Summer precipitation dynamics in high resolution climate simulations

David M. Flory

Iowa State University

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Summer precipitation dynamics in high resolution climate simulations

by

David M. Flory

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

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Major: Physics

Program of Study Committee:
John Hauptman, Co-major Professor
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Iowa State University
Ames, Iowa
2003

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This is to certify that the master's thesis of

David M. Flory

has met the thesis requirements of Iowa State University

Signatures have been redacted for privacy
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ABSTRACT

This study evaluates the ability of a nested mesoscale model to simulate precipitation at the sub-catchment scale (≤ 5 km). The objective is to determine if such high resolution simulation produces appropriate summer precipitation in the central U.S. for a basin with little topographic influence. A climate version of the non-hydrostatic fifth-generation Penn State/UCAR mesoscale model (MM5) was used to downscale a global climate scenario to cloud resolving scales over a domain centered over central Kansas. Climate integrations with 5 km grid spacing were done for May, June, July, and August of 1997. Model output was compared to three precipitation datasets each originating from the National Climactic Data Center's (NCDC) cooperative observer program (co-op) network. While simulations at both 5-km and 20-km grid spacing adequately reproduce precipitation spatial patterns and event timings for our purpose, they over-estimate the amount of precipitation in the domain. The good agreement of aggregated 5-km model intensities with co-op observations suggests the resolution of the observation network is too low to properly evaluate the 5-km output. Model vertical motion, relative vorticity, and mesoscale circulation associated with accumulated precipitation patterns suggest the model is generating mesoscale features consistent with MCCs.
CHAPTER 1. INTRODUCTION

The past decade has seen a progression from coarse resolution atmospheric general circulation models (AGCMs) to high resolution regional climate models (RCMs) driven by lateral boundary conditions. With the increasing resolution, it has become clear that physical processes such as radiation, convection, and condensation are as important as the model fluid dynamics (Sellers et al., 1997). The continued increase in computational power has allowed simulations at small spatial scales. Researchers have performed simulations, discussed further below, at resolutions representative of river basin and catchment scales that are of interest to hydrologists. The objective here is to determine if such high resolution simulation produces appropriate summer precipitation in the central U.S. for a basin with little topographic influence.

Hydrologists interested in the water resource management of a particular basin require accurate predictions of runoff, evaporation, and drainage. Hay et al. (2002) analyzed runoff in the Animas River Basin located in southwestern Colorado, United States, using a distributed hydrologic model and 8-day forecasts of precipitation and maximum and minimum temperature from phase one of the National Center for Environmental Prediction (NCEP) reanalysis which has a horizontal grid spacing of approximately 210 km (Kalnay et al., 1996). While significant improvements in the NCEP-based runoff forecasts were seen over those based on climatology, the authors regarded the NCEP precipitation forecasts as poor. They hypothesized that useful forecasts most likely occurred using NCEP output because the Animas river basin is dominated by snow melt. Snow-melt basins are strongly influenced by temperature, with daily variations in precipitation being less important than the value of precipitation over the accumulation period. Consequently, the forecast quality may not hold in other river basins where surface
hydrology is predominantly influenced by rainfall. In addition, the horizontal resolution of current-generation atmospheric models is often much larger than the basin size used in operational hydrologic applications. This requires disaggregating coarse-resolution output to the hydrological model’s grid over the catchment based on knowledge of hydro-climatic variability at the sub-catchment scale provided from station data. Unfortunately, most meteorological observing networks do not have adequate station density to characterize this sub-catchment variability and hydrologic models must use empirical approaches to generate input. For better estimates of basin and sub-catchment runoff, higher resolution precipitation simulations are required.

Mass et al. (2002) studied the effects of horizontal resolution on forecast accuracy while also providing a review of high resolution studies from the past two decades. High resolution studies, generally using a grid spacings of four kilometers or less, have been done on squall lines (Weisman et al., 1988; Skamarock et al., 1994; Weisman et al., 1997), cyclic mesocyclogenesis (Adlerman and Droegemeier, 2002), extreme rain events (Nielson-Gammon and Strack, 2000), and hurricane structure (Kuo et al., 2001). High resolution simulations over complex topography include McQueen et al. (1995) [Susquehanna River valley], Doyle (1997) [Central California Coast], Colle and Mass (1998) [Cascades], and Colle and Mass (2000) [Pacific Northwest]. While the primary focus of the review was on forecasting, it does provide insight on the impact increasing resolution has on simulating convective systems, stating that more realistic mesoscale structure and evolution are found as the grid spacing decreases to single digits (km). A collective conclusion of the above studies is that high resolution appears to be most useful for strongly forced convection, or convection associated with fronts, drylines, or topography (Mass et al., 2002). Details and further analysis of these studies can be found from the review of Mass et al. (2002) or from the papers themselves.

Convective processes generally occur on scales smaller than can be resolved by current model resolutions, and, as a consequence, must be parameterized. Convective parameterizations are often the source of error in model precipitation (Grell et al., 2000b). Eliminating the use of convective parameterizations requires running at high resolution and explicitly resolving
all vertical motions causing precipitation. Yu et al. (1999) explored the simulation of precipitation and runoff of single storm events using high resolution, two-way nested runs (36km, 36+12km, 36+12+4km) of MM5 (Grell et al., 1994) over the Susquehanna River basin in the northeastern United States. Using output from their hydrologic model system (HMS) with inputs of observed precipitation and modeled precipitation, the MM5 setup using both nests (36+12+4km) provided the most appropriate setup for the hydrologic simulation. Grell et al. (2000b) used a non-hydrostatic version of MM5 to simulate precipitation over the complex terrain of the Alps using model resolutions of 15, 5, and 1 km. Higher resolution runs resulted in precipitation increases and more realistic precipitation patterns due to better resolution of the local orographic features. However, the authors expressed concern as to whether or not continued increases in model resolution would lead to a convergence in the precipitation patterns and amounts.

Mesoscale convective complexes [MCCs, Maddox (1980)] are prominent contributors to summer precipitation in the central U.S. (Maddox, 1980). MCCs are large, long-lived mesoscale convective systems (MCS) that exhibit a quasi-circular cloud shield (infrared temperature $\leq -52{}^\circ C$) in infrared satellite imagery (Table 1). MCCs most frequently appear in the Central Plains of the United States and represent a region of mid-to-upper-tropospheric upward mass flux that is convectively driven and accompanied by a widespread area of stratiform precipitation (McAnelly and Cotton, 1989; Maddox, 2001; Fritsch and Maddox, 1981). Maddox (1980) observed that much of the precipitation in the summertime Midwest is related to this class of storm. Fritsch and Maddox (1981) quantified the point further stating in a study of 74 MCC’s from 1982 and 1983 that MCC’s typically account for 20-50% of annual rainfall over the Central Plains states.

Anderson and Arritt (1998) define a PEC (Persistent Elongated Convective system) as a large, long-lived MCS that fulfills the size and duration criteria for an MCC (Maddox, 1980), but not the shape criterion. MCCs and PECs tend to develop in regions of strong southerly flow for which low-level warm advection contributes mesoscale ascent and thermodynamic support (Cotton et al., 1989; Anderson and Arritt, 1998). Comparing weeks of frequent and infrequent
Table 1.1  Mesoscale Convective Complex (MCC), based upon analyses of enhanced IR satellite imagery. From Maddox(1980)

<table>
<thead>
<tr>
<th>Category</th>
<th>Physical characteristics</th>
</tr>
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<tbody>
<tr>
<td>Size:</td>
<td>A - Cloud shield with continuously low IR temperature $\leq -32^\circ C$. Must have an area $\geq 100,000 \text{ km}^2$.</td>
</tr>
<tr>
<td></td>
<td>B - Interior cold cloud region with temperature $\leq -52^\circ C$ must have an area $\geq 50,000 \text{ km}^2$.</td>
</tr>
<tr>
<td>Initiate:</td>
<td>Size definitions A and B are first satisfied.</td>
</tr>
<tr>
<td>Duration:</td>
<td>Size definitions A and B must be met for a period $\geq 6$ hours.</td>
</tr>
<tr>
<td>Maximum extent:</td>
<td>Contiguous cold cloud shield (IR temperature $\leq -32^\circ C$) reaches maximum size.</td>
</tr>
<tr>
<td>Shape:</td>
<td>Eccentricity (minor axis/major axis) $\geq 0.7$ at time of maximum extent.</td>
</tr>
<tr>
<td>Terminate:</td>
<td>Size definitions A and B no longer satisfied.</td>
</tr>
</tbody>
</table>

MCC occurrences in 1986 and 1987, Augustine and Howard (1991) found the active weeks were dominated by a deep tropospheric ridge centered over the south-eastern U.S. with a low-level moisture flux into the Central Plains region. Similarly, the combination of a synoptic-scale low-level trough in the large scale flow, the low-level jet (LLJ), and frontogenesis along frontal zones are thought to be key ingredients in MCC development.

A well known fallibility of numerical climate models is their propensity to over-estimate low-intensity and under-estimate high-intensity precipitation events (Mearns et al., 1995; Chen et al., 1996; Giorgi and Marinucci, 1996). Through analysis of correlations of observed precipitation at two points versus their separation distance for 6-h accumulation periods, Gutowski et al. (2003) suggest a model needs to resolve spatial scales of approximately 50 km and less to adequately replicate precipitation intensity distributions. This implies a grid spacing of smaller than about 15 km and that increasing model resolution could lead to better simulation of high-intensity precipitation events. The correlations also demonstrate low-intensity events are associated with larger spatial scales, and decreased grid spacings are expected to have little or no impact on these events. The authors also note that results of the Giorgi and Marinucci
(1996) study show, in some cases, decreasing resolution gives a detrimental increase in the frequency of low-intensity precipitation events.

The focus of the above studies has been primarily on high resolution climate runs over complex topography or case studies of convection for a single day or several days. In contrast, this study will address the ability of a nested mesoscale model to simulate spatial and temporal variations in hydroclimate for a basin with little topographic influence and substantial convective precipitation.

Mass et al. (2002) contend that while subjective comparison of observed and forecast structure suggest the value of high resolution, objective evaluation can result in a somewhat different conclusion, particularly if there are timing or positional errors of the mesoscale features. Consequently, analysis of temporal and spatial precipitation patterns will take place from a climatological perspective. Cumulative and monthly precipitation maps, domain average accumulations, precipitation time series, and precipitation intensity and vertical motion statistics will be the basis for examining how high resolution simulation may yield improvements of known deficiencies of a lower-resolution counterpart. In addition, analysis of mesoscale circulations associated with precipitation events will be used as a first look at whether models resolve the circulations associated with MCCs and MCSs.

This study consists of 6 chapters. Chapter 2 will describe the region of interest and integration time period, and Chapter 3 the model setup. Chapter 4 will present the analysis of model precipitation, Chapter 5 will look at the general dynamics, and, finally, Chapter 6 provides a summary of the work and suggestions for future research.
CHAPTER 2. REGION/PERIOD OF INTEREST

The Walnut River watershed is east of Wichita in southeast Kansas (Figure 2.1). The area measures 100 km from north to south and 60 km from east to west. The western half is cropland with encroaching urbanization, while the eastern half is grassland at the edge of the Flint Hills. The area experiences a strong east-west gradient of annual precipitation receiving 76 cm on the west side and 86 cm on the east. Annual snowfall in the region is about 35 cm, or 4-5% of the liquid water content of the annual precipitation (LeMone et al., 2000). The watershed lies within the Arkansas-Red river basin, contains extensive data collection networks for meteorological and hydrological data, and has been the focus of numerous intensive observing and simulation projects [ARM/CART (Stokes and Schwartz, 1994), GCIP (Coughlan and Avissar, 1996), PILPS (Henderson-Sellers et al., 1993), CASES (LeMone et al., 2000)].

The ultimate goal of this study is to evaluate the ability of high resolution models to simulate hydroclimate in a region with little or no topographic influence, ideally providing better precipitation forecasts for hydrological models. The impact of orography in the Arkansas/Red River basin is small, and precipitation from convective systems produces a significant fraction of the total. As stated in Chapter 1, June-August precipitation in the Central Plains is dominated by MCCs (Fritsch and Maddox, 1981). Anderson and Arritt (1998, 2001) demonstrate a high frequency of MCC and PEC occurrences in June, July, and August for the years 1992, 1993, 1997, and 1998. As discussed in chapter 3, boundary conditions for our high resolution domain were only available for 1994-1999, limiting our selection of years. 1997 and 1998 were El-Niño years (Ropelewski and Halpert, 1996), so increased rainfall should have been expected in our region of interest. Anderson and Arritt (2001) found a positive precipitation
anomaly in 1997 of over 5 cm for most of south-western Kansas and greater than 15 cm in far
south-west and south-central portions of the state. The authors also indicate that while the
number of MCC events was near the climatological average (33) for both 1997 (32) and 1998
(29), significantly more convective systems (MCCs and PECs) initiated and tracked through
the Red-Arkansas river basin in 1997 (Figure 2.2).

With these factors in mind, a domain encompassing the Arkansas/Red River basin was
selected for this study and simulations were done for the warm summer months (May, June,
July, and August) of 1997.
Figure 2.1 Arkansas/Red River Basin and Walnut River watershed.
Figure 2.2 Precipitation anomaly (cm) and tracks of -52° cloud shield centroid based on the position at initiation, maximum, and termination for each MCC (Mesoscale Convective Complex) and PEC (Persistant Elongated Convection). Adopted from Anderson and Arritt (2001).
CHAPTER 3. MODEL SETUP

The limited area model used for this study is a version of MM5 (Grell et al., 1994) expanded for climate and chemistry applications [MM5/MCCM, Grell et al. (2000a)]. This non-hydrostatic model uses the fully compressible mass continuity equation, neglecting the diabatic heating term that contributes to the pressure tendency. For its vertical coordinate, the model uses the terrain following σ coordinate defined as

\[ \sigma = \frac{p - p_t}{p_s - p_t} \]  

(3.1)

where p is pressure, \( p_t \) is the specified, constant pressure at the top of the model, and \( p_s \) is the surface pressure. The prognostic three-dimensional variables used in the model are temperature, water vapor, horizontal and vertical winds, along with surface pressure.

Parameterizations employed for our study include the Grell convective parameterization scheme (Grell, 1993), the rapid radiative transfer model (RRTM) of Mlawer et al. (1997), the Burk and Thompson boundary-layer turbulence parameterization (Burk and Thompson, 1989), and the soil/vegetation/snow parameterization of Smirnova et al. (2000). This implementation of turbulence parameterization requires kinetic-energy as an additional model prognostic variable (Grell et al., 2000a). The land-surface parameterization is directly coupled to the turbulence parameterization scheme and is not included in the standard release of the MM5 model. The scheme incorporates an energy-conserving solution for the fluxes of heat and moisture at the soil surface. A one-dimensional equation of diffusive and gravitational motions is used for soil moisture transfer and a one-dimensional diffusion equation is applied for heat conduction. Transfer of moisture from the soil takes place through evaporation from the canopy and transpiration as well as evaporation from bare soil. Additional details on this land-surface parameterization appear in Smirnova et al. (2000) and further MM5/MCCM model details
in Grell et al. (2000a) and Grell et al. (2000b). A summary of model parameterizations and computational parameters used in this study appear in Table 3.1.

**Table 3.1  Model parameterizations and setup.**

<table>
<thead>
<tr>
<th>Parameterized Quantity</th>
<th>Scheme</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cumulus:</td>
<td>Grell - Ensemble</td>
</tr>
<tr>
<td>Turbulence (PBL):</td>
<td>Burk-Thompson</td>
</tr>
<tr>
<td>Explicit Moisture:</td>
<td>Dudhia Simple Ice</td>
</tr>
<tr>
<td>Radiation:</td>
<td>RRTM (Rapid Radiative Transfer Model)</td>
</tr>
<tr>
<td>Land Surface:</td>
<td>Smirnova</td>
</tr>
<tr>
<td>Time Stepping:</td>
<td>Leap-Frog (long) and forward-in-time (short)</td>
</tr>
<tr>
<td>Lateral Boundary:</td>
<td>Updated at 6 hour intervals</td>
</tr>
<tr>
<td>Model Time step:</td>
<td>10 seconds (5-km)</td>
</tr>
</tbody>
</table>

The 5-km resolution domain (D02, Figure 3.1) was chosen to include the Walnut River watershed and to be far enough west to allow the highest probability of initiation and tracking of MCSs within the domain without placing the domain edge on the complex topography of the Rocky Mountains. The domain uses 121 north-south and 165 east-west grid points. Vertical discretization varies from the surface to the top of the atmosphere (100 mb), using 26 full-sigma levels (25 half-sigma) with a higher density of levels at the surface.

Dr. G. Grell and colleagues at the Forecast System Laboratory (FSL) in Boulder, Colorado have performed a six year (1994-1999) integration on a 20-km resolution domain (D01, Figure 3.1) deriving lateral boundary conditions from the NCEP/NCAR reanalysis (Kalnay et al., 1996). The output from this integration provides lateral boundary conditions for the 5-km domain. To insure the comparisons of model output were only impacted by the change in model resolution, the physical parameterizations used for the 5-km simulations were the same as those in the 20-km simulations. The lateral boundary conditions are ingested by nudging the outermost 4-5 grid points (relaxation zone) of the nested domain. The nudging assigns the outermost row or column of the model grid to the value specified from the boundary conditions. For the next four rows and columns from the boundary, the model is nudged toward the boundary conditions. The strength of this nudging decreases linearly away from the boundary. Plots of model grid points appear in Figure 3.2 and Figure 3.3, while the resolution, dimen-
Figure 3.1 Model domains. D01 is coarse resolution domain (20 km grid spacing) simulated by Dr. G. Grell and colleagues at the Forecast System Laboratory (FSL) in Boulder, Colorado, and D02 the fine resolution domain (5 km grid spacing).
sions, and number of grid points for each model domain is shown in Table 3. For readability, Figure 3.2 and Figure 3.3 show only every 5th model grid point.

Table 3.2 Model domain characteristics.

<table>
<thead>
<tr>
<th>Domain</th>
<th>Resolution</th>
<th>Domain size</th>
<th>Points in domain of interest</th>
<th>Points in Analysis Domain</th>
</tr>
</thead>
<tbody>
<tr>
<td>D01</td>
<td>20-km</td>
<td>120x150</td>
<td>18000</td>
<td>755</td>
</tr>
<tr>
<td>D02</td>
<td>5-km</td>
<td>121x165</td>
<td>19965</td>
<td>12150 (90x135)</td>
</tr>
<tr>
<td>D02-Agg</td>
<td>~20-km</td>
<td>22x33</td>
<td>726</td>
<td>726</td>
</tr>
</tbody>
</table>

To overcome difficulties associated with interpolation of soil moistures and temperatures from the 20-km coarse grid to the 5-km fine grid, and to insure realistic initialization of the values for the analysis period, integration of the 5-km domain began in October 1996. Analysis of the model output will focus on May, June, July, and August of 1997.

Due to the high computational requirements of running a large 5-km grid, the distributed memory implementation of the model was used. Model runs were performed on a 350 node 2.4 Ghz Xeon Linux cluster at Argonne National Laboratory under the generous sponsorship of Dr. John Taylor. Simulations used 32, 64, and 128 processors, though all analyzed runs utilized 128 processors and required between seven and eight hours per simulated month.
The 5-km domain

Figure 3.2  Model grid points for 5-km domain (D02). For readability, every 5th grid point is shown.
The 20-km domain

Figure 3.3 Model grid points for 20-km domain (D01). For readability, every 5th grid point is shown.
CHAPTER 4. PRECIPITATION ANALYSIS

4.1 Observations

The comparison of model precipitation to observations in this study utilized three different datasets. Each used observations from stations participating in National Climactic Data Center's (NCDC) cooperative observer program (co-op).

Dr. Martyn Clark (University of Colorado) provided a filtered daily precipitation dataset created from the NCDC's daily precipitation summary (Eischeid et al., 2000). The filtered dataset contains flags indicating missing, incomplete, or otherwise questionable precipitation reports. For this study, only flag-free station reports and stations that reported daily for over 90% of the time period of interest were used. While this technique likely resulted in the removal of quality data from the dataset, it provided a high quality dataset based on stations that consistently provided reliable reports. The 5-km model domain contained co-op stations from Kansas (233), Nebraska (41), Oklahoma (41), Missouri (11), Colorado (4) and Texas (1). In total, 331 stations were used from this observational dataset. For the rest of this study, this dataset will be referred to as the daily co-op observational dataset.

The Climate Prediction Center (CPC) manages a daily precipitation dataset on a 0.25° x 0.25° grid for the continental United States (Higgins et al., 2000). This dataset uses observations from the NCDC daily co-op network, CPC datasets (River Forecast Centers data and 1st order stations), and daily accumulations from the NCDC hourly co-op dataset. The observations were gridded using a Barnes (1964) objective analysis scheme with a varying radius of influence (~300 km). As we will see later, application of the objective analysis scheme resulted in a smoothed dataset when compared to the daily co-op observational dataset. This dataset will be referred to as the CPC observational dataset.
Hourly precipitation data were used to evaluate the diurnal cycle of simulated precipitation. Precipitation data from the co-op hourly dataset (Hammer and Steurer, 2000) was initially parsed to include only co-op stations that fell within the analysis domain and were reporting observations during the time period of interest (May-August, 1997). These stations were then compared to valid stations in the filtered daily co-op observation dataset mentioned above. Stations in the hourly list that were not in the daily station list were removed from the hourly dataset. As with the daily co-op dataset, we suspect that this method led to the exclusion of quality data, but ultimately resulted in a high quality data set based on stations that consistently provided reliable reports. Of the 618 stations active in the model domain during the time period, 312 qualified as participating stations; 62 of the 312 reported measurable hourly precipitation during the period resulting in 4708 reports. It was assumed that the remaining 250 stations not reporting hourly measurable precipitation either did not receive precipitation during the period, though this is highly unlikely, or did not report hourly observations. This dataset will be referred to as the hourly co-op dataset.

A summary of the observational datasets is shown in Table 4.1.

<table>
<thead>
<tr>
<th>Dataset</th>
<th>Period</th>
<th>Qualifying Stations</th>
<th>Valid Stations</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>COOP</td>
<td>Daily</td>
<td>362</td>
<td>331</td>
<td>(Eischeid et al., 2000)</td>
</tr>
<tr>
<td>CPC</td>
<td>Daily</td>
<td>529</td>
<td>529</td>
<td>(Higgins et al., 2000)</td>
</tr>
<tr>
<td>COOP</td>
<td>Hourly</td>
<td>618</td>
<td>312</td>
<td>(Hammer and Steurer, 2000)</td>
</tr>
</tbody>
</table>

4.2 Model Output and Post-processing

Spatially averaged daily and accumulated precipitation for May-August 1997 were used to compare results of model runs using different numbers of processors (32, 64, and 128), and, consequently, different domain decompositions. An algorithm within the model determines domain decomposition at run initiation using the number of processors assigned by the user for use in the north-south and east-west directions. Comparison of daily and May-August total accumulated precipitation (not shown) reveal no difference between model runs using different
domain decompositions, implying that further analysis could proceed without concern for the number of processors used. The remainder of this study uses model output from the 128 processor run.

Domain averages for the 5-km integrations were calculated by averaging the accumulations from all grid points within the domain except those within 15 grid points from the boundary. This insured our analysis would be sufficiently removed from the nudging effects of the boundaries. The final analysis domain had 95 north-south grid points and 135 east-west grid points. References to the 5-km domain from this point forward refer to this reduced domain. Spatial averages for the 20-km domain were calculated by selecting only the 20-km grid points that fell within the 5-km domain of interest. A summary of model output domains appears in Table 3.

To compare the 5-km spatial precipitation fields with the 20-km fields and CPC observations, the 5-km data was passed through a 9-point smoothing technique similar to that described in Haltiner and Williams (1980). The response function for a 9-point smoother in two-dimensions can be represented as

\[ R(S, \lambda) = (1 - 2S \sin^2 \frac{\Delta x}{\lambda})(1 - 2S \sin^2 \frac{\Delta y}{\lambda}) \] (4.1)

where \( S \) is a determined constant, \( \lambda \) is the wavelength of interest, and \( \Delta x \) and \( \Delta y \) are the grid spacings in the x and y directions, respectively. Choosing \( S = 1/2 \), results in \( R = 0 \) or the removal of \( 2\Delta x \) and \( 2\Delta y \) (2 grid-point) waves from the field and reduces the amplitude of slightly longer waves. Since wavelengths less than \( 2\Delta x \) (\( 2\Delta y \)) cannot be resolved on a grid with spacing \( \Delta x \) (\( \Delta y \)), and to prevent these wavelengths from showing up in longer wavelengths through aliasing, \( S = 1/2 \) were used in this study. Each successive pass through the smoother results in an additional application of the response function. The 5-km data was passed through the smoother four times. The first resolvable wavelengths are the \( 3\Delta \) wavelengths or 60 km wavelengths in the 20-km model. These are \( 12\Delta \) waves in the 5-km model. The four successive passes through the smoother result in a total response function of \( R = 0.57 \) for these waves and \( R = 0.87 \) for wavelengths twice this length (\( 24\Delta \) waves in the 5-km model).

Intensity histograms were created to evaluate the statistics of daily precipitation events. All bin widths easily satisfied the minimum width criteria suggested by Wilks (1995) for avoiding
excessively fine and potentially noisy gradations in precipitation intensity. Comparison of daily CPC precipitation intensities with intensities from the daily co-op observations reveal that the objective analysis smooths the CPC precipitation field. Histograms of this dataset have a reduced frequency of intense precipitation events (> 4 cm/day, Figure 4.1), an excess in the frequency of light to moderate events (< 4 cm/day). Both of these behaviors can be attributed to the large, varying radius of influence (~300 km) used in the objective analysis.

By assuming the co-op stations are equally spaced throughout the domain, an average distance between stations can be estimated. The daily co-op dataset contains 331 stations (See Table 4.1), or an estimated average grid distance of ~30 km. This assumption is slightly flawed as a higher density of co-op stations can be found in the eastern portion of the domain (Figure 4.2), yet it is adequate for our purpose. By aggregating points in the 5-km domain, we can create an effective grid spacing comparable to that of the 20-km and daily observation dataset. By a method similar to that of the 9 point smoother mentioned earlier, aggregation of the 5-km fields is used to remove small scale variations unresolvable by the 20-km domain and the observational dataset creating a new data field with comparable resolution. Aggregation of the 5-km was accomplished by averaging 4x4 blocks of grid points, resulting in a effective model grid spacing of 20 km (810 grid points versus the original 12,960).

4.3 Results

4.3.1 Spatial and temporal averages

Spatial averages of model precipitation amounts were calculated to quantify better the inter-model differences. The domain average of the 5-km accumulation (Figure 4.3) is in excess of the co-op stations by approx. 80 mm (20 %) by the end of the time period. By comparison, the domain average of the 20-km simulation accumulation has better agreement with the observations with an excess of around 20 mm (5 %). It is interesting to note that accumulations for the daily co-op stations are around 10 mm less than the smoothed CPC dataset. This deficit can be explained through the larger frequency of light intensity events when compared to the co-op stations. Even though the daily co-op dataset has a higher
Figure 4.1 Daily precipitation histogram for daily CPC and daily co-op data. Bin widths set at 0.25 cm/day. Ordinate gives the ratio of bin counts to total number of counts contributing to histogram. CPC refers to observational data from the CPC unified U.S. precipitation dataset.
Figure 4.2 National Climactic Data Center (NCDC) co-op station locations.
frequency of intense events, the light to moderate events make up a much greater fraction of the overall events. A secondary contributing factor may be the quality control imposed on the daily co-op data. As previously mentioned, our control method may have resulted in the exclusion of good data. This could lead to either an over-estimation or under-estimation of domain precipitation depending on the nature of the discarded events.

A breakdown of the domain averaged accumulated precipitation into convective and explicit fractions is shown in Figure 4.4 and Figure 4.5 for the 5-km and 20-km domain, respectively. In this context, parameterized precipitation is precipitation produced by the convective parameterization scheme; explicit precipitation refers to precipitation produced by the explicitly resolved moisture dynamics, or that which results from saturation of moist parcels. Both figures show the primary component of the precipitation is explicit (resolved) in nature. Seventy-two percent (340 of the 440 mm total) of the precipitation is resolved in the 5-km simulations compared to approximately sixty-nine percent (285 of 410 mm) in the 20-km simulation. At higher resolutions, more of the precipitation-producing vertical motions will be explicitly resolved, and less will need to be parameterized by the cumulus parameterization scheme. However, even at 5-km, a cumulus parameterization scheme is needed to represent cloud-scale vertical motions or vertical motions with a horizontal area less than 25 km². As resolutions increase beyond 5-km, the fraction of parameterized precipitation will continue to drop until a cumulus parameterization scheme is no longer needed.

Average hourly precipitation is shown in Figure 4.6 where the observation diurnal cycle was produced from the hourly co-op dataset. The average diurnal cycle was calculated by accumulating precipitation amounts in hourly bins and dividing by the total number of observations including non-precipitation events. Both models reproduce the nocturnal summertime precipitation pattern observed in the U.S. Midwest (Wallace and Hobbs, 1977). The 5-km average hourly precipitation contains a distinct diurnal cycle with a precipitation maximum occurring around 01Z (7 pm local standard time). In contrast, the diurnal cycle of the 20-km integration has a broader peak, smaller amplitude, and better matches the observed diurnal cycle. It is clear that the amplitude of the diurnal cycle for the 5-km simulations is too large. This
Figure 4.3 Domain averaged accumulated precipitation (mm) for May-August 1997.
Figure 4.4 Convective/Non-convective breakdown of 5-km domain average accumulated precipitation (mm) for May-August 1997.
Figure 4.5 Convective/Non-convective breakdown of 20-km domain average accumulated precipitation (mm) for May-August 1997.
over-estimation results in the excess seen in the domain average accumulations (Figure 4.3). The amplitude of the 20-km simulation is also slightly larger than observed, yet clearly outperforms the 5-km run in total accumulated precipitation.

Four month accumulations are shown for the high-resolution (5-km), low-resolution (20-km), and the CPC dataset in figures 4.7, 4.8, and 4.9, respectively. An observation based climatology from 1948-1998 is shown in Figure 4.10. The climatology shows a west to east gradient in precipitation during May-August which can be explained in part through the increased influence of the low-level jet (LLJ) on the eastern portion of the domain. The observations show 1997 strays from the climatological average during the period producing over 600 mm of accumulated precipitation over a large area in south-central Kansas. This is consistent with the location of the precipitation anomaly listed in Anderson and Arritt (2001) (Figure 2.2). Both simulations do an adequate job representing the overall precipitation pattern with higher accumulations in south-central and south-eastern portions of the state. Both simulations tend to reproduce the U-shaped precipitation region in south-eastern Kansas also found in the observations, although the model accumulations are excessive, producing areas over 800 mm. Both simulations reproduce the dry (low precipitation) region in central Kansas. This dry hole is also found in the observations, but is shifted to the north. This may be a result of the large radius of influence used to build the gridded dataset. 5-km precipitation is excessive in the pan-handle of Oklahoma and in parts of both northwest and southwest Missouri. The latter problem also occurs with the 20-km simulations. The precipitation maximum near the Walnut Creek watershed in south central Kansas also seems to be excessive, with over 700 mm in the four-month period. While an observation maximum does exist in the region, the total accumulation does not exceed 700 mm.

A monthly breakdown of the total accumulation is shown in Figures 4.11, 4.12, and 4.13, and a monthly climatology (1948-1998) appears in Figure 4.14. Both models do a good job reproducing monthly climatology of the region, with May and June being the wettest months of the four. For May, the simulations shift the precipitation maximum to the south-west. The 5-km simulation has the largest shift, moving the observed east-central Kansas maximum to the
Figure 4.6  Average hourly precipitation rate for May-August 1997.
south-central portion of the state. Precipitation magnitudes across the domain are respectable, although the 5-km simulation extends the region of 100-150 mm accumulation throughout the domain. Both simulations do a good job of reproducing the overall precipitation pattern in June, although both tend to exaggerate the maximum. The 20-km simulation exhibits a precipitation maximum in the northwestern portion of Kansas, a behavior hinted at by the 5-km simulation, and only slightly suggested by the observations. July is climatologically the driest month of the four studied. Both simulations do a good job of representing the spatial pattern and magnitude of the precipitation during this month. The observation maxima in south-central and eastern Kansas are both present in the model fields, and the model precipitation maximum (both 5-km and 20-km) in northwestern Kansas is hinted at by the observational field. Both models underestimate the domain wide average precipitation in August. It is unclear if this is a consequence of the observational smoothing of the precipitation over the domain, or if the models tend to be too dry. While both models produce a maximum in the western portion of the domain, the 20-km model does a better job with the magnitude. Both models generated a maximum near northeastern Kansas which is not seen in the observational dataset.

Plots of the domain average daily time series for the 5-km and 20-km integrations are shown in figures 4.15 and 4.16, respectively. The models tend to over-produce daily amounts while adequately reproducing the daily time series of the observations. Both models heavily over-predict the first event of May. The 5-km simulation appears to do better job with the event around day 128 when the 20-km almost doubles the daily accumulation, although the fine grid run does spread the event over several days not realized in the observations. The models perform well for the event around day 139, although both are a bit over-active in the days leading up to the event. Several events are essentially missed (e.g., days 195-205), or drastically under-predicted by the models (e.g., days 208-217) in the second half of the integration. Both models heavily over-predict the day 221-228 event, although seem to do a respectable job on the timing of the disturbance. A comparison of model time series is provided in figure 4.17.
Figure 4.7 Smoothed 5-km model accumulated precipitation (mm) for May-August 1997.
Figure 4.8 20-km model accumulated precipitation (mm) for May-August 1997.
Figure 4.9  CPC accumulated precipitation (mm) for May-August 1997. CPC refers to observational data from the CPC unified U.S. precipitation dataset.
Figure 4.10  CPC precipitation (mm) climatology for May-August. Climatology taken as average from 1948-1998. CPC refers to observational data from the CPC unified U.S. precipitation dataset.
Figure 4.11 Smoothed 5-km model monthly accumulations (mm).
Figure 4.12 20-km model monthly accumulations (mm).
Figure 4.13  CPC monthly accumulations. CPC refers to observational data from the CPC unified U.S. precipitation dataset.
Figure 4.14 CPC precipitation (mm) monthly climatology for May-August. Climatology taken as average from 1948-1998. CPC refers to observational data from the CPC unified U.S. precipitation dataset.
Daily Precipitation Time Series
May–August 1997

Figure 4.15 Domain averaged precipitation (mm) time series for 5-km run and CPC unified U.S. precipitation dataset.
Daily Precipitation Time Series
May–August 1997

Figure 4.16 Domain averaged precipitation (mm) time series for 20-km run and CPC unified U.S. precipitation dataset.
Figure 4.17  Domain averaged precipitation (mm) time series comparison for 5-km and 20-km model runs.
4.3.2 Intensity histograms

The histogram in Figure 4.18 compares the 5-km intensities to the daily co-op data. The 5-km integration produces several extreme precipitation (> 15 cm/d) events which are not present in the observations. While the observations do contain extreme precipitation rates, for daily precipitation rates greater than 5 cm/d, the frequency of modeled rates generally exceeds that of the observations.

The aggregation of the 5-km model data result in good agreement with the co-op observations for both high and low intensity events (Figure 4.19). This suggests the spatial resolution of the co-op network may not be high enough to resolve very high intensity precipitation events. Comparison of the 20-km run's intensity histograms with observations shows that the 20-km model is also over-estimating the frequency of extreme high precipitation events (> 5 cm/d) while under-estimating the frequency of light to moderate intensity events (< 5 cm/d, Figure 4.20). In addition, the 20-km histogram includes extreme precipitation intensities in excess of 30 cm/d. Thus, while the 20-km run did a satisfactory job in reproducing the domain average accumulated precipitation (Figure 4.3), it appears it did so through a propitious balance of several extreme precipitation events with an underestimate in the frequency of light to moderate events.
Figure 4.18 Daily precipitation histogram for 5km domain and daily co-op data. Bin widths set at 0.25 cm/day. Ordinate gives the ratio of bin counts to total number of counts contributing to histogram.
Figure 4.19  Daily precipitation histogram for 5-km domain aggregated to a 20-km grid and daily co-op data. Bin widths set at 0.25 cm/day. Ordinate gives the ratio of bin counts to total number of counts contributing to histogram.
Figure 4.20 Daily precipitation histogram for 20km domain and daily co-op data. Bin widths set at 0.25 cm/day. Ordinate gives the ratio of bin counts to total number of counts contributing to histogram.
CHAPTER 5. MODEL DYNAMICS

5.1 Vertical motion

In an attempt to explain the occurrence of extreme precipitation events seen in the daily precipitation histograms, histograms of model vertical motion for the 5-km and 20-km simulations were used. Figures 5.1, 5.2, and 5.3 show the vertical motion histograms for the $\sigma = 0.3$ (325mb), $\sigma = 0.35$ (375mb), and $\sigma = 0.4$ (425mb) levels, respectively. These levels were chosen according to the composite analysis of MCC events provided by Cotton et al. (1989). In their composite of 90 organized MCC cases, the vertical profile of pressure coordinate vertical motion, $\omega$ (Figure 13 in their paper), shows a minimum in the mature and dissipation stages of development around 300-400 mb. Thus, if the model vertical motions are a result of MCC type behavior, the largest values are expected around 300-400 mb.

The histograms show extremely large vertical motions occurring in both models. Although their frequency of occurrence is small, vertical motions of between 20-30 m/s occur in the 5-km model, and between 10-20 m/s in the 20-km model. While the magnitudes of extreme vertical motion in the 20-km model is less than the 5-km, the 20-km grid cells represent an area 16 times that of the 5-km cells, and vertical velocities of 10-20 m/s over such a large area are just as extreme as 20-30 m/s over the smaller 5-km area. Vertical motions greater than 5 m/s are indicators of significant storm scale updrafts. While the model values of vertical velocity are extreme, they do not exceed the theoretical maximum vertical velocity,

$$w_{\text{theor,max.}} = 2\sqrt{\text{CAPE}},$$

(5.1)

where CAPE is the convective available potential energy, or the maximum kinetic energy a statically unstable parcel can acquire neglecting the effects of water vapor and condensed water
on the buoyancy (Holton, 1992). CAPE values in excess of 1000 J/kg are often associated with severe thunderstorms.

The 20-km vertical motion histograms have limited sampling available with vertical motion only available at six hour intervals within the model output. Initially, an increase in the number of extreme vertical motions was expected if the output were available hourly as in the 5-km simulations. Although further inspection of the 20-km diurnal cycle (Figure 4.6) suggests that the period in which peak vertical motions are occurring (times of peak rainfall rate) is sufficiently wide to achieve a representative sample, even at six hour intervals.

To better understand the nature of these extreme vertical motions, composite vertical profiles of extreme vertical motion are shown in Figure 5.4. Profiles were broken down into intervals of $\Delta \sigma = 0.2$. Only profiles that had a vertical motion greater than 10 m/s within its sigma interval were included in that interval. The number of profiles contributing to each sigma interval is listed in the legend. As we are primarily interested in the extreme values, the profiles only include vertical motions from 22Z-04Z or when the highest average hourly precipitation rates were observed according to Figure 4.6. Vertical motion profiles for all hours other than 22Z-04Z (not shown) reveal that the majority of extreme motions are occurring during this time period. The number of contributing profiles was roughly a factor of three less than those contributing to the 22Z-04Z profiles. The number of counts in each profile (see legend) suggest that extreme vertical motions are occurring most frequently at the upper levels. When extreme vertical motions are found in the lower levels, they also exist in the mid-to-upper levels, suggesting a deeply convective case.

The profiles are consistent with Cotton et al. (1989) composite MCC vertical motion profiles. The composites show negative values of $\omega$ (vertical motion) through the depth of the atmosphere for all stages, except the mature stage. While the mature and dissipation stage both contain a minimum in the $\omega$ vertical motion field around 300-400 mb, the mature stage also contains positive $\omega$ (downward motion) at lower levels centered around 850mb. This downward motion at lower levels is presumed to be associated with evaporatively cooled air from MCC generated precipitation. The simulation of the downward motion is not evident in the
5-km profiles, although visual inspection of the raw data show that weak downward vertical motions (-0.5 m/s) are present at the lower levels.

Figures 5.5-5.9 show a time series of 300 mb vertical velocity contours overlaid on filled 1-hr precipitation contours for 18-22 LST, 20 June 20, 1997 (00Z-04Z, 21 June). Contours of vertical velocity and precipitation are 2 m/s and 5 mm, respectively. Figure 5.10 shows a magnified image of the 20 LST (02Z) image focused on an intense precipitation region. It can be seen that model precipitation is associated with the strong vertical motions as suggested by the vertical motion profiles.

The dynamics leading to these extreme vertical motions are unclear at this time. Analysis of the overall and component breakdown of the vertical accelerations associated with these motions would lead to a better understanding of the principal processes and potentially point to model deficiencies. This analysis is outside the scope of this thesis.
Figure 5.1 Vertical motion histograms for $\sigma = 0.30$ ($\sim 325$mb). Bin widths for each histogram are 0.5 m/s.
Figure 5.2 Vertical motion histograms for $\sigma = 0.35$ ($\sim 375\text{mb}$). Bin widths for each histogram are set at 0.5 m/s. Ordinate reflects ratio of bin counts to total number of counts contributing to histogram.
Figure 5.3  Vertical motion histograms for $\sigma = 0.40$ ($\sim 425$mb). Bin widths for each histogram are set at 0.5 m/s. Ordinate reflects ratio of bin counts to total number of counts contributing to histogram.
Figure 5.4  Average profile of vertical motion from 22Z-04Z. Profiles created in $\Delta \sigma = 0.2$ layers. Average profiles made up of only profiles containing 10 m/s updraft within evaluation layer. Number of profiles contributing to each layer are shown in the legend. Level refers to model level which increases from the top of the atmosphere (level = 1) to the surface (level = 25).
Figure 5.5 300 mb vertical motion (line) and 1-hr accumulated precipitation (fill) for 18 LST, 20 June, 1997 (00Z, 21 June). Contour intervals for vertical motion and precipitation are 2 m/s and 5 mm, respectively.
Figure 5.6  Same as Figure 5.5 for 19 LST, 20 June, 1997 (01Z, 21 June).
Figure 5.7  Same as Figure 5.5 for 20 LST, 20 June, 1997 (02Z, 21 June).
Figure 5.8  Same as Figure 5.5 for 21 LST, 20 June, 1997 (03Z, 21 June).
Figure 5.9  Same as Figure 5.5 for 22 LST, 20 June, 1997 (04Z, 21 June).
Figure 5.10 Magnification of Figure 5.7 focused on high precipitation region.
5.2 Vorticity and Mesoscale Circulation

General comparison of model dynamics associated with mesoscale convective complexes (MCC) will use the MCC composite structure of vertical motion and vorticity as outlined in Cotton et al. (1989). The upward mass flux associated with large vertical motions leads to horizontal divergence at the tropopause. A peak in anti-cyclonic vorticity (vorticity minimum) exists at the 200-300mb level, reaches a maximum during the mature stage, and remains large during the dissipation stage. Weak cyclonic vorticity is present in the lower troposphere throughout the life-cycle.

To diagnose the potential of the model to simulate MCC behavior, 300 mb and 800 mb relative vorticity were overlaid on 3-hour precipitation accumulations and analyzed for the integration period. Periods with both anti-cyclonic relative vorticity at 300 mb and cyclonic relative vorticity at 800 mb associated with areas of precipitation were considered events of interest. Relative vorticity and 3-hr accumulated precipitation at 300 and 800 mb are shown for 21 LST, 20 June (03Z, 21 June) in Figure 5.11 and Figure 5.12, respectively. Similar plots for 00 LST, 21 June (06Z, 21 June) are shown in Figure 5.13 and Figure 5.14. Relative vorticity fields have been smoothed using a Barnes objective analysis scheme (Barnes, 1964) with a radius of influence of 25 grid points. This radius of influence was chosen to filter out features with wavelengths longer than those associated with MCCs. Both time periods show regions of strong anti-cyclonic vorticity at 300 mb and strong cyclonic vorticity at 800 mb associated with the three-hour precipitation accumulations. While we cannot definitively conclude that this is an MCC, it should be considered as a candidate for further analysis.

A mesoscale filter similar to that used by Takle et al. (1999) and defined by Giorgi et al. (1993) was applied to further evaluate the localized flow features. This technique is intended to remove large scale flow features from the circulation leaving the mesocale departure of the local wind. In their paper on PIRCS (Takle et al., 1999), the authors selected a 9 x 9 point box centered on a local point of interest. The spatially averaged wind vector was calculated by averaging wind vectors at all model points within the box. This value was then subtracted from the local wind value. With a grid spacing of ~52 km, the average wind vector was taken
over an approximately 450 x 450 km box. Features remaining in the average wind vector are assumed to be those of synoptic scale (> 500 km) systems. With the forcing frame removed, our 5-km domain consists of 95 north-south and 135 east-west grid point. This results in a domain 475 x 675 km domain, or a domain only slightly large than the 9 x 9 box used in the Takle et al. (1999) study. Thus, all of the analysis domain grid points were used to calculate the average wind vector in this study. This average wind vector was subtracted from the local wind vectors and analyzed for the periods of interest found in the vorticity analysis above.

Figures 5.15 and 5.16 show the 300 mb horizontal vector wind field, the 300 mb mesoscale circulation, and the 3-hr accumulated precipitation for 21 LST, 20 June (03Z, 21 June) and 00 LST, 21 June (06Z, 21 June), respectively. An anti-cyclonic flow is present in the mesoscale circulation during both time periods. These are consistent with the anti-cyclonic vorticity found at 300 mb during these times periods (Figures 5.11 and 5.13) lending further support to the presence of an MCC. It should again be stated that this does not definitively point to the existence of an MCC during this time period, yet does indicate that this event should be given strong consideration for future analysis.
Figure 5.11  300 mb relative vorticity \((10^{-6} \text{ s}^{-1})\) and 3-hour accumulated precipitation (mm) for 21 LST, 20 June (03Z, 21 June). Relative vorticity (lines) and precipitation (fill) are contoured at \(2\times10^{-6} \text{ s}^{-1}\) and 5 mm, respectively.
Figure 5.12  800 mb relative vorticity (10⁻⁶ s⁻¹) and 3-hour accumulated precipitation (mm) for 21 LST, 20 June (03Z, 21 June). Relative vorticity (lines) and precipitation (fill) are contoured at 2x10⁻⁶ s⁻¹ and 5 mm, respectively.
Figure 5.13 Same as Figure 5.11 for 00 LST, 21 June (06Z, 21 June).
Figure 5.14  Same as Figure 5.12 for 00 LST, 21 June (06Z, 21 June).
Figure 5.15  Analysis of mesoscale circulation for 21 LST, 20 June (03Z, 21 June). The full horizontal vector wind fields in shown in a), the mesoscale circulation (full field - average) in b), and contours of the 3-hr accumulated precipitation in c).
Figure 5.16  Same as Figure 5.15 for 00 LST, 21 June (06Z, 21 June).
Increasing computational power is leading to simulations at small spatial scales that are of interest to hydrologists. Low-resolution models are known to over-estimate light-intensity precipitation events and under-estimate high-intensity events. As resolution increases, models are beginning to resolve small scale, high-intensity events and have the potential to provide better simulations. This study has evaluated the ability of a nested mesoscale model to simulate precipitation at the sub-catchment scale ($\leq 5$km) with the objective of determining if increasing resolution improves the simulation of summer convective precipitation in the central U.S. for a basin with little topographic influence.

A limited area model (MM5) with climate and chemistry modifications (Grell et al., 2000a) was run for the months of May, June, July, and August of 1997 for a domain centered over Central Kansas. The number of observed MCC (Mesoscale Convective Complex) and PEC (Persistent Elongated Convection) events that initiated and tracked through the region coupled with a positive precipitation anomaly over more than half the domain make this an attractive time period (Anderson and Arritt, 2001).

High-resolution (5-km grid spacing) model output was compared to a low-resolution integration (20-km grid spacing) completed by Dr. G. Grell and colleagues at the Forecast System Laboratory, Boulder, Colorado. The runs were performed using 128-nodes of a 350 node Linux cluster at Argonne National Laboratory through collaboration with Dr. John Taylor. The 5-km simulation was constructed to use the same physical parameterizations as the 20-km simulations to insure increasing resolution was the only factor impacting model output. Model output was compared to three observational datasets each derived from the National Climactic Data Center's (NCDC) cooperative observer program (co-op).
Comparison of domain averaged daily precipitation shows that both models adequately reproduce the timing of the observed time series, although they are less reliable with regards to accumulated amounts, over-producing in most cases. A similar pattern is found with May-August and monthly accumulations. Spatial patterns similar to those in the observations appear, although the model accumulations are often excessive. Domain averaged precipitation for the 5-km simulation are in excess of the daily co-op observations by approximately 80 mm (20%) and the 20-km simulation is closer to the observations with an excess of 20 mm (5%). While the 20-km simulation seems to outperform the 5-km simulation, intensity histograms show it does so through an over-estimation in the frequency of high intensity, less frequent events, and an under-estimation in frequency of the more frequent, light to mid-intensity events. The 5-km simulations also too frequently produce high intensity events compared to the co-op observations. By aggregating the 5-km output to match the effective grid spacing of the observation network, the aggregated 5-km output has a much better fit to the observations. This suggests the resolution of the observation network is too coarse to resolve events at the 5-km model resolution.

Histograms of vertical motion reveal that both models contain overly vigorous updrafts. Vertical motion profiles show large upward motions are primarily occurring at the upper levels and are at times consistent with the maximum precipitation rates observed in the diurnal cycle. Visual inspections of 3-hour accumulated precipitation overlaid with vertical velocity contours confirm that high intensity precipitation events are coincident with large upward vertical velocities. The profiles are consistent with Cotton et al. (1989) composite MCC vertical motion profiles which have upward vertical motion occurring throughout the profile with maximums occurring in the mature and dissipation stages around 300-400mb.

Visual inspection of 3-hour accumulated precipitation and relative vorticity suggest the model is generating mesoscale features consistent with MCC relative vorticity composites Cotton et al. (1989). Events with significant precipitation accumulations accompanied by strong anti-cyclonic relative vorticity at the upper levels (200-300 mb) and strong cyclonic relative vorticity at lower levels (800-900 mb) are observed. In addition, subtracting the domain aver-
age wind vector from each grid point in the domain demonstrate that these events also contain an upper level (300 mb) mesoscale circulation consistent with MCC structure. One such case (20 June, 1997) is analyzed in this study. While we cannot definitively conclude any of these events are MCCs, they should be considered as strong candidates for further analysis.

One downfall of the current study is the size of the model domain. MCC signatures were frequently seen on the domain boundaries or originated near a domain boundary and quickly propagated through the domain. A larger domain would have allowed us to track the MCC behavior more thoroughly and determine their lifetime within the model.

At this time, it is difficult to say whether one model out performs the other. Previous studies have shown high resolution can be of value in regions with strongly forced convection, or convection associated with fronts, drylines, or topography (Mass et al., 2002). In our domain, the 5-km simulation suggests that the model is producing precipitation events which, in fact, may be occurring, but are not resolved by the observational network. For a region with little or no topographic features, more analysis is needed to assess the value of increasing resolution. This study has generated several suggestions for follow-on analysis. They include:

1. Detailed analysis of vertical accelerations. Excessive vertical motions are occurring in both models. The cause of these over-ambitious vertical motions needs to be determined. A component breakdown of the vertical accelerations associated with these motions may point to a deficiency in the model framework which leads to the large updrafts, and, consequently, excessive precipitation intensities.

2. Change/modify convective parameterizations. A grid spacing of 5-km is still too large to explicitly resolve cloud scale updrafts/downdrafts and a convective parameterization scheme is required. The results of this study are dependent on the convective scheme used and the details of the scheme. Alternate convective schemes or variation in the current scheme parameters may lead to improved results.

3. Vorticity composites similar to those performed in Cotton et al. (1989). Model vorticity composites would go a long way in answering whether mesoscale circulations similar to
those found in MCCs exist in the model.

If the results of this and future analysis exhibit deficiencies in high-resolution models similar to their low-resolution counterparts, the benefits derived from high-resolution runs may not be worth the computational cost. Alternative methods, such as adaptive grid techniques, should then be explored.
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