent boundary present as was observed in 13 km 00-hr RUC analysis and that the cold pool was confined to a much smaller area. As a result, the WRF did not correctly simulate linear convection or MCS activity which was observed (with the simulated convection specifically lacking stratiform precipitation), hinting to the idea that deep layer shear oriented parallel to low level boundaries may be vital for MCS development in certain circumstances, a point often brought up such as in Dial and Racy (2004) and Schumann and Roebber (2010).
2.7 Using Rapid Update Cycle Analysis as a Substitute for in-situ Observations

Given the lack of spatial and temporal data with the limited in-situ observations available for studying the LLJ and resultant MCS activity, using gridded 00-hr model output from the 13 km RUC analysis as a substitute was deemed necessary (as discussed in chapter 1). Thompson et al. (2003) first employed the RUC analysis to derive soundings in a high horizontal spatial resolution to evaluate environments within close proximity to tornadic supercells. After comparing 149 observed soundings to RUC analysis point soundings, it was determined via an error analysis that the RUC soundings were within reasonable agreement to the observed soundings, rendering the RUC analysis an appropriate tool to use for observational studies. Schumacher and Johnson (2005) also used the RUC analysis to evaluate synoptic and mesoscale atmospheric features that contributed to extreme rainfall producing nocturnal convective events and found that trailing convection and stratiform precipitation occurred in regions of high moisture content with instability, located on the north side of a surface boundary.

Similar to Thompson et al. (2003), Hane et al. (2008) employed the RUC analysis to derive vertical profiles for the ambient environment associated with nearly 50 MCSs given the spatial and temporal gaps in upper air observations. Comparing RUC derived soundings to observed soundings from multiple radiosonde locations across the central plains at 1200 UTC revealed that the wind, temperature and moisture output from the RUC agreed reasonably well with the observed soundings, except that the wind magnitudes below 950 hPa were about half of what observations showed. Besides this one exception, the RUC analysis proved to be a reasonable substitute for in situ data in
this study. Citing Hane et al. (2008) as well as Thompson et al. (2003), Coniglio et al. (2010) used the 20 km hourly RUC analysis to construct composites of 94 nocturnal environments for MCS activity. It was from these RUC analysis composites that Coniglio et al. (2010) was able to determine the relevance of the ageostrophic component of the wind to MCS longevity mentioned earlier.

Given the success of using the RUC analysis in earlier works, Snively and Gallus (2014) used the hourly 20 km RUC analysis to diagnose the stability of the environment for the time and location of MCS initiation for 37 cases and compared them to 4 km WRF output. As such, they were able to conclude that the WRF simulated higher deep layer shear, shallower PBLs and weaker MUCAPE overall than what was observed in the RUC, suggesting that much of the convection in the WRF was elevated in nature and that cases with higher shear present in the model resulted in better simulations of MCSs.

2.8 Implementation of Microphysics in Atmospheric Modeling

With recent increases in computational resources, simulations of MCSs are often now performed using fine enough grid spacing that convective parameterizations are not needed, increasing the importance of accurate MP parameterizations in these models (Weisman and Klemp 1982). Each MP scheme dictates the evolution of convection through the distribution and type of hydrometeors present given the prognostic and diagnostic equations provided (Thompson et al. 2004; Morrison et al. 2009). Bulk hydrometeor distributions are governed by the particle size distribution.
\[ N(D) = N_0 D^\mu e^{-\lambda D} \]  \hspace{1cm} (10)

where \( N(D) \) is the number of particles (as a function of diameter size \( D \) of the particle), \( N_0 \) is the intercept, \( \lambda \) is the slope parameter, and \( \mu \) is the shape parameter of the size distribution given a gamma distribution. In this case, (10) represents that classic exponential (Marshall-Palmer) distribution. For all hydrometeors, a power law (defined in Locatelli and Hobbs 1974) is used for the slope parameter calculation, where the mass and fall speed of a hydrometeor are dependent on diameter size is applied in MP schemes with the particle size distribution

\[ m(D) = a m D^b \]  \hspace{1cm} (11)

where \( m \) is the mass of the hydrometeor, \( D \) is the hydrometeor diameter, \( a \) is the effective density, and \( b \) is the fractal dimension of the particle as defined in Liu (1995).

Hydrometeor velocity relationships are also vital in determining the temporal distribution of both liquid and ice hydrometeor types. Ferrier (1994) developed a hydrometeor velocity relationship that was used in multiple schemes and is defined as

\[ v(D) = \left(\frac{\rho_0}{\rho}\right)^{1/2} \alpha D^\beta e^{-f D} \]  \hspace{1cm} (12)

where \( v \) is the terminal velocity as a function of hydrometeor diameter, \( \rho_0 \) is the surface air density while \( \rho \) is the air density at a given level. The coefficients \( \alpha \), \( \beta \), and \(-f\) are constant parameters that are explicitly defined in Ferrier (1994) for each hydrometeor type.

In the research results to follow, a single and double moment microphysics scheme were employed in the 4 km WRF-ARW. A single moment scheme is purely diagnostic in nature for particle number concentrations, where \( N_0 \) in (10) for rain, snow, ice and
graupel hydrometeors is held constant. A double moment scheme can have a predictive element to it, where the hydrometeor number concentration is prognostic given the $N_0$ relationship in (10)

$$N_0 = \frac{N\lambda^{\mu+1}}{\Gamma(\mu+1)}$$

(13)

where $N$ is the total number of particles, $\lambda$ and $\mu$ are as defined in (1), and $\Gamma$ represents the Euler Gamma Function (Morrison et al. 2009).

The single moment MP scheme employed in this study was the WSM6 devised by Hong et al. (2006). Given the difficulties in simulating ice nuclei in numerical modeling, Hong et al. (2004) first developed an ice scheme that was more fitted to observations by assuming ice nuclei concentrations were only a function of temperature and that ice crystal number concentrations were dependent on the total amount of ice present throughout a given layer. Hong et al. (2006) added graupel to the hydrometeor spectrum, and developed the WSM6 scheme which computes the development and evolution of all hydrometeors by centering all MP sub-processes on the development and movement of graupel in convection as seen in Fig. 1-1.
Given that a purely prognostic MP scheme can be too expensive in computer power and memory to employ in either a research or operational setting (Ferrier 1994), Thompson et al. (2008) improved the single moment liquid MP scheme from Thompson et al. (2006), with a prognostic cloud ice number calculation employed. Cloud ice was on the belief that it had the biggest influence on all other hydrometeors. For cloud ice, (13) was substituted into (14), a refined version of (10) which assumes a normal vs. exponential distribution, with the intercept parameter $N_0$ modified such that:

$$N(D) = \frac{N_t}{\Gamma(\mu+1)} \lambda^{\mu+1} D^\mu e^{-\lambda D}$$

where $N_t$ is the total number of particles in the distribution, and D, $\lambda$, and $\mu$ are the same as in (9).
For cloud water, a critical value of $N_c$ was set to determine when cloud development was initiated based on aerosols in air above the boundary layer. For Thompson et al. (2008), $N = 100 \text{ cm}^{-3}$ was chosen. For rain number concentration to remain a diagnostic variable, (14) is still applied, but when $\mu > 0$ for rainwater, $N_0$ can no longer be defined explicitly, and is thus determined by the relationship

$$N_{0,r} = \frac{(N_1-N_2)}{2} \tanh \left( \frac{q_{tr}-q_{r}}{4q_{r0}} \right) + \frac{(N_1+N_2)}{2}$$

where $N_1 = 9 \times 10^9 \text{ m}^{-4}$, $N_2 = 2 \times 10^6 \text{ m}^{-4}$ and $q_{r0} = 1 \times 10^{-4} \text{ kg kg}^{-1}$ are the upper/lower intercept limits and transition value limit, respectively. The definition of the intercept parameter for the graupel size distribution is calculated by using the graupel mixing ratio. Taking into account that convective updrafts produce higher amounts of graupel, the intercept parameter calculation was modified such that

$$N_{0,g} = \max \left( 10^4, \min \left( \frac{100}{q_g}, 10^6 \right) \right)$$

where the graupel density $q_g$ is constant and set to 400 kg m$^{-3}$. Lastly, snow was calculated using a rescaling technique utilized in Field et al. (2005), where the snow number concentration was defined as:

$$N(D) = \frac{M_2^4}{M_3^{3.4}} \left[ k_0 e^{-\frac{M_3^2 A_0}{M_3^2}} + k_1 \left( \frac{M_2}{M_3} D \right)^{\mu_s} e^{-\frac{M_2^2 A_1}{M_3^2}} \right]$$

where $k_0 = 490.6$, $k_1 = 17.46$, $A_0 = 20.78$, and $A_1 = 3.29$, and $\mu_s$ is as defined in Thompson et al. (2008), and $M_n$ is the nth moment of the distribution:

$$M_n = \int D^n N(D) dD$$
MCSs along with many other forms of linear convection or convective clusters often have trailing areas of stratiform precipitation associated with them. Morrison et al. (2009) used both a single and double moment MP scheme to simulate linear convection with trailing stratiform rain and found that the stratiform precipitation was more prevalent in the double moment simulations. This was because the double moment scheme had greatly reduced evaporation rates in the region of trailing stratiform rain, allowing much of the precipitation to reach the ground. In addition, enhanced evaporation in portions of the convective updrafts allowed for the weakening of updraft strength and incomplete consumption of atmospheric instability, allowing for positively buoyant air parcels to be exhausted from the storm updrafts to regions where stratiform rainfall would occur. This would promote regions of enhanced rainfall behind linear convection. Such effects can have a significant impact on precipitation from MCSs in the Great Plains.

2.9 High Resolution Modeling and the Planetary Boundary Layer

The Great Plains LLJ is dependent on boundary layer processes, and given that this current research included simulating LLJs in the 4 km WRF, where the boundary layer must be parameterized through the use of a PBL scheme given the coarse temporal and spatial resolution of a model grid with respect to the scale of turbulence (Coniglio et al. 2013), a critical review of the three PBL schemes employed in the present research was considered important. The PBL consists of the 1-2 km deep layer of the atmosphere that is closest to the Earth’s surface. This layer is heavily influenced by friction induced by the planet’s terrain and by vertical heat fluxes often induced by incoming solar radiation.
Given the chaotic nature of these fluxes in the PBL, turbulence is generated and redistributed across the PBL (Stull 2009). Turbulence is considered to be one of the most important processes to the PBL (Nakanishi 2001) given that turbulent fluxes are a primary source of temperature, moisture and momentum distribution in the vertical (Coniglio et al. 2013). Generally, PBL schemes are divided into local and non-local mixing schemes, where local mixing schemes base their calculations of eddy diffusivity and turbulent mixing based on data at a given point, while non-local schemes determine PBL variables based on data from surrounding points (Holtslag and Boville 1993).

The MYJ scheme is the adjustment of the Mellor Yamada (MY 2.5) scheme, where a viscous sub layer of the PBL was added to simulations that operated on the eta grid as defined in Janjic (1994). The MY 2.5 scheme defined the state of the boundary layer through a prognostic turbulence kinetic energy (TKE) equation (19) defined below:

\[
\frac{d}{dt} \left( \frac{q^2}{2} \right) = \frac{\partial}{\partial z} \left( \rho q \frac{\partial q^2}{\partial z} \right) + P_s + P_b - \epsilon \tag{19}
\]

\[
P_s = -\langle wu \rangle \frac{\partial U}{\partial z} - \langle wv \rangle \frac{\partial V}{\partial z} \tag{20}
\]

\[
P_b = \beta g \langle w \theta v \rangle \tag{21}
\]

\[
\epsilon = \frac{q^3}{B_1 l} \tag{22}
\]

\[
\langle wu \rangle = -K_M \frac{\partial U}{\partial z} \tag{23}
\]

\[
\langle wv \rangle = -K_M \frac{\partial V}{\partial z} \tag{24}
\]

\[
\langle w \theta v \rangle = -K_H \frac{\partial \theta v}{\partial z} \tag{25}
\]
\[ \langle wS \rangle = -K_H \frac{\partial S}{\partial z} \]  
\[ K_{M,H} = lqS_{M,H} \]  
\[ A_2 = S_M [6A_1A_2G_M] + S_H [1 - (3A_2B_2G_H) - (12A_1A_2G_H)] \]  
\[ A_1 = [S_M [1 + (6A_1^2G_M)] - (9A_1A_2G_H)] - S_H [12(A_1^2G_H) + (9A_1A_2G_H)]/(1 - 3C_1) \]  
\[ G_M = \frac{l^2}{q^2} \left( \frac{\partial U^2}{\partial z} + \frac{\partial V^2}{\partial z} \right) \]  
\[ G_H = -\frac{l^2}{q^2} \beta g \frac{\partial \theta_v}{\partial z} \]  

Here, the brackets \( \langle \ \rangle \) denote fluxes or covariances (through previous equations and henceforth in this literature review). The quantity \( (q^2/2) \) is the TKE, \( S_q = 0.20 \) and \( \beta = 0.004 \) (both unitless), \( l \) is a general length scale, \( P_s \) and \( P_b \) are TKE production terms by shear and buoyancy, respectively, which are calculated using fluxes determined by Fickian diffusion simplifications with eddy diffusivity coefficients (23)-(26). \( \varepsilon \) is the dissipation of TKE (22). The vertical turbulent exchange coefficients for the turbulent momentum and heat fluxes are as defined in (27). Constants \( A_1, A_2, B_1, B_2, \) and \( C_1 \) were determined empirically from past research. \( G_m \) and \( G_H \) are shear and buoyancy parameterizations that are dependent on the general length scale selected. While the structure of this scheme is quite straightforward, the complexity behind PBL evolution has often introduced conflicts for applying this scheme outright as determined in several studies, including Janjic (1990). Adding the viscous sub layer is one of numerous changes made to MY 2.5 that defines the MYJ scheme. The viscous layer, cited by both Liu et al. (1979) and Janjic (1994) is defined as
\[ U_z - U_{sfc} = D_1 \left( \frac{F_U}{u_*} \right) \left[ 1 - \exp \left( -\frac{z u_*}{D_1 \nu} \right) \right] \]  
(32)

\[ \theta_z - \theta_{sfc} = D_2 \left( \frac{F_\theta}{u_*} \right) \left[ 1 - \exp \left( -\frac{z u_*}{D_2 X} \right) \right] \]  
(33)

\[ q_z - q_{sfc} = D_3 \left( \frac{F_q}{u_*} \right) \left[ 1 - \exp \left( -\frac{z u_*}{D_3 \gamma} \right) \right] \]  
(34)

where (32), (33) and (34) define the sub-layer for momentum, potential temperature, and atmospheric water vapor, respectively. D_1, D_2 and D_3 are parameters (defined in more detail in Janjic (1994)). \( u_* \) is the friction velocity. \( F_U, F_\theta, \) and \( F_q \) are the turbulent fluxes of momentum, heat and water vapor, respectively, while \( \nu, X, \gamma \) are the associated molecular diffusivities for momentum, heat and water vapor. The eta model that employed the MY 2.5 scheme with the addition of a viscous layer was tested on a spurious rainfall event and a tropical storm (with a successful 36-hour forecast), and it was concluded that the eta model simulations had performed reasonably well, allowing for the MYJ scheme to be incorporated into the operational eta model by 1993. Nearly a decade later, Janjic (2001) updated the MY 2.5 scheme in the NCEP Meso model by incorporating changes to the general length scale, which impacted the structure of the MYJ scheme given its strong relation to the MY 2.5 scheme.

The MYNN local scheme adds three additional prognostic equations and allows for counter gradient fluxes, with significant emphasis on pressure and buoyancy covariances along as well as stability effects throughout the PBL at the turbulent length scale (Nakanishi 2009). As such, it utilizes liquid water potential temperature (\( \theta_l \)) and atmospheric total water content \( q_w \) (defined as both atmospheric water vapor and liquid
contents combined) as the basic thermodynamic variables. As such, the governing equations for the second turbulent quantities are

\[
\frac{\partial q^2}{\partial t} = -\frac{\partial}{\partial z}\left[w\left(\frac{u^2+v^2+w^2+2p}{\rho_0}\right)\right] - 2\left[\langle uw \rangle \frac{\partial u}{\partial z} + \langle vw \rangle \frac{\partial v}{\partial z}\right] + 2 \frac{\partial}{\partial \theta_v} \langle w\theta_v \rangle - 2\varepsilon \tag{35}
\]

\[
\frac{\partial (\theta_i^2)}{\partial t} = -\frac{\partial}{\partial z} (w\theta_i^2) - 2\langle w\theta_i \rangle \frac{\partial \theta_i}{\partial z} - 2\varepsilon_{\theta_i} \tag{36}
\]

\[
\frac{\partial (\theta_1 q_1)}{\partial t} = -\frac{\partial}{\partial z} (w\theta_1 q_w) - \langle wq_w \rangle \frac{\partial \theta_1}{\partial z} - \langle w\theta_1 \rangle \frac{\partial q_w}{\partial z} - 2\varepsilon_{\theta q} \tag{37}
\]

\[
\frac{\partial (q_w^2)}{\partial t} = -\frac{\partial}{\partial z} (wq_w^2) - 2\langle wq_w \rangle \frac{\partial q_w}{\partial z} - 2\varepsilon_{q w} \tag{38}
\]

where (35) represents the turbulent quantity of TKE, (36) represents the covariance for the turbulent quantity of liquid water potential temperature, (37) is the covariance between the liquid water potential temperature and atmospheric total water content, and (38) is the covariance of the turbulent quantity of total water content. The variables \( p, \rho_0, \theta_v, \theta_1, \) and \( Q_w \) are pressure, density, virtual potential temperature, horizontal advection (derived from the thermal wind relationship) and vertical atmospheric total water content flux, respectively. The MYNN scheme uses this body of equations in an attempt to solve for the unknown second order fluxes \( \langle uw \rangle \) and \( \langle w\theta_1 \rangle \). The difference between the 2.5 and 3.0 schemes is that the MYNN 3.0 scheme uses prognostic equations to predict TKE \( (q^2/2) \) and variance in temperature \( (\theta^2) \), while the MYNN 2.5 scheme only provides a prognostic equation for TKE (Mellor and Yamada 1982). Specifically, TKE is as specified in (19) and the variance in temperature is:

\[
\frac{D(\theta^2)}{Dt} = \frac{\partial}{\partial z} \left[ lqS_{\theta} \frac{\partial (\theta^2)}{\partial z} \right] - 2\langle w\theta \rangle \frac{\partial \theta}{\partial z} - \frac{2q}{\Lambda_z} \langle \theta^2 \rangle \tag{39}
\]
where $S_\theta$ is a constant, $l$ and $\Lambda_2$ are length scales, $\theta^2$ is the temperature variance, and $\Theta$ is the potential temperature in the MYNN 3.0 scheme. For MYNN 2.5 though, potential temperature variance is diagnosed as:

$$\langle \theta^2 \rangle = -\frac{\Lambda_2}{q} \langle w\theta \rangle \frac{\partial \Theta}{\partial z}$$

(40)

Nakanishi (2009) pointed out that both the MYNN 2.5 and 3.0 schemes improve the development of a convective boundary layer, and better estimates the magnitude of TKE and the turbulent length scale. This is due to the MYNN 2.5 and 3.0 schemes relying heavily on the new parameterizations of the pressure covariances and stability functions for third order turbulent fluxes mentioned earlier.

The Yonsei University (YSU) scheme is non-local in nature, where the state of the boundary layer is determined by surrounding point data. The YSU scheme factors in entrainment processes at the top of the PBL. The YSU scheme is governed by the a single prognostic equation, where variables ($u$, $v$, $\theta$, $q$) within the mixed layer are substituted into:

$$\frac{\partial C}{\partial t} = \frac{\partial}{\partial z} [K_c \left( \frac{\partial C}{\partial z} - \gamma_C \right) - (w'c')_h \left( \frac{z}{h} \right)^3]$$

(41)

where the PBL height ($h$) is defined by the level at which a minimum flux (lowest $(w'c')_h$) exists at the inversion layer. $K_c$ represents the eddy diffusion coefficient, and $\gamma_C$ is a correction to the local gradient of the chosen variable entered into (41), which includes the effect of large scale eddies to the total flux. $(w'c')_h$ is the flux of the subject variable at the inversion layer. The key difference between the YSU and older versions of
non-local mixing schemes is that \((\overline{w'}c')h\left(\frac{x}{h}\right)^3\) is explicitly defined vs. being implicitly defined in previous non-local schemes. The YSU scheme thus causes boundary layer mixing when the PBL becomes absolutely unstable (i.e. a free convection regime) and decreases mixing when a wind shear driven convection regime dominates the PBL. As such, excessive mixing (present in previous schemes with medium range forecasts) is resolved and rapid overgrowth of the PBL is nearly eliminated (Hong et al. 2006).

Studies have recently been conducted with sensitivity tests of high resolution WRF runs with multiple PBL schemes. Hu et al. (2010) employed a nested WRF (with 4 km grid spacing for the inner most nest of a region over eastern Texas, west Louisiana and southwest Arkansas), with the MYJ, YSU and Asymmetric Convective Model II (ACM2) employed. Using over 92 sets of daily simulations in comparison to surface observations across the nested domain, it was determined that the YSU and ACM2 schemes performed well compared to observations. The MYJ was associated with a cool, moist bias to its simulated PBL, mainly due to the fact that it did not factor in entrainment of air from above the PBL top. Running the 4 km WRF, Coniglio et al. (2013) simulated the PBL with three local schemes (MYJ, MYNN 2.5, QNSE) and two non-local schemes (ACM2, YSU). Similar to Hu et al. (2010), the PBL’s in the morning were too dry and cool (except the YSU). The local schemes (except the MYNN 2.5) were too moist in comparison to the non-local schemes. Still, the MYNN 2.5 scheme performed the best overall with the smallest amount of bias associated with temperature and moisture, allowing Coniglio et al. 2013 to conclude that the MYNN 2.5 scheme was an optimal choice when running high resolution convective allowing models.
CHAPTER 3: WRF FORECAST SKILL OF THE GREAT PLAINS LOW-LEVEL JET AND ITS CORRELATION TO FORECAST SKILL OF MESOSCALE CONVECTIVE SYSTEM PRECIPITATION

by

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3.1 Abstract

The Great Plains low-level jet (LLJ) fosters an environment that is supportive of nocturnal mesoscale convective systems (MCSs) across the central U.S. during the summer months. The goal of this research is to determine if forecast skill in the LLJ correlates with forecast skill in MCS precipitation in high spatial resolution (4 km) WRF runs. It was found that the forecast skill of the geostrophic and ageostrophic wind direction components correlated to forecast skill of MCS precipitation. The diurnal geostrophic wind maximum across the plains (induced by terrain sloping and heating) set up the background flow for the LLJ to develop and established the orientation of the LLJ (which would impact MCS evolution). The forecast skill in the ageostrophic wind direction correlated with MCS forecast skill since the ageostrophic winds induced convergence at the terminus of the LLJ, which benefited MCS longevity only if the ageostrophic wind held a perpendicular component to the MCS. Forecast skills of 700 hPa temperature advection, mixed layer convective available potential energy, mixed layer convective inhibition, 0-1 km, 0-3km and 0-6 km shear along with surface and 850 hPa frontogenesis and horizontal moisture flux convergence were compared to MCS precipitation forecast skill to see if correlations in non-LLJ parameters existed. No forecast skill correlations were found with the non-LLJ parameters.

3.2 Introduction

Mesoscale Convective Systems (MCSs) are the primary source of precipitation across the Great Plains and Midwest during the summer months (Jirak and Cotton 2007;
Coniglio et al. 2010; French and Parker 2010) and provide the rainfall needed for agricultural purposes (Schaffer et al. 2011) thus, better forecasts of nocturnal convection benefits farmers (Stensrud and Fritsch 1993, Jirak and Cotton 2007). Nocturnal MCS development and sustenance is often tied to the occurrence of the Great Plains southerly Low-Level Jet (LLJ) (Cotton et al. 1989; Augustine and Caracena 1994; Mitchell et al. 1995; Higgins et al. 1997; French and Parker 2010), a narrow current of air with a wind speed maxima located between 150 and 2000 m above ground level (AGL) (Bonner and Peagle 1970; Mitchell et al. 1995; Whiteman et al. 1997; Song et al. 2005). As such, better simulations of the LLJ may yield better forecasts of MCS precipitation (Mitchell et al. 1995; French and Parker 2010).

Over recent years, higher spatial and temporal resolution in weather prediction models has allowed better numerical simulations of weather phenomena including MCSs (French and Parker 2010; Snively and Gallus 2014) and boundary layer evolution, which impacts the development of the southerly LLJ (Hu et al. 2010, Coniglio et al. 2013). It was found in a 250 m horizontal grid spacing theoretical cloud modeling study (French and Parker 2010) that longer lived MCSs were influenced by the storm-relative inflow and low level shear that were induced by the addition of an idealized LLJ. Strong low level shear kept convection from being undercut by outflow induced by the cold pool and weakening. Snively and Gallus (2014) evaluated the convective modes of storms simulated in the 4 km WRF using 00-hr 13 km RUC analyses as substitutes to in-situ observations given the extremely coarse spatial and temporal resolution of radiosonde and vertical profiler data. It was found that linear convective modes with stratiform precipitation were better simulated when the deep layer shear vectors were oriented
parallel to low level boundaries or frontogenesis (Dial and Racey 2004; Schumann and Roebber 2010). This would encourage the congealment of cold pools, fostering the upscale growth of surface based convection into MCSs during the evening hours.

Hu et al. (2010) (using surface observations and aircraft data for observations) and Coniglio et al. (2013) (using 13 km RUC analyses for observations) ran the WRF under multiple local and nonlocal mixing-based Planetary Boundary Layer (PBL) schemes and found that the Mellor Yamada Janjic (MYJ) scheme resulted in a boundary layer that was too cool and moist. As such, Hu et al. 2010 found that the MYJ simulated a low level wind maximum (likely the entrance region of a LLJ) too strong in nature, while the YSU scheme was too weak with regards to wind magnitude. The MYJ scheme did not factor in entrainment from the top of the PBL and thus its boundary layer depth became too shallow, too quickly, likely promoting the strong wind speeds rapidly developing above the PBL. It should be noted that this study was conducted in south Texas during the summer months, hence the core of the LLJ was not sampled and conclusions on PBL schemes with respect to LLJ development could not be made. Coniglio et al. (2013) found that the MYNN 2.5 scheme did the best for simulating the PBL in comparison to the RUC analyses, thus being the scheme most desired for high resolution modeling for pre-convective environments.

While the previous works highlighted conditions that made MCS development favorable or PBL conditions favorable for LLJ development, little, if any, work has been done to simulate LLJs and subsequent MCS development in real cases using high (vertical and horizontal) spatial and temporal resolution. The present research aims to
simulate real cases of LLJ and subsequent MCS development and evolution with high vertical and horizontal resolution using the 4 km WRF with multiple PBL and microphysics (MP) schemes. It is hypothesized that cases with better simulations of the LLJ would yield better MCS precipitation forecasts. As such, correlations of forecast skill of the LLJ and forecast skill in MCS precipitation are emphasized in the present research.

3.3 Data and Methodology

With in-situ observations of winds in the lower troposphere lacking for validation of high spatial resolution modeling, 13 km RUC analyses were substituted for in-situ data, a research decision often implemented with success (e.g., Thompson et al. 2003, Hane et al. 2008, Schumacher and Johnson 2009, Coniglio et al. 2010, and Snively and Gallus 2014).

The present research employed the WRF-ARW (version 3.5) model with 4 km horizontal grid spacing and 50 vertical levels, with 28 levels below 850 hPa having 5 hPa grid spacing. The selected domain spanned a 1600 km x 1600 km area across the eastern Great Plains and Midwest states (Fig. 3-1).

20 cases were chosen based on a strong LLJ being present in the RUC analyses, with subsequent MCS development (determined via archived mosaic radar data) as well as availability of Lamont, OK vertical profiler data, and 12 km NAM forecast output used to initialize the WRF lateral and boundary conditions. In terms of wind magnitudes, LLJs were classified based on Bonner’s (1968) criteria. LLJs were also classified based on their synoptic environment (Chen and Arritt 2013, personal communication). A
streamline analysis with wind magnitudes of the 900 hPa and 200 hPa fields along with overlaid 900 hPa and 200 hPa divergent fields were employed to determine LLJ type. The evaluation of coupling of 900 hPa convergence and 200 hPa divergence and associated implication of LLJ type stems from the work of Uccellini and Johnson (1979), who stated that LLJs and upper level jet streaks are coupled in strongly forced synoptic setups, where the convergent exit region of a LLJ is overrun by a divergent entrance region of an upper level jet streak, allowing for large scale vertical motion (that often promotes widespread storm organization and severe weather). Cases where strong 900 hPa southerly flow (where magnitudes reached or surpassed 15 ms$^{-1}$) and strong cyclonic flow at 200 hPa (where magnitudes reached or surpassed 30 ms$^{-1}$) are present and where 900 hPa convergence is coupled with 200 hPa divergence are considered cyclonic LLJ events (Fig. 3-1) (referred to henceforth as type C LLJs). Cases where strong southerly flow is still present at 900 hPa but with weak anticyclonic flow at 200 hPa and little to no jet coupling are considered anticyclonic LLJs or type A (Fig. 3-2), where the inertial oscillation (Blackadar 1957), terrain sloping and heating (Wexler 1961; Bonner and Paegle 1970, Pan et al. 2004; Parish and Oolman 2010), and large scale monsoonal southerly flow induced by the summertime Bermuda High (Chen and Kpaeyeh 1993) play the biggest roles in forcing the LLJ. Cases where at least some synoptic forcing appeared evident (i.e. an approaching or departing upper level jet streak or longwave trough) but were not purely type C or type A in nature were considered transition cases, or type T LLJs. This category was the most subjective in nature.

MCSs were classified as convective clusters that encompassed at least a 50 km x 5-10 km area of 45+ dbZ cores with lifespans of at least 3 hours. MCSs that were located
closer to the center of the WRF domain were selected to minimize influences from lateral boundaries.

Sensitivity tests were also performed to understand the role of MP and PBL schemes in the correlation of LLJ errors and MCS precipitation errors. To determine if hydrometeor distribution would affect the correlation of forecast skill, WRF runs for each case were conducted under two different MP schemes. The WRF-Single Moment 6-class (WSM6) scheme was chosen because it was specially developed for high resolution simulations involving ice, graupel and snow processes in atmospheric cloud simulations by revolving hydrometeor development processes around the production/distribution of graupel rather than using predictive equations (Hong et al. 2006). The Thompson MP scheme (defined in Thompson et al. 2008) uses prognostic ice and rain number concentrations with time split fall terms to determine hydrometeor evolution with time. To understand how PBL development might influence the LLJ and the correlation of skill, three PBL schemes were utilized: the MYJ, MYNN 2.5, and YSU. The MYJ is a local scheme (MY 2.5 scheme revised) which uses a prognostic calculation for turbulence, with the addition of a viscous sub-layer to the PBL through molecular diffusion (Janjic 1994). The local mixing based MYNN 2.5 scheme uses a prognostic equation to calculate the turbulent kinetic energy in the boundary layer, but leaves thermodynamic variables as diagnostic in nature (Mellor and Yamada 1984; Nakanishi 2001). The YSU scheme is a nonlocal mixing scheme which evaluates entrainment of air into the PBL mixed layer from above the inversion (Hong et al. 2006).

Mean Absolute Error (MAE) was computed to measure the difference between the 13 km RUC-Analysis and the 4 km WRF simulations for some parameters such as LLJ
magnitude. The 4km WRF data were re-gridded via interpolation to the 13 km grid used in the RUC analyses. For determining the forecast skill of the LLJ, all grid points outside of the 65\textsuperscript{th} percentile criteria for the magnitude of the total wind and with a northerly direction or negative v component were filtered out of the regridded WRF and RUC-Analysis outputs. Subjective analysis revealed that the 65\textsuperscript{th} percentile captured the wind magnitudes that roughly fit criteria I-III defined in Bonner (1968). The directional component of the total wind and directional and magnitude components of the ageostrophic and geostrophic winds, potential temperature and atmospheric water vapor content were calculated only at the same points as the total wind magnitude (so that the MAE of these variables are calculated in the LLJ exclusively). The MAE was defined as:

$$MAE = \frac{1}{n} \sum_{i=0}^{n} |X_i - Y_i|$$

(1)

where n is the total number of grid points at or above the 65\textsuperscript{th} percentile of the total wind magnitude in both the RUC and the WRF in a 200-2000 m (at 100 m intervals) layer, X is the WRF output at a given grid point i, and Y is the RUC-Analysis output at the same grid point i. In addition, the MAE was calculated for the magnitude of shear in the 0-1km, 0-3km and 0-6km layers, mass convergence and horizontal moisture flux convergence (MFC), 700 hPa positive temperature advection, 850 hPa frontogenesis and surface frontogenesis, mixed layer convective available potential energy (MLCAPE) and mixed layer convective inhibition (MLCIN). The 0-3km and 0-6 km shear were calculated using the bulk wind difference between the surface and 3000 m and 6000 m respectively. With 0-1 km shear, mass convergence, MFC and frontogenesis (induced by boundaries and the terminus of the LLJ), the grid-points selected to calculate the MAEs were within a 1000 km x 1000 km area (at heights from 200-2000 m at 100 m intervals).
centered around the point of MCS initiation. 700 hPa temperature advection, MLCAPE, and MLCIN MAEs were calculated the same way given that the average 700 hPa positive temperature advection maximum and MLCIN minimum occurred at the MCS initiation point (Cotton et al. 1989) in the RUC analyses. All MAE calculations were performed at 0300 UTC, 0600 UTC and 0900 UTC.

STAGEIV 6-hourly precipitation observations from derived satellite/composite radar analyses (Lin 2005) were also re-gridded to the 13 km grid and compared to WRF simulated 6-hourly rainfall to determine the WRF precipitation skill, using the Equitable Threat Score (ETS) and bias, defined as:

\[
ETS = \frac{H - \frac{FO}{N}}{F + O - H - \frac{FO}{N}}
\]  
\[
Bias = \frac{F}{O}
\]

where H is the number of grid points correctly forecast to have rainfall exceeding a certain threshold, O is the number of points observed to exceed the threshold, F is the number of points forecast by the WRF to exceed the threshold, and N is the total number of grid-point calculations in the WRF 13 km re-gridded domain. Higher ETSs indicate better skill at forecasting precipitation. ETSs were calculated at the 0.1 inch threshold for a sub region of the WRF domain that was only affected by MCS activity. ETSs were calculated for the 0000-0600 UTC, 0300-0900 UTC and 0600-1200 UTC time periods in order to capture the evolution of precipitation during the developmental, mature and dissipation stages of the MCSs. LLJ MAEs at 0300 UTC, 0600 UTC, and 0900 UTC
were compared to ETS values in these three time periods to explore correlation during various LLJ/MCS stages. It was mentioned by Mason (1989), Hamill (1999) and Mesinger (2008) that a model wet bias could artificially inflate ETSs, diminishing the validity of using the ETS in such cases. As such, ETS scores were ranked with their associated biases to explore if such behavior was present.

The square of the Pearson Correlation Coefficient ($R^2$) was used to determine the magnitude of correlation between forecast skill of the LLJ (MAEs) and forecast skill of MCS precipitation (ETSs) in 4km WRF runs involving all 20 cases. In Mendenhall and Sincich (2007), $R^2$ is defined as:

$$R^2 = \frac{\sum_{i=1}^{n}(x_i - \bar{x})(y_i - \bar{y})^2}{\sum_{i=1}^{n}x_i^2 \sum_{i=1}^{n}y_i^2}$$  \hspace{1cm} (4)

where $i$ represents each grid point out of $n$ total number of gridpoints of the sample size, $x_i$ is the observed value at a given point, $y_i$ is the forecast value at a point, and both $\bar{x}$ and $\bar{y}$ are the sample means of the observed and forecast variables. Since bias and sampling variation problems can exist with the use of $R^2$ (Helland 1987), where deceivingly low values may occur in association with a model involving a subject of low predictability given small samples sizes (as was the case with this research), the p-value was employed to determine if a correlation existed and how significant it was. A p-value at or below 0.05 (where the confidence interval at rejecting the null-hypothesis was set at or above 95%) was the threshold for determining if an $R^2$ correlation was significant. Helland (1987) among others cautioned trusting an $R^2$ value alone and advised graphical analysis for checking results. In the present study, scatterplots with a line of best fit were generated and evaluated to validate $R^2$ results. Significant $R^2$ values that were associated
with a negative trend (higher ETS for lower MAE) were bolded in font (Tables 3-3 through 3-8). It is important to note that the $R^2$ value does not conclude that the forecast skill of one variable dictates the forecast skill of another variable. Rather, $R^2$ depicts a linear trend between the forecast skills of two variables, suggesting an association between both variables.

Composite analyses were conducted in the same manner described in Cotton et al. (1989). The average location of MCS initiation was overlaid with 700 hPa temperature advection, MLCAPE, MLCIN and 200-2000 m (at 100 m intervals) averaged fields of atmospheric water vapor mixing ratio, total wind and mass convergence. For each case, the subdomains were selected in the same manner as the shears, and mass convergence MAEs.

A Gaussian filter was applied to the WRF base variables to eliminate atmospheric features with a wavelength less than $2\Delta x$ of the coarser RUC analyses (26 km) so that the environment could be evaluated on the scale of the RUC for a fair comparison.

3.4 Results

a) Characteristics of the observed LLJ

A nearly even distribution of type C, A, and T LLJs was observed among the 20 cases (Table 3-1), allowing for a variety of ambient environments influencing the LLJ to be represented. Most of the LLJs peaked around 0900 UTC (with the majority of the remaining cases peaking around 0600 UTC), a finding that concurs with Bonner and Paegle (1970), Mitchell et al. (1995), and Whiteman et al. (1997). Most of the LLJs (regardless of their synoptic environment) matched Bonner III criterion (with an overall
average peak wind of 27.30 ms\(^{-1}\)). Mitchell et al. (1995) and Kumjian et al. (2006) also found that most LLJs in their studies fit the Bonner III criteria, possibly because they sampled a large proportion of type C LLJs. In addition, the average height of the peak wind for all 20 cases was 878 m, similar to what Bonner (1968) and Mitchell et al. (1995) observed (peak magnitude heights at 785 m and approximately 1000 m, respectively), but much higher compared to Whiteman et al. (1997) and Song et al. (2005) (596 m and approximately 300 m, respectively). It is important to note though that Bonner (1968) only had 13 pilot rawinsondes to sample the LLJ across the Great Plains, and that the 31 low powered vertical profilers in the Wind Profiler Demonstration Network employed by Mitchell et al. (1995) only sampled wind magnitudes above 500 m. In addition, Whiteman et al. (1997) employed a cross-chain wind finding technique using only available rawinonde and vertical profiler data (where at least three in-situ measurements had to be simultaneous for an observation to be recorded), and the data quality became questionable closer to the surface. As such, the height of the true LLJ peak may have been missed. The same can definitely be said for Song et al. (2005), who only used hourly, 915 MHz data at 60 m intervals (beginning just above ground level) from only two stations that were within 50 km of each other. The LLJ peak wind slopes northwesterly with height in the RUC analyses in nearly all of the 20 cases investigated (Fig. 3-3), and assuming that the LLJ is depicted with relative accuracy, this sloping effect may explain the reason for the LLJ peak being found higher aloft in the present study compared to some past research.

Type A LLJs had a lower peak magnitude (25.38 ms\(^{-1}\)) than type C or type T LLJs (26.89 ms\(^{-1}\) and 29.77 ms\(^{-1}\), respectively), and also had a lower peak altitude (692 m)
compared to type C and T LLJs (863 m and 1083 m, respectively). Type C LLJs were overall stronger and located higher aloft than type A, likely due to the influence of large scale dynamic features such as jet coupling from isentropic up-glide or cyclogenesis, as explained in Uccellini and Johnson (1979). This was found to be the case in Hoecker (1963) and Bonner (1966), where LLJ magnitudes and height of the speed maxima could extend past the boundary layer to roughly the 850 hPa level. What is somewhat unusual is that type T LLJs had the strongest average peak wind and average peak altitude and not type C jets, which were more synoptically influenced. Further research is needed to understand this difference.

\[b) \text{ Correlation of forecast skill of LLJ variables with forecast skill of MCS precipitation in 4 km WRF runs}\]

As seen in the Thompson-MYJ runs (Table 3-2), only a few cases had a bias above 1.0, showing that only a few cases represented a moist bias. The Thompson-MYNN, Thompson-YSU and WSM6 (MYJ, MYNN and YSU) schemes at all time-intervals displayed similar behavior.

No discernable correlations could be made between the forecast skill of the total, geostrophic or ageostrophic wind magnitudes (not shown) within the LLJ and forecast skill of MCS precipitation in the WRF. Correlations between forecast skill of the geostrophic and ageostrophic directional components of the LLJ winds and forecast skill of MCS precipitation, however, were noted. The geostrophic wind direction forecast skill in the LLJ was more correlated to MCS precipitation than all other variables (Tables 3-3,
3-4 and Fig. 3-4a). All tested WRF runs showed at least weak to moderate correlations, with an exception being Thompson-MYNN where no correlation was found in the early stages of LLJ and MCS development. Composites between the RUC analyses and WRF runs of 2100 UTC geostrophic winds and peak time of LLJ winds, convergence and atmospheric water vapor (all within an averaged 200-2000 m layer), show that regions of highest magnitude and terminus of the geostrophic winds align well with the wind maximum and terminus of the LLJ (Figs. 3-5 and 3-6). Bonner and Paegle (1970) and Parish and Oolman (2010) found that the diurnal geostrophic wind maximum (influenced by terrain sloping and heating) provided a background wind profile for the LLJ, despite slightly weakening in the evening due to the change in direction (aloft) of the thermal wind. The results in Figs 3-5 and 3-6 support Bonner and Paegle as well as Parish and Oolman’s claims. In addition, Augustine and Caracena (1994) noted that nocturnal precipitation was often most widespread north of the diurnal geostrophic wind and eventual LLJ maximum, which was noted in composites of 12-hr precipitation in observations and all WRF runs (Fig. 3-7), lending further credibility to the claim that the diurnal geostrophic wind profile contributes to LLJ and subsequent MCS evolution.

While not as strong of a correlation for all WRF runs, the 0300 UTC MAE of the ageostrophic wind direction correlates well with the 0600-1200 UTC MCS precipitation ETSs with all WSM6 WRF runs (Tables 3-5 and 3-6). This suggests that forecast skill of the divergent wind early on in LLJ development can be associated with forecast skill in MCS evolution in the WRF (Fig. 3-4b). Fig. 3-6 shows that the centroid for the initiation of all observed and simulated MCSs occurs within or just upstream the maximum region of convergence (induced by the directional component of the ageostrophic wind),
suggesting that air mass convergence at the LLJ terminus plays a key role in MCS sustenance (Coniglio et al. 2010). Earlier, Bonner (1968) suggested that low level convergence was mostly important for extending the life time of an MCS that was already in progress. In the present research, it was noted in the observations and WRF runs that for individual MCSs, the veering of the LLJ wind direction alone led to MCS weakening or dissipation (also noted by Coniglio et al. 2010), meaning that the convergence had to originate from the ageostrophic winds that had a perpendicular component to the MCSs (discussed in more detail shortly). This can be seen by comparing the observed and simulated reflectivities of a given MCS (Figs. 3-8 and 3-9, respectively) with the associated LLJ profile in the RUC analyses and WRF (Figs. 3-10 and 3-11, respectively). In the observations, the first MCS moving into an environment of veered winds weakens and a new MCS develops to the west and matures in a region where the winds remain backed at the LLJ terminus. It is currently not understood, however, why ageostrophic wind direction correlations were not observed in WRF runs using the Thompson MP. It was also noted that later MAEs in the ageostrophic winds correlated with MCS precipitation ETSs for the 0600-1200 UTC time frame in the WSM6-MYJ runs.

MAEs of potential temperature within the LLJ at early stages (e.g. 0300 UTC) correlated well with early evening MCS precipitation ETSs (0000-0600 UTC and to an extent, 0300-0900 UTC; Fig. 3-4c). Tables 3-7 and 3-8 show relatively strong correlations under the WSM6 WRF runs, with some weaker correlations noted in the Thompson-MYJ runs. It also appears that several correlations between MAEs of potential temperature and MCS precipitation ETSs at different times were prevalent under the
WSM6-MYJ runs (which were also noted with the ageostrophic and total wind MAE correlations). This is likely because the WSM6-MYJ better simulates the thermodynamic state of the PBL and thus better depicts instability, but more research into instability parameters in the WSM6-MYJ runs are necessary to validate potential temperature forecast skill correlations with MCS precipitation forecast skill.

c) Other factors influencing MCS development and sustenance

While this research focused on evaluating correlations between forecast skill of LLJ variables and MCS precipitation in the WRF, a few other parameters that could affect MCS formation and evolution were also examined.

No correlations of higher ETSs associated with lower MAEs were observed in 0-1 km shear, or 0-3 km shear (not shown). While low level shear in the lowest 1 km and (to an extent) lowest 3 km can aid in organization and longevity of MCSs (Rotunno et al. 1988; Weisman 1992; Weisman and Rotunno 2004; Jirak and Cotton 2007; French and Parker 2010), it is not a necessity for MCS maintenance. Coniglio et al. (2010) found that 0-1 km shear and 1-3 km shear environments could not discriminate between short lived and long lived MCS environments. Given Coniglio et al. (2010) and that short lived and long lived MCSs were observed and successfully simulated in the WRF in the present study despite varying low level shear environments, it may be deduced that forecast skill in low level shear does not correlate to forecast skill in MCS precipitation.

Mass convergence and horizontal MFC did not correlate with MCS forecast skill. Mass convergence and MFC develop due to the directional components of the
ageostrophic winds orienting themselves perpendicular to the MCS. In an example case, ambient mass convergence weakened ahead of the MCS along the terminus of the LLJ due to the LLJ veering with time (Figs. 3-10 and 3-11). With the LLJ winds veering more parallel to MCS motion, the weaker magnitudes of mass convergence and MFC allowed convection to weaken. Mass convergence and MFC must be maximized in the immediate vicinity and downstream of convective activity to supply the large scale rising motion needed for extending the lifespan of an MCS (an observation noted in most cases in this study).

700 hPa temperature advection may also promote an enhanced environment for MCS initiation and sustainment. Cotton et al. (1989) found that MCSs initiated in environments of maximum positive temperature advection given that large scale rising motion is present. Both the RUC and WRF 700 hPa temperature advection composites (Fig. 3-12) concur with Cotton’s assessment, with the average location of MCS initiation located within the region of maximum positive temperature advection, suggesting from quasigeostrophic theory that large scale rising motion not related to the convergence at the LLJ terminus is involved in MCS initiation and sustenance. It is interesting to note that the 700 hPa positive temperature advection magnitudes were weaker for the Thompson MYJ composites compared to the other WRF runs. No correlations between MAEs of 700 hPa temperature advection and ETSs of MCS precipitation (not shown) were found.

Deep layer shear (0-6 km) has been considered an important ingredient to storm mode and plays a significant role in MCS development in conjunction with frontogenesis. As
determined by Dial and Racey (2004), Schumann and Roebber (2010) and Snively and Gallus (2014), parallel orientation of the deep layer shear vectors to surface or low level boundaries encouraged cold pool mergers shortly after convective initiation, promoting upscale growth of convection and resultant MCS development. As such, the MAEs of 0-6 km shear, surface and 850 hPa frontogenesis were compared to the ETSs of MCS precipitation. No correlations were evident between the forecast skills of 0-6 km shear and MCS precipitation, or frontogenesis and MCS precipitation (not shown). While the 0-6 km shear existed on the synoptic level, frontogenesis occurred on the mesoscale, likely reducing forecast skill. In addition, 0-6 km shear must work in tandem with frontogenesis to foster MCS development, a factor that bulk statistics such as MAE applied to single variables cannot take into account.

Stability parameters were evaluated since thermodynamic variables likely also play a role in MCS development as suggested by the forecast skill correlations between potential temperature and MCS precipitation that were noted. MLCAPE and MLCIN composites (Figs. 3-13 and 3-14) revealed that MCSs initiated along the northern bounds of more unstable, yet capped air masses. No correlations were found between MLCAPE and MLCIN forecast skill and MCS precipitation forecast skill. The LLJ pumps a thermodynamically favorable air mass northward into the terminus, where mass convergence is present, indicating an association between thermodynamic air mass properties and dynamic lift induced by convergence via the ageostrophic wind which was suggested in Coniglio et al. (2010). It is interesting to note that the Thompson runs had lesser precipitation than WSM6 (Fig. 3-7), and yet had stronger MLCIN overall (Fig. 3-14), suggesting that model microphysics and hydrometeor evolution alone may not
determine MCS precipitation totals and that convective inhibition likely limits convective development and longevity to a degree. Still, stability and MCS precipitation associations cannot be easily quantified in simple bulk statistics.

### 3.5 Conclusions and Discussion

Despite the complex nature of the relationship between LLJs and resultant MCS precipitation, some correlations were found between forecast skill for the LLJ and that of the precipitation in the associated MCSs. The forecast skill of the geostrophic wind direction had the highest correlation of forecast skill with MCS precipitation, since the geostrophic wind sets up the background flow for the LLJ to develop upon. While it would seem intuitive that the geostrophic magnitude forecast skill would have the greatest correlation with MCSs, the directional component of geostrophic wind is more significant as it sets up the orientation of the LLJ in its developmental stages, which influences that large scale environment, leading to large scale errors that impact MCS initiation and sustenance. To a lesser degree, the forecast skill of the ageostrophic (divergent) wind direction held some significant correlation with MCS precipitation forecast skill since convergence is induced at the terminus of the LLJ by the divergent wind. Mass convergence and MFC are only present to support MCS activity once they are induced by the ageostrophic winds, a factor that bulk statistics such as MAE cannot take into account, hence the lack in correlation between the forecast skill in mass convergence and MFC with that of MCS precipitation. Banacos and Schultz (2005) noted that a relationship between mass convergence/MFC and convective initiation exists, but
multiple other factors play a role in convective initiation/sustenance. Banacos and Schultz cited capping, entrainment of dry air parcels aloft and sloping of convergence with height as possible causes for ambiguity between the convergence/convective initiation relationship, suggesting that a focus on physical processes leading up to convective initiation is more advisable than monitoring convergence magnitudes themselves. Evaluating the directional component of the geostrophic and ageostrophic winds would thus be more practical than measures of convergence.

Forecast skill of the potential temperature field in the LLJ showed some correlation to MCS activity in the early stages of both LLJ and MCS development. It is not clear why, and more research is needed to investigate this finding. In addition, several correlations between the forecast skill of the LLJ and forecast skill of MCS precipitation were found mostly with WSM6 model runs. While the reason for this is also unclear, and more investigation is necessary for a definitive explanation, it is possible that the MP scheme chosen may ultimately dictate the forecast skill relationships between several variables.

In addition, it was found that 700 hPa temperature advection, which in quasigeostrophic theory supplies large scale rising motion, likely fostered an environment favorable for initiating and sustaining deep moist convection in the 20 cases examined. Still, no correlations in forecast skill between 700 hPa positive temperature advection and MCS precipitation were found. This result is likely due to the mixed sample of cases with synoptically and inertial oscillation driven LLJs. It would be interesting to repeat this research with more cases, perhaps with type C or A LLJs alone to see if this result changes.
Deep layer shear and frontogenesis played a role in both observations and forecasts. Forecast skill of deep layer shear (and also low level shear) do not show correlations with MCS forecast skill, likely because they work in tandem with other ingredients or processes to be effective in initiating and sustaining MCSs, a factor that cannot be analyzed by the bulk statistics applied in the present study. Specifically, deep layer shear vectors must run roughly parallel to low level frontogenesis to encourage convective cold pool mergers and upscale growth to MCS structures.

No correlations were noted between forecast skill of stability and MCS precipitation. While composites show that MCS initiation favors regions of minimum convective inhibition along the northwestern edge of an MLCAPE field, no quantitative conclusions could be made, solidifying the notion that identifying forecast skill via physical processes is more appropriate than by parameters.

If possible, it would be better if the number of cases with well-developed LLJs and MCSs could be doubled or tripled to see if an increase in sample size would change the results of this study. Doing so without mixing LLJ types in this study, particularly those with synoptic influence such as those described in Uccellini and Johnson (1979), may clear up some of the ambiguity of the results for LLJ variable forecast skill correlations with MCS forecast skill. In addition, an investigation into PBL and convection hydrometeor evolution and the differences in the use of PBL and MP schemes could be conducted to see if different parameterizations within the WRF are more optimal for high resolution simulations of the LLJ and resultant MCS activity. Ideally, RUC data should be replaced with a high density of in situ vertical profiler data from across the plains,
using high powered stations like the one near Lamont, OK. Analysis of WRF output was performed at 3-hourly intervals due to computational resource limitations. Refinement to hourly analyses may reveal more insightful details on LLJ and concurrent MCS evolution.

3.6 Acknowledgements

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3.7 References


3.8 Figures

Figure 3-1. For the 24 May 2007 case at 06 UTC, a) 900 mb streamlines and wind magnitudes color filled at 10 and 15 ms$^{-1}$, b) 200 mb streamlines and wind magnitudes color filled at 15 and 30 ms$^{-1}$. c) 900 mb convergence ($10^{-5}$ s$^{-1}$) in red, and 200 mb divergence in blue ($10^{-5}$ s$^{-1}$).
Figure 3-2. Same as Fig. 3-1, but for the 09 August, 2010 case at 0600 UTC.
Figure 3-3. RUC total wind (in ms$^{-1}$) at a) 200, b) 600, c) 1000, and d) 1400 meters above ground level, respectively for the 24 May, 2007 case at 0900 UTC.
Figure 3-4. Scatterplot displaying the correlation in forecast skill for a) (0300 UTC MAE) of the geostrophic wind direction in the LLJ with forecast skill (0300-0900 UTC ETS) in MCS precipitation for the Thompson-MYJ WRF runs, b) 0300 UTC MAE of the ageostrophic wind direction and 0600-1200 UTC ETS in WSM6-MYJ WRF runs, and c) 0300 UTC MAE of potential temperature and 0000-0600 UTC ETS in WSM6-MYNN WRF runs.
Figure 3-5. 200-2000 m averaged geostrophic wind composite (ms$^{-1}$) at 2100 UTC for a) RUC analysis, b) Thompson-MYJ, c) Thompson MYNN, d) Thompson YSU, e) WSM6-MYJ, f) WSM6-MYNN, and g) WSM6-YSU. ‘X’ delineates the centroid of all MCS initiation points.
Figure 3-6. 200-2000 m averaged total winds (barbs in ms$^{-1}$), atmospheric water vapor content (line contours in g kg$^{-1}$), and mass divergence (filled contours in 10$^{-5}$ s$^{-1}$, where negative values delineate convergence) at peak LLJ times for a) RUC analysis, b) Thompson-MYJ, c) Thompson MYNN, d) Thompson YSU, e) WSM6-MYJ, f) WSM6-MYNN, and g) WSM6-YSU. ‘X’ delineates the centroid of all MCS initiation points.
Figure 3-7. Composites of 0000-1200 UTC accumulated MCS precipitation (in mm) for a) STAGEIV observations, b) Thompson-MYJ, c) Thompson MYNN, d) Thompson YSU, e) WSM6-MYJ, f) WSM6-MYNN, and g) WSM6-YSU.
Figure 3-8. Observed reflectivity of MCSs from 21 June, 2010 Case for a) 2100 UTC, b) 0000 UTC, c) 0300 UTC, d) 0600 UTC, e) 0900 UTC, f) 1200 UTC. Images courtesy www.mmm.ucar.edu/imagearchive.
Figure 3-9. Same as Fig. 3-8, but for simulated reflectivity with the Thompson MYJ parameterization.
Figure 3-10. RUC analysis of 200-2000 m (at 100 m interval) averaged fields of total wind (barbs in m s\(^{-1}\)), mass divergence (filled contours at $10^{-5}$ s\(^{-1}\), where negative values delineate convergence), and atmospheric water vapor (line contours in g kg\(^{-1}\)) for the 21 June, 2010 Case at a) 2100 UTC, b) 0000 UTC, c) 0300 UTC, d) 0600 UTC, e) 0900 UTC, f) 1200 UTC.
Figure 3-11. Same as Fig. 3-10, but for the Thompson MYJ WRF simulation.
Figure 3-12. 700 hPa temperature advection composites at the time of MCS initiation for the a) RUC analyses, b) Thompson MYJ, c) Thompson MYNN, d) Thompson YSU, e) WSM6-MYJ, f) WSM6-MYNN, and g) WSM6-YSU. Wind barbs represent 700 hPa total wind (ms$^{-1}$), and filled contours represent 700 hPa vorticity advection (10$^{-5}$ Ks$^{-1}$). The ‘X’ delineates the average position of MCS initiation.
Figure 3.13. Composites of MLCAPE (Jkg$^{-1}$) during MCS initiation for a) RUC analyses, b) Thompson MYJ, c) Thompson MYNN, d) Thompson YSU, e) WSM6-MYJ, f) WSM6-MYNN, and g) WSM6-YSU. The ‘X’ delineates the average position of MCS initiation.
Figure 3-14. Same as Fig. 3-13, but for MLCIN (J kg$^{-1}$).
### Table 3-1. LLJ characteristics for each case in the 00-hr 13 km RUC analysis representing observations.

<table>
<thead>
<tr>
<th>Date</th>
<th>LLJ Type</th>
<th>Magnitude Peak wind (ms(^{-1}))</th>
<th>Time of Peak Wind (UTC)</th>
<th>Height of Peak Wind (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>05/24/2007</td>
<td>C</td>
<td>27.5</td>
<td>0900</td>
<td>750</td>
</tr>
<tr>
<td>05/30/2007</td>
<td>C</td>
<td>22.20</td>
<td>0900</td>
<td>550</td>
</tr>
<tr>
<td>06/01/2007</td>
<td>C</td>
<td>26.92</td>
<td>0900</td>
<td>850</td>
</tr>
<tr>
<td>08/17/2009</td>
<td>C</td>
<td>25.45</td>
<td>0900</td>
<td>450</td>
</tr>
<tr>
<td>05/13/2010</td>
<td>C</td>
<td>32.25</td>
<td>0900</td>
<td>1250</td>
</tr>
<tr>
<td>05/25/2010</td>
<td>C</td>
<td>30.02</td>
<td>0300</td>
<td>1650</td>
</tr>
<tr>
<td>05/26/2010</td>
<td>C</td>
<td>19.82</td>
<td>0300</td>
<td>550</td>
</tr>
<tr>
<td>06/18/2010</td>
<td>C</td>
<td>30.95</td>
<td>0600</td>
<td>850</td>
</tr>
<tr>
<td><strong>Ave</strong></td>
<td></td>
<td><strong>26.89</strong></td>
<td></td>
<td><strong>863</strong></td>
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<td>06/07/2010</td>
<td>A</td>
<td>25.15</td>
<td>0900</td>
<td>550</td>
</tr>
<tr>
<td>06/19/2010</td>
<td>A</td>
<td>25.86</td>
<td>0600</td>
<td>450</td>
</tr>
<tr>
<td>06/20/2010</td>
<td>A</td>
<td>28.04</td>
<td>0900</td>
<td>1200</td>
</tr>
<tr>
<td>08/09/2010</td>
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<td>23.40</td>
<td>0900</td>
<td>650</td>
</tr>
<tr>
<td>06/16/2012</td>
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<td>0900</td>
<td>450</td>
</tr>
<tr>
<td>05/27/2013</td>
<td>A</td>
<td>29.82</td>
<td>0900</td>
<td>850</td>
</tr>
<tr>
<td><strong>Ave</strong></td>
<td></td>
<td><strong>25.38</strong></td>
<td></td>
<td><strong>692</strong></td>
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<tr>
<td>08/19/2009</td>
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<td>0600</td>
<td>1600</td>
</tr>
<tr>
<td>06/02/2010</td>
<td>T</td>
<td>31.31</td>
<td>0600</td>
<td>550</td>
</tr>
<tr>
<td>06/11/2010</td>
<td>T</td>
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<td>0900</td>
<td>550</td>
</tr>
<tr>
<td>06/21/2010</td>
<td>T</td>
<td>29.01</td>
<td>0600</td>
<td>1200</td>
</tr>
<tr>
<td>06/15/2012</td>
<td>T</td>
<td>25.00</td>
<td>0900</td>
<td>1500</td>
</tr>
<tr>
<td>05/28/2013</td>
<td>T</td>
<td>35.48</td>
<td>0600</td>
<td>1100</td>
</tr>
<tr>
<td><strong>Ave</strong></td>
<td></td>
<td><strong>29.77</strong></td>
<td></td>
<td><strong>1083</strong></td>
</tr>
<tr>
<td><strong>Overall Ave</strong></td>
<td></td>
<td><strong>27.30</strong></td>
<td></td>
<td><strong>878</strong></td>
</tr>
</tbody>
</table>
Table 3-2. 0000-0600 UTC Equitable Threat scores and biases for the .1 inch rainfall threshold for MCSs in the Thompson-MYJ runs. Cases are sorted in ascending order of ETS scores.

<table>
<thead>
<tr>
<th>Date</th>
<th>Equitable Threat Score</th>
<th>Bias</th>
</tr>
</thead>
<tbody>
<tr>
<td>8/18/2009</td>
<td>0.064</td>
<td>0.667</td>
</tr>
<tr>
<td>5/25/2010</td>
<td>0.139</td>
<td>0.913</td>
</tr>
<tr>
<td>6/19/2010</td>
<td>0.145</td>
<td>0.903</td>
</tr>
<tr>
<td>8/16/2009</td>
<td>0.211</td>
<td>0.921</td>
</tr>
<tr>
<td>5/31/2007</td>
<td>0.228</td>
<td>0.852</td>
</tr>
<tr>
<td>6/20/2010</td>
<td>0.237</td>
<td>0.645</td>
</tr>
<tr>
<td>6/6/2010</td>
<td>0.278</td>
<td>0.528</td>
</tr>
<tr>
<td>6/15/2012</td>
<td>0.299</td>
<td>0.878</td>
</tr>
<tr>
<td>6/1/2010</td>
<td>0.318</td>
<td>0.835</td>
</tr>
<tr>
<td>5/26/2013</td>
<td>0.348</td>
<td>1.168</td>
</tr>
<tr>
<td>6/10/2010</td>
<td>0.355</td>
<td>0.967</td>
</tr>
<tr>
<td>5/29/2007</td>
<td>0.365</td>
<td>0.810</td>
</tr>
<tr>
<td>8/8/2010</td>
<td>0.367</td>
<td>1.582</td>
</tr>
<tr>
<td>5/27/2013</td>
<td>0.393</td>
<td>0.984</td>
</tr>
<tr>
<td>6/17/2010</td>
<td>0.414</td>
<td>0.685</td>
</tr>
<tr>
<td>5/12/2010</td>
<td>0.423</td>
<td>0.981</td>
</tr>
<tr>
<td>5/23/2007</td>
<td>0.428</td>
<td>0.883</td>
</tr>
<tr>
<td>5/24/2010</td>
<td>0.437</td>
<td>0.823</td>
</tr>
<tr>
<td>6/14/2012</td>
<td>0.439</td>
<td>0.761</td>
</tr>
<tr>
<td>6/18/2010</td>
<td>0.469</td>
<td>1.262</td>
</tr>
</tbody>
</table>

Table 3-3. R² and associated p-values relating the MAE of the geostrophic wind direction (in degrees) in the LLJ to the six hourly ETS of MCS precipitation involving the Thompson WRF runs. R² values in bold have a corresponding p-value of .05 or less, with a negative slope observed in scatter plot analyses (where higher equitable threat scores were observed with lower MAEs).

<table>
<thead>
<tr>
<th>Correlation</th>
<th>Thompson MYJ</th>
<th></th>
<th>Thompson MYNN</th>
<th></th>
<th>Thompson YSU</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>R²</td>
<td>P-value</td>
<td>R²</td>
<td>P-value</td>
<td>R²</td>
<td>P-value</td>
</tr>
<tr>
<td>ETS_0_6Z_MAE_geo_dir_03Z</td>
<td>0.188</td>
<td>0.055</td>
<td>0.091</td>
<td>0.194</td>
<td>0.341</td>
<td>0.007</td>
</tr>
<tr>
<td>ETS_3_9Z_MAE_geo_dir_03Z</td>
<td>0.333</td>
<td>0.007</td>
<td>0.158</td>
<td>0.082</td>
<td>0.348</td>
<td>0.006</td>
</tr>
<tr>
<td>ETS_6_12Z_MAE_geo_dir_03Z</td>
<td>0.259</td>
<td>0.021</td>
<td>0.081</td>
<td>0.224</td>
<td>0.143</td>
<td>0.099</td>
</tr>
<tr>
<td>ETS_3_9Z_MAE_geo_dir_06Z</td>
<td>0.373</td>
<td>0.004</td>
<td>0.184</td>
<td>0.058</td>
<td>0.262</td>
<td>0.020</td>
</tr>
<tr>
<td>ETS_6_12Z_MAE_geo_dir_06Z</td>
<td>0.416</td>
<td>0.002</td>
<td>0.185</td>
<td>0.057</td>
<td>0.191</td>
<td>0.053</td>
</tr>
<tr>
<td>ETS_6_12Z_MAE_geo_dir_09Z</td>
<td>0.508</td>
<td>0.000</td>
<td>0.249</td>
<td>0.025</td>
<td>0.257</td>
<td>0.022</td>
</tr>
</tbody>
</table>
Table 3-4. Same as Table 3-3 but for the WSM6 WRF runs.

<table>
<thead>
<tr>
<th>Correlation</th>
<th>WSM6 MYJ</th>
<th>WSM6 MYNN</th>
<th>WSM6 YSU</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>R^2</td>
<td>P-value</td>
<td>R^2</td>
</tr>
<tr>
<td>ETS_0_6Z_MAE_geo_dir_03Z</td>
<td>0.317</td>
<td>0.009</td>
<td>0.300</td>
</tr>
<tr>
<td>ETS_3_9Z_MAE_geo_dir_03Z</td>
<td>0.620</td>
<td>0.000</td>
<td>0.420</td>
</tr>
<tr>
<td>ETS_6_12Z_MAE_geo_dir_03Z</td>
<td>0.421</td>
<td>0.002</td>
<td>0.264</td>
</tr>
<tr>
<td>ETS_3_9Z_MAE_geo_dir_06Z</td>
<td>0.479</td>
<td>0.001</td>
<td>0.409</td>
</tr>
<tr>
<td>ETS_6_12Z_MAE_geo_dir_06Z</td>
<td>0.291</td>
<td>0.014</td>
<td>0.215</td>
</tr>
<tr>
<td>ETS_6_12Z_MAE_geo_dir_09Z</td>
<td>0.232</td>
<td>0.031</td>
<td>0.212</td>
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</table>

Table 3-5. Same as table 3-3, but for the Thompson WRF runs involving ageostrophic wind direction.

<table>
<thead>
<tr>
<th>Correlation</th>
<th>Thompson MYJ</th>
<th>Thompson MYNN</th>
<th>Thompson YSU</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>R^2</td>
<td>P-value</td>
<td>R^2</td>
</tr>
<tr>
<td>ETS_0_6Z_MAE_ageo_dir_03Z</td>
<td>0.040</td>
<td>0.400</td>
<td>0.001</td>
</tr>
<tr>
<td>ETS_3_9Z_MAE_ageo_dir_03Z</td>
<td>0.135</td>
<td>0.110</td>
<td>0.008</td>
</tr>
<tr>
<td>ETS_6_12Z_MAE_ageo_dir_03Z</td>
<td>0.182</td>
<td>0.060</td>
<td>0.047</td>
</tr>
<tr>
<td>ETS_3_9Z_MAE_ageo_dir_06Z</td>
<td>0.081</td>
<td>0.223</td>
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</tr>
<tr>
<td>ETS_6_12Z_MAE_ageo_dir_06Z</td>
<td>0.100</td>
<td>0.174</td>
<td>0.009</td>
</tr>
<tr>
<td>ETS_6_12Z_MAE_ageo_dir_09Z</td>
<td>0.142</td>
<td>0.101</td>
<td>0.002</td>
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</table>

Table 3-6. Same as table 3-5, but for the WSM6 WRF runs involving ageostrophic wind direction.

<table>
<thead>
<tr>
<th>Correlation</th>
<th>WSM6 MYJ</th>
<th>WSM6 MYNN</th>
<th>WSM6 YSU</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>R^2</td>
<td>P-value</td>
<td>R^2</td>
</tr>
<tr>
<td>ETS_0_6Z_MAE_ageo_dir_03Z</td>
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<td>0.276</td>
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<tr>
<td>ETS_3_9Z_MAE_ageo_dir_03Z</td>
<td>0.135</td>
<td>0.110</td>
<td>0.012</td>
</tr>
<tr>
<td>ETS_6_12Z_MAE_ageo_dir_03Z</td>
<td>0.407</td>
<td>0.002</td>
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<td>ETS_3_9Z_MAE_ageo_dir_06Z</td>
<td>0.132</td>
<td>0.114</td>
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<tr>
<td>ETS_6_12Z_MAE_ageo_dir_06Z</td>
<td>0.395</td>
<td>0.003</td>
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</tr>
<tr>
<td>ETS_6_12Z_MAE_ageo_dir_09Z</td>
<td>0.285</td>
<td>0.015</td>
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</tbody>
</table>
Table 3-7. Same as table 3-3, but for the Thompson WRF runs involving potential temperature.

<table>
<thead>
<tr>
<th>Correlation</th>
<th>Thompson MYJ</th>
<th>Thompson MYNN</th>
<th>Thompson YSU</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>R2</td>
<td>P-value</td>
<td>R2</td>
</tr>
<tr>
<td>ETS_0_6Z_MAE_theta_03Z</td>
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<td>0.008</td>
<td>0.025</td>
</tr>
<tr>
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<td>0.022</td>
<td>0.001</td>
</tr>
<tr>
<td>ETS_6_12Z_MAE_theta_03Z</td>
<td>0.108</td>
<td>0.156</td>
<td>0.002</td>
</tr>
<tr>
<td>ETS_3_9Z_MAE_theta_06Z</td>
<td>0.025</td>
<td>0.508</td>
<td>0.019</td>
</tr>
<tr>
<td>ETS_6_12Z_MAE_theta_06Z</td>
<td>0.105</td>
<td>0.162</td>
<td>0.047</td>
</tr>
<tr>
<td>ETS_6_12Z_MAE_theta_09Z</td>
<td>0.018</td>
<td>0.568</td>
<td>0.023</td>
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</table>

Table 3-8. Same as table 3-7, but for the WSM6 WRF runs involving potential temperature.

<table>
<thead>
<tr>
<th>Correlation</th>
<th>WSM6 MYJ</th>
<th>WSM6 MYNN</th>
<th>WSM6 YSU</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>R2</td>
<td>P-value</td>
<td>R2</td>
</tr>
<tr>
<td>ETS_0_6Z_MAE_theta_03Z</td>
<td>0.567</td>
<td>0.000</td>
<td>0.258</td>
</tr>
<tr>
<td>ETS_3_9Z_MAE_theta_03Z</td>
<td>0.587</td>
<td>0.000</td>
<td>0.166</td>
</tr>
<tr>
<td>ETS_6_12Z_MAE_theta_03Z</td>
<td>0.360</td>
<td>0.005</td>
<td>0.086</td>
</tr>
<tr>
<td>ETS_3_9Z_MAE_theta_06Z</td>
<td>0.293</td>
<td>0.013</td>
<td>0.122</td>
</tr>
<tr>
<td>ETS_6_12Z_MAE_theta_06Z</td>
<td>0.291</td>
<td>0.014</td>
<td>0.164</td>
</tr>
<tr>
<td>ETS_6_12Z_MAE_theta_09Z</td>
<td>0.076</td>
<td>0.238</td>
<td>0.017</td>
</tr>
</tbody>
</table>
CHAPTER 4: ADDITIONAL RESULTS

4.1 Using in-situ Observations to Validate RUC Analyses

Many studies have been conducted which utilized the RUC analysis as a substitute for in-situ observations (e.g., Thompson et al. 2003, Hane et al. 2008, Schumacher and Johnson 2009, Coniglio et al. 2010, and Snively and Gallus 2014). Since a wide array of data is incorporated into the RUC, many of the previous studies as well as the current study compared the RUC with available in-situ observations.

The 915 MHz vertical profiler at the Lamont, OK central facility was the only source of high vertical resolution wind data that was available hourly for the present research. Composites of profiler data were created for each LLJ type for the total wind magnitude and direction at or near the time of peak LLJ intensity (depending on the availability of quality profiler data) for the Lamont, OK site and RUC analyses at the representative grid point for Lamont. For type C cases, reasonable agreement existed for LLJ peak wind altitude (Fig. 4-1), but the RUC had weaker winds below the LLJ peak height (around 900 m) than Lamont, with a shallower decrease in wind speeds above 900 m than what the profiler data suggested. The RUC analysis also had the winds backed more than Lamont throughout the 200-2000 m layer (LLJ layer). For type A cases (Fig. 4-2), more agreement for the wind magnitude between both data sources was noted in the 1000-1500 m layer, but strong disagreement (by as much as 7 ms\(^{-1}\)) existed below that. Winds in the RUC composite were also too backed throughout the LLJ layer compared to Lamont. Lastly, type T cases (Fig. 4-3) showed the strongest disagreement between the RUC and
Lamont, where the RUC was up to 7-8 m s\(^{-1}\) weaker than Lamont (below 1 km) with the RUC wind direction too backed in a similar fashion to type C and A composites.

These composites were generated at a single point and are not representative of the three dimensional structure of the LLJ. Several sources of error existed, including quality issues with the Lamont profiler and other high powered profilers, which caused their exclusion as data sources to be assimilated in the 13 km RUC (ruc.noaa.gov/ruc13_docs/RUC13ppt.htm). The Lamont, OK station was also located at the edge of the LLJ for some cases, where subtle (to occasionally well-defined) boundaries delineated significant differences in wind speed and direction with height. The slight displacement of these boundaries (with respect to their location to the Lamont station) and convection occasionally impacting the Lamont profiler introduced considerable difference between the profiler data and RUC output.

Numerous differences for both the high profiler data and RUC analysis output exist, with common differences apparent between both sources that require attention. The RUC analysis portrays the LLJ peak wind too high compared to the Lamont data. In addition, the RUC wind magnitudes are weaker and more backed than observations suggested. Fig. 4-4 depicts a case where the LLJ was undisturbed by convection at the Lamont, OK location and relatively high quality of vertical profiler data was present. This case best sums up the common differences observed between the Lamont profiler, the RUC analysis, and WRF forecasts in all cases. It is interesting to note that the Lamont profiler and WRF are more similar than the RUC, suggesting that differences between the WRF and RUC analyses may not entirely represent errors in the WRF forecasts for the LLJs.
4.2 Figures

Figure 4-1. Composite of type C LLJ cases for the Lamont vertical profiler data and RUC analyses at the representative grid point for Lamont, OK for a) total wind magnitude and b) total wind direction.
Figure 4-2. Same as Fig. 4-2, but for type A cases.
Figure 4-3. Same as Fig. 4-2, but for type T cases.
Figure 4-4. Vertical profiler time series (wind barbs in m s\(^{-1}\)) for a) Lamont, OK vertical profiler, b) RUC analysis and c) WRF forecast (Thompson-MYJ), initialized at 1200 UTC.
CHAPTER 5: CONCLUSIONS AND ADDITIONAL REMARKS

5.1 General Conclusions

The Great Plains LLJ is a significant contributor to the development of summer nocturnal precipitation across the central US. Being able to determine the skill of forecasting the LLJ’s characteristics and the concurrent impact on nocturnal MCS evolution in high resolution modeling was the primary goal of this research. With a dense network of in-situ observations being unavailable for this project, 13 km gridded 00-hr RUC analyses were employed as a substitute to observations for comparing against 4 km WRF forecasts. Given the impact of hydrometeor distributions on convective evolution and given the impact the evolution of the PBL has on LLJ development, all cases in the WRF were run under 2 separate MP schemes, each employing 3 different PBL schemes to conduct a sensitivity test, which would determine if a WRF MP/PBL combination ranked superior in simulating reality.

Breaking up the total wind contained within the LLJ to its geostrophic and ageostrophic magnitude and directional components revealed that forecast skill in both the geostrophic and ageostrophic directional components correlated with MCS precipitation forecast skill. The diurnal geostrophic wind set up the background flow for the nocturnal LLJ to develop, while the geostrophic wind directional component established the orientation of the LLJ, rendering prediction of the geostrophic wind before and during LLJ development to be paramount in understanding and predicting LLJ and subsequent MCS activity. The ageostrophic wind direction determines the magnitude
of convergence that occurs at the terminus of the LLJ, which provides enhanced lift to support MCS evolution. In addition, positive correlations in potential temperature forecast skill within the LLJ with MCS precipitation were evident during the early stages of LLJ evolution, likely due to the influx of positive buoyancy into the LLJ terminus, which fostered the initiation and maturity of MCSs.

While MCSs were found on average to initiate in regions of higher 700 hPa positive temperature advection, no correlation in forecast skill was found between 700 hPa temperature advection and MCSs given that LLJs under different types of synoptic setups were sampled in this study, including setups where 700 hPa positive temperature advection was not necessary to present enhanced lift for supporting nocturnal MCS activity. Mass convergence, MFC, deep layer shear, frontogenesis and stability parameter forecast skill did not positively correlate with MCS forecast skill since their roles vary greatly based on the synoptic setup present. The varying measure of influence these parameters have on MCS evolution is a factor that a forecast skill bulk statistic (such as MAE) cannot take into account.

5.2 Critical Reflections

A recommendation for future work or replication of research would be choosing one LLJ type to analyze. The jet coupling and positive vorticity advection in type C LLJs supplied additional upper level support and large scale lift for MCS initiation and sustenance that was not present for type A cases. The forecast skill for parameters that vary in the amount of influence they have on MCS evolution based on synoptic
background setups may explain why variables like 700 hPa temperature advection, deep layer shear, mass convergence and MFC forecast skill did not correlate with MCS precipitation forecast skill. There was little to no use in classifying type T LLJs. While they did not have the classic synoptic background like type C jets, synoptic influence was still involved, thus it would have been better to classify these cases as type C.

Theoretically, it would be ideal if this research was repeated with only type A cases (which would force the only mechanism for MCS development to be the LLJ), especially with a larger number of cases to solidify results.

Some uncertainties are also present regarding how appropriate the RUC analyses are for representing the observed LLJ. High powered vertical wind profiler data (via the Lamont, OK central facility) and radiosonde data (at 1200 UTC at locations suspected to be within the core of the LLJ) were compared to RUC analyses output at the respective representative grid points to determine if the RUC was appropriate for being used in this study. While some of the observations suggested the RUC to be less than optimal in nature, particularly below 400 m (Hane et al. 2008), consideration must be given to observational error associated with radiosonde and vertical profiler data. Profiler data was excluded from assimilation into the RUC analysis given the high tendencies for errors. A high powered profiler network (similar to the NOAA profiler network, but with better resolution in the lowest 1-2km) would have been ideal for this project, but the lack of such a network was the reason for employing the RUC. The RUC analysis may be portraying a weaker LLJ because of the PBL scheme implemented, where the effects of a rapidly developing LLJ were dampened. To prove this, RUC analyses at a given time for the LLJ would need to be compared to a RUC forecast (initialized earlier in time) to see
if time is needed for the LLJ to develop more intensely in the RUC. This task was attempted, but archived RUC forecasts could not be located (only the analyses are archived).

Although different MP and PBL schemes were used in this research, the focus was not so much on differences in simulation of the LLJs and MCSs as it was on the relationship between errors in forecasting LLJs and MCSs. Future work should explore differences in the structures of the LLJs and MCSs simulated by the different schemes. A cursory exploration here showed that the WSM6 WRF runs had larger correlations in forecast skill between LLJ variables and MCS precipitation than the Thompson runs. The WSM6 runs often produced too much convection, with stratiform regions of rainfall and thunderstorm cores producing too much rainfall compared to observations or the Thompson runs, an observation also made by Weisman et al. (2007). Overloading of graupel in convection, which in turn can effect precipitation amounts and the thermodynamic characteristics of the immediate surrounding environment is believed to be the culprit for excess rainfall. Cold pools may be more easily generated, and with overly more intense convection (compared to Thompson and observations) excess latent heating may occur, altering the moisture, temperature and wind fields within the LLJ region (nearest to convection) in a way that significantly impacts the ETSs and MAEs (hence the WSM6 correlations). Extensive analysis of latent heating in the RUC analyses and WRF runs is recommended for future work.

Given that PBL evolution influences LLJ development, a separate study on LLJ and MCS forecast skill is recommended with simplification to using one MP scheme
(assuming an optimal scheme was identified), perhaps while restricting LLJ cases to type A only. A PBL based sensitivity test conducted with a mixture of both local and non-local mixing schemes would make differences in LLJ structure and impacts on the LLJ/MCS forecast skill correlations more apparent. Given the aforementioned complications in this research, it was rather difficult to see if a particular PBL scheme performed better than the others. Hue et al. (2010) and Coniglio et al. (2013) noted a cool, moist bias with the MYJ scheme. In addition, Coniglio et al. determined that the MYNN scheme was optimal at simulating the PBL in the WRF. A simplified replication of this research may produce the same result (which was a hypothesis for the current research, but was not observed given the aforementioned complications mentioned throughout this chapter).
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