ABSTRACT

Caldwell, Raymond Jason. Analysis of Model QPF Errors During the 2 - 4 December 2000 Snowstorm in North Carolina. (Under the direction of Gary L. Lackmann)

Model forecasts of an early season snowstorm for 2 - 4 December 2000 followed the historical blizzard of 24 - 25 January 2000 that dumped 20.3 inches of snowfall at the Raleigh-Durham International Airport. Much like the January 2000 storm, operational models exhibited a significant lack of skill, particularly in the realm of quantitative precipitation forecasting. As early as 1800 UTC 1 December, operational models from the National Centers for Environmental Prediction (NCEP), including the 32-kilometer Eta, generated liquid equivalent precipitation totals approaching two inches for the Raleigh-Durham metropolitan area. Later forecasts indicated as much as 2.77 inches of precipitation would fall. In reality, only a trace of precipitation was observed at the Raleigh-Durham airport.

A local, real-time version of the fifth-generation, mesoscale modeling system (MM5) was operational at the time of the event and provided a much-improved forecast scenario compared to the NCEP Eta model. Remarkably, the initial conditions and lateral boundary conditions in the MM5 were identical to those used to initialize the Eta model at 1200 UTC 2 December. In this study, an examination of the potential sources of error in the quantitative precipitation forecast is performed to...
challenge prior studies that suggest that data quality issues with sea surface temperature analyses led to spurious precipitation generation. The study includes a case study of the 2 – 4 December snowstorm, model sensitivity experiments, and quasi-geostrophic analysis to identify and diagnose the quantitative precipitation errors in the Eta model and the superior forecast guidance available from the local MM5 model.

The case study showed that several potential sources of model error existed including missing upper air soundings, sea surface temperatures, model design, and misdiagnosed topographic flow. This study will test the hypothesis that errors at the 500-hPa level led to limited precipitation early in the period and, hence, produced errors in the cold air damming, coastal front, and cyclogenesis in later periods responsible for the heaviest precipitation in model forecasts.

Results from sensitivity experiments with the MM5 model failed to exhibit significant differences in the representation of topographically induced phenomena or the westward extent of the precipitation shield into central North Carolina. The Eta model produced an anomalously strong 850 hPa jet at the North Carolina coast which transported warm air and moisture inland over the region. Better representation of the initial 500-hPa shortwave trough and associated vorticity maximum in the MM5 model is shown in the results to strengthen the low-level damming episode and shift the coastal front farther offshore.
The results of this study provide basis for further investigation into both models and concludes that the effect of the upper-level forcing on the evolution of the low-level topographically induced flow and the surface-based forcing of upper-level dynamics can be of equal magnitude and importance in winter season precipitation forecasting. Results of this study will be coupled with local efforts to improve forecasting through conceptual model development by providing operational forecasting with the knowledge that individual models can have independent and opposing response to initial condition errors based on the physical and dynamical make-up of the mesoscale modeling system.
ANALYSIS OF MODEL QPF ERRORS DURING THE 2-4 DECEMBER 2000 SNOWSTORM IN NORTH CAROLINA

by

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A thesis submitted to the Graduate Faculty of North Carolina State University in partial fulfillment of the requirements for the Degree of Master of Science

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Chair of Advisory Committee
DEDICATION

To face the truth is never more difficult than when your life is threatened by a terminal illness. It is at that moment that you must decide to believe in the greatest truth – God’s ability to perform miracles. With courage, faith, and perseverance, the challenges encountered in life become the successful journeys we will always fondly remember.

For the last thirty years, my mother has provided the love, hope, and encouragement necessary to surmount obstacles and achieve my dreams. For that reason, this thesis is dedicated to my mother, the foundation of my life and a living testament to the beauty of a miracle.

Thanks for everything! I love you, Mom!

“It is never too late to be what you might have been.”

--George S. Elliott
PERSONAL BIOGRAPHY

Jason Caldwell was born in Greenville, North Carolina, on September 25, 1974. As an infant, he and his family moved to the historical town of Cowpens, South Carolina - the place Jason calls home. In 1992, he began his college career at the University of South Carolina-Spartanburg and transferred to Northeast Louisiana University in 1993 to pursue a degree in Atmospheric Science. Jason graduated magna cum laude from NLU in 1997 and began work later that year as a Certified Weather Observer for a FAA-contract company at the Charlotte-Douglas International Airport in North Carolina. In 2000, Jason returned to his educational career when he received a graduate assistantship at North Carolina State University. He initially instructed several introductory meteorology labs and participated in a tropical wind reconstruction project. In 2001, Jason began research on winter season, model performance as part of the consortium known as the Southeast Centers for Mesoscale Environmental Prediction. In 2003, Jason joined the South Carolina State Climatology Office in Columbia, South Carolina, as the Severe Weather Liaison. Currently, Jason serves as a HAS Forecaster with the Lower Mississippi River Forecast Center in Slidell, Louisiana, where he focuses on applied research to improve the skill in quantitative precipitation and river forecast guidance.
ACKNOWLEDGMENTS

Maintaining a focus and forward progress has been the most significant challenge in completing my research during the last four years. Without the ever-present support of several people, the successful completion of my research project would have remained only a dream. First, I would like to express my sincere gratitude to Dr. Gary Lackmann and Dr. Allen Riordan for the patience and guidance they exhibited during the duration of the research. In addition, I am truly thankful for their valuable contributions to the final preparation of my thesis. Without their belief in my ability to perform and a true excitement for the weather, I would have lost sight and failed to achieve my academic goals.

There are a number of others who also deserve thanks for providing an environment conducive to the advance of scientific research in the realm of operational forecasting. Thanks to the staff of the National Weather Service in Raleigh, the State Climate Office of North Carolina, Capitol Broadcasting (WRAL-TV), and the North Carolina Supercomputing Center (NCSC) for the opportunity to participate in the advancement of science through collaboration across multiple aspects in the field of meteorology. The synthesis of ideas that occurred between floors in the Research III building on the Centennial Campus will be a lasting impression of the benefits of cooperative research. To the Forecasting Lab
staff, thanks for your support during some critical moments, helping to re-energize my productivity and renew my enthusiasm. Thanks to John McHenry and Kiran Alapati of NCSC for their assistance with the model development and testing through assistance with computer programming and instruction on the MM5 modeling system.

Since October 2003, I have worked intermittently on the research project and thesis while employed at the South Carolina State Climatology Office. I appreciate the extra time and extra nudges supplied by my friends at the South Carolina Department of Natural Resources to emphasize the importance of completing my degree, particularly to Freddy Vang, Hank Stallworth, Danny Johnson, Hope Mizzell, and Bud Badr. To the climate staff, thanks for sharing in the seemingly endless process called a thesis.

Finally, I would like to thank my family and friends for the endless support through the most challenging, yet most rewarding, period of my life. To my mom, Sandy, and dad, Carl, thanks for your undying faith in my ability and the wealth of love that has provided a stable foundation since birth.
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1. INTRODUCTION

1.1 Operational Model Errors

During the year 2000, the operational models available from the National Centers for Environmental Prediction (NCEP) [i.e. Eta, AVN/MRF, and NGM] exhibited a significant lack of skill during winter weather events along the East Coast (Zhang et al. 2002, Brooks et al. 2001, Petersen and McQueen, Eds. 2001, Grumm and Bryan 2001, Rogers et al. 2000). The most prolific errors occurred in three independent cases: (i) 24 – 25 January 2000, (ii) 2 – 4 December 2000, and (iii) 29-30 December 2000. In each storm, short-term model guidance (0 to 36 hours) failed to accurately represent the amount and distribution of wintry precipitation in regions from the Carolinas to New England. In January 2000, the Southeast received record snowfall totals of nearly two feet, including a major portion of central North Carolina. The total accumulated snowfall of 20.3 inches at the Raleigh-Durham International Airport (RDU) broke snowfall records dating back to 1887. Zhang et al. (2001) suggested that problems with the initialization fields, including missing upper air soundings, may have led to the errors in precipitation forecasts. While position and intensity errors in the model prediction of the coastal cyclone were not exceptional in the 24-25 January 2000 case, precipitation was drastically under-estimated from the Carolinas to Maryland. The Eta
model generated light precipitation only along the immediate coast and no precipitation inland where the highest totals occurred. Unlike the model forecast guidance in the January 2000 snowstorm, both December cases over-predicted precipitation in North Carolina.

At the University Center for Atmospheric Research (UCAR), Drs. Bill Bua and Stephen Jascourt designed a case study entitled ‘When Good Models Go Bad’ to use in the Cooperative Program for Operational Meteorology, Education, and Training (COMET). The development of the online case study evolved from problems with the sea surface temperature fields used to initialize the model in late 2000 (Thiebaux et al 2001). Several additional studies (Grumm and Bryan 2001, Brooks et al. 2001) indicate that the model precipitation errors in the 30 December 2000 event were a function of the data quality of initial conditions and lateral boundary conditions used to initialize and nudge the Eta model during integrations, including the sea surface temperatures. The model skill in winter season precipitation forecasts during December 2000 led to the development of a ‘Tiger Team’ at NCEP to address the model failures and to identify the potential source of errors. Through a series of changes in the operational configuration, the sea surface temperatures were changed from the 50-kilometer MCSST (McClain et al. 1985) in the 2-4 December case, to 1-by-1-degree Reynolds Optimum Interpolation (ROI)
(Reynolds et al. 1994) on 20 December, and, finally, in January 2001 to a new analysis method, the real-time global sea surface temperature analysis (RTG_SST) (Thiebaux et al. 2001).

The 50-kilometer MCSST data used at the time incorporated satellite-only retrievals on a grid composed of 7-day averages. Extensive cloud cover existed along the East Coast prior to the 2-4 December event; therefore, any migration of the Gulf Stream and change in sea surface temperatures in the region were representative of the conditions as far back as mid-November (Thiebaux et al. 2001). Supposedly, considerable smoothing of contours due to the averaging process, coupled with the inability to update data in regions of persistent cloudiness, created a problematic scenario for this particular data set.

1.2 Local Initiatives

The role of sub-synoptic scale processes in the development and evolution of winter weather scenarios in the southeastern United States is quite important. Variable topography from the Appalachian Mountains to the Gulf Stream in the western Atlantic Ocean provides the forcing required for the development of mesoscale meteorological phenomena with only regional or local impacts across the Carolinas and Virginia. Numerous studies, primarily by operational meteorologists, have
concentrated on the challenge of winter weather forecasting, particularly when topographically induced flow affects the region (Gurka et al. 1995, Maglares et al. 1995, Keeter et al. 1995).

In a recent effort, National Weather Service offices across the Carolinas and Virginia received funding through the Collaborative Science, Technology, and Applied Research (CSTAR) program to address these issues. Previous modeling experiments on the performance of larger-scale, operational (Fritsch 1992), and even mesoscale (Stauffer and Warner 1987, Kramer 1997), models during cold air damming and coastal frontogenesis events have failed to show skill beyond the synoptic scale. For that reason, the CSTAR program focused on improving the skill of winter weather predictions through the implementation of additional forecast tools, training programs, and methodologies at the operations level throughout the region. Initial research concentrated on the development of a comprehensive climatology of cold air damming and coastal front behavior to identify the synoptic environment at key times in the evolution of topographic flow regimes along the East Coast (Bailey et al. 2003 and Appel 2001). The climatological plots allowed operational forecasters to distinguish the damming classification and sensible weather impacts likely during an event based on model forecasts of the synoptic environment and precipitation. Late in 2003, the National Weather
Service offices across the Southeast and North Carolina State University collaborated again in a continued effort (CSTAR2) to improve winter weather forecasts with a focus on quantitative precipitation forecasts (QPF) during the cool season. The significant snowstorm of 2 – 4 December 2000 is an excellent example of the potential errors that often arise in model QPF guidance.

In a similar initiative, the North Carolina Supercomputing Center (NCSC), Capitol Broadcasting Corporation (CBC), the North Carolina State Climate Office (SCO), and North Carolina State University (NCSU) directed research towards contributions of improved skill levels in numerical weather prediction techniques as one element in the consortium known as the Southeast Center for Mesoscale Environmental Prediction (SECMEP). Unlike the typical philosophy associated with model development, the objective of the ‘Evaluation and Development of the MM5 Real-Time Forecast System’ project was not to reduce the overall statistical forecast errors for all events. Instead, with the flexibility to configure the model (i.e. domain, physical parameterizations, boundary conditions), sensitivity studies would provide optimum, situation-specific configurations with the least forecast error associated with such phenomena as cold air damming, coastal frontogenesis, sea breeze circulation, and tropical storms. Another objective of the SECMEP project
was to evaluate the performance of the Penn State University/National Center for Atmospheric Research (PSU/NCAR) fifth generation, mesoscale modeling system (MM5) relative to the NCEP Eta model during the winter season in the southeastern United States. Through a quantitative assessment of the superior guidance and capability of the MM5 model, the public would reap the benefits of reduced loss of property and life due to hazardous weather conditions.

The evolution of mesoscale modeling over the last 20 years has presented an equal set of challenges and triumphs to the forecast community. Increased resolution and portability of the computer models have provided meteorologists in a diverse population of professionals (i.e. broadcasting, National Weather Service, environmental consulting) with the ability to design a region-specific tool to meet the needs of local stakeholders and decision-makers with remarkable savings of time and financial expense. As part of the WRAL-TV project, key constituents across central North Carolina compiled a list of significant weather events from 1999 through 2000 that were associated with significant model errors from the Eta model, MM5 model, or both. The expected result of the project was to provide high-resolution model guidance that outperformed the operational Eta model at NCEP.
1.3 Research Objectives

The present study examines the model forecasts for the snowstorm that left four to twelve inches of snow across eastern North Carolina on 2 – 4 December 2000 (Figure 1.1). Operational models, however, predicted storm total liquid equivalent precipitation amounts of one to nearly three inches across much of central and eastern North Carolina. For example, the Eta model run from 1200 UTC on 2 December 2000, only 12 hours prior to precipitation onset, produced 2.77 inches of liquid precipitation at the Raleigh-Durham International Airport (RDU) (Table 1.1). The observed precipitation at the station was only a trace. The online case studies from the National Weather Service offices in Wakefield and Raleigh proposed several contributing factors to the forecast bust, including: (a) data quality errors in sea surface temperature (SST) fields, (b) a misdiagnosed lead shortwave trough/vorticity maximum at the 500-hPa level, (c) two missing upper air soundings (RAOBs) in critical locations for the 1200 UTC model runs on 2 December, and (d) under-prediction of the southern extent, strength, and depth of the cold wedge with implications that errors also existed in the coastal front location and eventual cyclone development and track.

While the Eta and other operational models indicated snowfall totals that would rival the 24 – 25 January 2000 record snowstorm at RDU,
some support for a lighter, though still significant, snow in the Raleigh area was provided by the 1200 UTC run of the local, real-time mesoscale model (MM5) on 2 December 2000. Although errors in the sea surface temperatures are proposed as the primary source of error for the winter weather events in 2000 (Zhang et al. 2002), several sensitivity studies with the MM5 model suggest that this hypothesis is not of primary importance. Discussion and results of the sensitivity experiments are presented in Chapter 5. The MM5 model produced a superior forecast compared to the Eta model for the event, especially with respect to the westernmost edge of the precipitation shield (Figure 1.2). The fact that the MM5 model used the same initial conditions and forecasts from the Eta model for lateral boundary conditions implies that errors in the precipitation distribution and total accumulation are not related to errors in sea surface temperature analyses, but rather may be a function of the model design. The predicted storm total liquid precipitation from the MM5 was 0.75 inches with the heaviest six-hour precipitation totals from the 24- and 30-hour forecasts displaced slightly west of observations (Figures 1.2a and 1.2b). The MM5 model did identify the sharp east-to-west gradient in the isohyets across Wake County unlike the Eta model forecasts for the same period (Figures 1.2c and 1.2d) which over-predicted precipitation across much of central North Carolina. The
superior forecasts from the MM5 model compared to the Eta model present an important consideration when evaluating the role of initial conditions in the evolution of model-generated precipitation errors. The present study suggests that, despite a distinct warm bias in the sea surface temperatures near the North Carolina coastline, additional factors must also be present to induce the large errors in quantitative precipitation forecasts from the Eta model. Therefore, the current study will evaluate the role of the lead 500 hPa disturbance in the production of precipitation in the Eta model compared to the MM5 forecast guidance. The design of the current project is based on achieving the main objectives of the case study:

(1) Identify the meteorological factors and processes important in the development of winter season precipitation in the southeast United States,

(2) Describe the operational modeling systems available in December 2000,

(3) Perform a case study to compare the observed conditions, Eta forecasts, and MM5 forecasts of synoptic conditions and precipitation,

(4) Propose a hypothesis based on the initial case study,

(5) Test the role of model design in the superior performance of the MM5 model relative to Eta,
(6) Compare observations and forecasts of atmospheric conditions using quasi-geostrophic analysis to identify individual model errors and differences between models that potentially contributed to errors in QPF, and

7) Summarize the findings and propose future work.

To fully understand the complicated nature of winter weather forecasting in the Southeast during the winter months, the challenges associated with topographic flow regimes (Maglares et al. 1995, Keeter et al. 1995, and Gurka et al. 1995) are presented in Chapter 2 and serve as background for this project. Physical and dynamic processes and the interactive feedbacks with precipitation will conclude the chapter. The two mesoscale modeling systems are described in chapter 3 to convey the influence of model structure and design in the operational forecast of winter precipitation. Model performance and systematic biases of the Eta and MM5 model in topographic flow and precipitation forecasting during the cool season follow the basic model information. Chapter 4 contains an independent case study with observational data, in situ data, and model forecasts from the 2-4 December 2000 event to identify errors in model forecasts relative to the representation of topographically forced phenomena and the magnitude and distribution of precipitation across North Carolina. At the conclusion of Chapter 4, the contributing factors
from Chapter 1 are considered and the main hypothesis for the current research project is proposed. Chapter 5 introduces the tests and results that identify the lack of response by the MM5 model to various sensitivity experiments and justify testing additional sources of error. The difference in Eta and MM5 model cross-sections relative to the 500-hPa disturbance and topographically induced flow with implications on precipitation forecasts are also presented as results in Chapter 5. Finally, Chapter 6 summarizes the case study background, experiments, and major findings. Any new discoveries and considerations for future research conclude the thesis.
2. LITERATURE REVIEW

2.1 Southeast Winter Weather Forecasting

2.1.1 Cold Air Damming

Along the eastern seaboard, the Appalachian Mountains extend from New England south-southwest into the northern part of Georgia in the heart of the Southeast. Only several hundred miles to the east, the warm Gulf Stream flows northward along the North Carolina coastline bringing moisture-rich, tropical air into a region that is no stranger to the influx of arctic air during the winter season. Cold, dry air is typically impeded by the mountain chain as a ridge of high pressure forces arctic air southward from central Canada across the Plains and Midwest states on northwest winds. An average of two to three times per month, however, arctic high pressure from Canada slides to the east and southeast over the Great Lakes and New England into a favorable position for the topographically induced phenomenon of cold air damming to occur (Richwien 1980). First described by Baker (1970), the inverted ridge of high pressure over the eastern United States, in response to the orientation of the Appalachian Mountains, has been thoroughly investigated (Stauffer and Warner 1987, Forbes et al. 1987, Bell and Bosart 1988, Fritsch et al. 1992, Bailey et al. 2003). From this research, the processes involved in the development and evolution of cold air...
damming became a primary consideration of operational forecasters in the damming region east of the Appalachian Mountains. For the purposes of this report, the damming region is defined as the region east of the mountain chain in both Carolinas and Virginia.

The clockwise flow around an area of high pressure situated in southeastern Canada or across the northeastern part of the United States results in east or northeast winds orthogonal to the Appalachian Mountain chain in the Mid-Atlantic region. As the momentum of an air parcel approaching the barrier decreases, the northward directed Coriolis force weakens and the air parcel turns south in response to the unbalanced pressure gradient force and ageostrophic northerly flow. The initial trajectory of air parcels, coupled with the Coriolis deflection of the ageostrophic flow toward the mountain barrier, results in mass accumulation, adiabatic cooling, and a hydrostatic pressure rise. Cold advection on northeasterly winds occurs east of the mountain chain as the resultant total wind lies between the easterly geostrophic and northerly ageostrophic wind components in topographically forced flow. Rising surface pressures and lowering temperatures at low-levels continue to stabilize the atmosphere as the cool wedge moves southward to produce the typical “U-shaped” isobars known as the signature of the Baker ridge responsible for cold air damming.
The feedback role of cold advection on the damming can be described as contributions of increasing static stability, enhanced blocking, and resulting in additional accumulation of mass and hydrostatic pressure rises. The Froude number (Bailey et al. 2003) is used to determine the strength of the blocked flow as the relationship between the flow perpendicular to the barrier, the height of the boundary, and the Brunt-Väisälä frequency (N) which is proportional to the difference in the dry adiabatic lapse rate and the environmental lapse rate. Large values for static stability, therefore, correspond to higher values of N and lower Froude numbers.

2.1.2 Coastal Frontogenesis

As the cold air moves southward from the Mid-Atlantic region, another topographical feature specific to the southeastern United States becomes an important consideration to forecasters, the Gulf Stream. The temperature, moisture, and stability gradients that exist across the Carolinas and Virginia create an enhanced zone of baroclinicity that culminates in the generation of the coastal front. The eastern edge of the cold dome and the location of the coastal front have been correlated in several prior studies (Bosart et al. 1972, Riordan 1990, Appel 2001). The influence of the coastal front on local weather conditions is primarily with
regards to the incipient cyclogenesis; although, temperature forecasts can be complicated by the movement of the coastal front inland.

Kocin and Uccellini (1990) describe the process involved in the generation of the coastal front and its impact on coastal cyclogenesis relative to the development of significant East Coast snowstorms. As cold air damming becomes entrenched to the east of the Appalachian Mountain chain, low-level warm advection is suppressed over inland regions, which forces the maximum temperature gradients offshore. Warm advection patterns ahead of dissipating cyclones approaching from the west are shifted east of the cold dome. The same high pressure responsible for the cold air damming provides easterly flow over the ocean waters where friction is reduced and the flow remains nearly geostrophic at the surface. The deformation zone located at the convergence of easterly and northerly winds along the North Carolina coastline enhances low-level baroclinicity. This region of maximum low-level convergence, baroclinicity, warm advection, and surface vorticity can be identified by the location of maximum potential temperature gradient, also denoted as the coastal front (Bosart 1975). The development, location, and movement of the coastal front are indicative of the favored location for cyclogenesis and the track of the cyclone (Bosart 1981, Stauffer and Warner 1987, O’Handley and Bosart 1996).
2.1.3 Coastal Cyclogenesis

Along the East Coast of the United States, cyclones generally develop as Type A or Type B storms as categorized by Miller (Kocin and Uccellini 1990). Type A storms form along the polar front and follow the development stages of the typical extratropical cyclone from the Norwegian school of thought described by Bjerknes (1919). The Type B system is most common along the coastal Carolinas and Virginia and involves secondary cyclogenesis to the southeast of a primary dissipating low over the Ohio Valley to the west of the Appalachian Mountains. The secondary cyclone forms along the coastal front. Kocin and Uccellini (1990) selected twenty prominent storms from the period 1955 through 1985 to evaluate in a survey of snowstorms affecting the northeast United States. The storms were divided equally between Type A and B development scenarios. Of the ten Type A cases, however, seven storms exhibited center jumps as the cyclones moved northeastward along the Southeast coast, lending evidence to the importance of the coastal front even in events without secondary cyclogenesis.

The offshore waters of the Carolinas and Virginia are favored for cyclogenesis due to the stretching of vortices that pass over the mountain chain and deformation zones associated with concave coastlines (Petterssen 1956). Cold shelf waters along the immediate coast of North
Carolina in the winter enhance the steep temperature and moisture gradients over distances of less than 100 kilometers. The development of a coastal front at the position of the potential temperature gradient serves as a focus for converging winds at the surface. This natural baroclinic zone coincides with decreased static stability, ample moisture, and a region where the frequent passage of jet level disturbances enhance the potential for cyclogenesis. The interaction of upper-level forcing for ascent, low-level instability, and surface convergence at the coastal front leads to the climatologically favored region of cyclone development. From the extensive climatology of East Coast snowstorms, Uccellini and Kocin (1987) identified the meteorological phenomena shared by all the events as factors in the development of cyclones along the eastern seaboard and separated the processes by scale. The synoptic scale has horizontal dimensions greater than or equal to 2000 kilometers and time periods of 24 to 48 hours. Smaller scale phenomena, considered mesoscale or meso-α scale, extend over 100 to 2000 kilometers with life cycles of 3 to 24 hours. A schematic representation from Kocin and Uccellini (1990) showing the physical and dynamic processes involved in coastal cyclogenesis is presented in Figure 2.1.

On the synoptic scale, the upper-level trough and ridge patterns provide vorticity and kinetic energy through divergence and associated
ascent. Decreasing wavelength and increasing amplitude occur as diffluent flow develops downstream of the negatively tilted trough axis. Confluence over New England in the cyclonic flow around a second trough in southeastern Canada builds cold high pressure southward and assists in the development of cold air damming. So, while large-scale vertical motion and the source for cold air for snowfall generation result from processes on the synoptic scale, additional forcing for cyclonic development occurs at smaller scales.

Upper-level jet streaks increase the divergence aloft, available kinetic energy, and vorticity associated with the synoptic scale pattern through indirect and transverse circulations. Processes in the lower troposphere, related most strongly to the topographical variations across the eastern United States, enhance the gradients and transport of temperature and moisture quantities along the baroclinic zone. Development of a low-level jet generates warm advection over the top of the cold dome and enhances ascent through isentropic lift. Sensible and latent heat flux in the oceanic planetary boundary layer (PBL) in the vicinity of the Gulf Stream contribute to the buoyancy of the air mass through the addition of heat and moisture to the lower troposphere and, hence, to the decreased stability of the atmosphere.
2.1.4 Precipitation Feedbacks and Topographic Flow

The physical and dynamic processes involved in cold air damming, coastal frontogenesis, and coastal cyclogenesis influence the distribution of heat and moisture in both the vertical and horizontal dimensions. Cold air damming significantly increases the low-level static stability within the cold dome. Increased blocking associated with cold advection enhances subsidence and limits the upward vertical motion to regions above the low-level inversion. The vertical temperature profile is important in the determination of the precipitation type and efficiency related to cloud microphysical processes and ice crystal growth. The southern progression of the Baker ridge determines the eventual location and intensity of the thermal and moisture gradients that provide favorable atmospheric conditions for development of the coastal front. Low-Level convergence and decreased stability at the coastal front enhance vertical motion in a region where moisture availability is maximized. This complex situation is the reason for complicated winter season forecasting in the southeastern United States during the winter season due to the variety of feedbacks that occur and the sensitivity of precipitation forecasts over inland North Carolina to the location, track, and intensity of the coastal cyclone. In Chapter 3, the description of the model physics and design will provide insight on the mechanics of models with respect to QPF.
2.2 Quantitative Precipitation Forecasting

The relationship between precipitation and topographically induced weather phenomena in the Southeast comprises a series of complex feedbacks on a variety of meteorological scales of motion, ranging from the synoptic and planetary scale down to the mesoscale and microscale environment. The development of precipitation in the damming region and offshore near the Gulf Stream results in diabatic heating and cooling through evaporational cooling inland and latent heat release offshore. The evaporation of precipitation into the low-levels of the atmosphere within the cold dome can account for up to thirty percent of the total reduction in temperature observed within the lower troposphere (Bell and Bosart 1988) during cold air damming events. Fritsch et al. (1992) indicated that increased surface pressures on the order of 2 hPa could be attributed to the evaporation process and solar sheltering from precipitation and clouds within the damming region. Decreasing surface temperatures, coupled with warm advection in southwesterly flow aloft (Forbes et al. 1987), decreases the environmental lapse rate in the planetary boundary layer with the most significant contribution at the top of the PBL. The increased static stability results in lower values of the Froude number to enhance the development of the cold pool through impeded cross-barrier flow, additional mass accumulation, and hydrostatic pressure rises.
The impact of precipitation on winter weather forecasting in the Southeast is not limited to the diabatic processes associated with the phase changes during condensation and evaporation processes (Lackmann et al. 2002). During ice storm events, the latent heat release from the freezing of liquid precipitation at the surface can modify the thermal profile in the lowest levels to inhibit additional freezing or frozen precipitation. If precipitation rates are low and the supply of dry air is impeded, the latent heat of freezing in the low-levels can exceed the rate of cooling by evaporational processes. Likewise, melting snowflakes, in conjunction with evaporation, can cool the upper levels of the atmosphere sufficiently to promote snowfall as the predominant precipitation type by producing a thermodynamic profile that is nearly isothermal at or below 0 degrees C. These complex interactions provide only brief insight into the important feedback relationship that exists between precipitation and topographically forced flow regimes in the Carolinas and Virginia. The impact of the meteorological characteristics of the cold dome, coastal front, and coastal cyclone on the distribution of precipitation also illuminates the non-linear pattern of physical and dynamic response and forcing associated with winter weather in the Southeast.

In North Carolina, quantitative precipitation forecasts (QPF) during the winter season often include the mention of rain, freezing rain, sleet,
and snow within a single forecast at a single location. As evidence to the high impact associated with incorrect precipitation forecasts, the National Weather Service in Raleigh developed several empirical models, including a precipitation type nomogram based on model-generated vertical profiles to determine the phase state of precipitation reaching the ground. An accurate forecast for precipitation type is of particular concern to public interests; however, identifying the location, timing, and amount of freezing and frozen precipitation is equally important. For that reason, the quantitative precipitation forecast (QPF) entails the consideration of the precipitation potential, intensity, duration, and type.

Wetzel and Martin (2000), henceforth WM00, designed a multifaceted methodology to address the challenge of predicting winter precipitation using an ingredients-based technique. The five ingredients included in the recipe for cool season precipitation are remarkably correlated with the physical components and processes in East Coast cycogenensis from Kocin and Uccellini (1990) described in Section 2.1.3. WM00 propose using the following ingredients: (i) forcing for ascent, (ii) moisture, (iii) instability, (iv) precipitation efficiency, and (v) temperature (Figure 2.2). Doswell et al. (2002) commented on the ingredients based approach outlined by WM00 with particular concerns regarding the definition of an ingredient as a necessary contributor to the generation of
precipitation. While Doswell et al. (2002) noted problems with the WM00 method in quantitative forecasts of precipitation, a qualitative evaluation of the ingredients may identify model differences responsible for precipitation errors during the 2 – 4 December 2000 case. The methodology used by WM00 incorporates the quasi-geostrophic theory to describe the atmospheric tendency to produce vertical motion. In the following sections, the concept of quasi-geostrophic flow will precede the full description of the ingredients-based methodology used to diagnose model differences responsible for spurious precipitation generation in the testing performed in Chapter 5.

2.2.1 Quasi-geostrophic Theory and Vertical Motion

At the synoptic scale, horizontal flow aloft in the mid-latitudes is considered to be approximately geostrophic. Here, the pressure gradient force (PGF) and Coriolis (COR) are in near perfect balance. This balance of active forces allows the three-dimensional velocity vector to be largely approximated as flow parallel to the isobaric surfaces. The quasi-geostrophic theory allows simplification of the equations of motion to solve explicitly for the pressure vertical velocity, $\omega$, as a function of temperature and vorticity from the derivation in Holton (1992, p. 167) as
\[
(\nabla^2 + \frac{f_0^2}{\sigma} \frac{\partial^2}{\partial p^2}) \omega = \frac{f_0}{\sigma} \frac{\partial}{\partial p} \left[ \Phi \cdot \nabla \left( \frac{1}{f_0} \nabla^2 \Phi + f \right) \right] + \frac{1}{\sigma} \nabla^2 \left[ \Phi \cdot \nabla \left( -\frac{\partial \Phi}{\partial p} \right) \right]
\]

where the right hand side (RHS) represents the forcing for vertical motion as the horizontal temperature advection (Term C) and the rate of change of differential horizontal vorticity advection with height (Term B). The left hand side (LHS) is interpreted as the response to the forcing for ascent on the RHS of the equation.

The quasi-geostrophic approximation assumes that the atmospheric flow is nearly geostrophic and hydrostatic. Therefore, the advection of vorticity and thermal fields from the RHS represents disruption of the geostrophic and hydrostatic balance mandated in the assumptions. In Term C, temperature changes occur through advective processes and adiabatic ascent and descent. The adiabatic flow associated with QG theory allows parcels to maintain the same potential temperature; therefore, parcels in a theoretical sense must travel along isentropic surfaces. Upward sloping surfaces along warm frontal boundaries are generally associated with warm advection and, hence, upward vertical motion. The opposite is true in regions of cold advection. The thermal advection term can also be related to the negative of the Laplacian of
thickness advection or, more generally, the thickness advection. The inverse relationship requires warm advection and increasing thickness to guarantee that Term C is positive. Cold advection is considered in the same manner and results in downward vertical motion as the depth between adjacent geopotential height surfaces fluctuate to uphold the constraints of the ideal gas law in Equation 2,

\[ p\alpha = RT \]  

(2)

As the mean layer temperature increases (decreases) then the specific volume also must increase (decrease) to compensate for thermal advection, provided that the process is both hydrostatic and isobaric.

Vorticity changes in Term B arise from the advection of relative and/or planetary vorticity that in turn reflect in the divergence or convergence at a particular level. Term B is defined as the differential vorticity advection with height. For disturbances with short wavelengths less than 3000 kilometers, the values of relative vorticity exceed the magnitude of planetary vorticity. As the geostrophic wind blows from maximum values of relative vorticity at the ridge crest to minimum values at the base of the trough, the result is anticyclonic, or negative, vorticity advection (AVA) at that level. At the surface, however, planetary vorticity is favored which is opposite in sign to the relative vorticity at the ridge crest. With cyclonic, or positive, vorticity advection (CVA) present below
the region of AVA, the net change in absolute vorticity with height becomes more negative with time; therefore, AVA (CVA) at upper levels of the atmosphere implies downward (upward) vertical motion.

From another perspective by Charles Doswell (Doswell 1995), an idealized geopotential height field is depicted with the cyclonic vorticity maximum at the base of the trough and the anticyclonic vorticity maximum at the ridge crest (Figure 2.3). The influence on vertical motion can be described through the concept of conservation of momentum. If parcel A travels along the flow toward higher values of absolute vorticity, the parcel is mandated to converge to maintain equilibrium with the environment. The same concept of conservation of angular momentum is involved with an ice skater when the arms are pulled in (convergence) towards the center of circulation to increase the rate of spin (increased vorticity). Therefore, regions of AVA upstream and CVA downstream of the trough axis are associated with convergence and divergence at the respected level, respectively. There is a crucial assumption that is required to imply three-dimensional motion and vertical ascent/descent from the properties of vorticity advection. Figure 2.4 identifies the requirement that the level of non-divergence must exist below the level of consideration to support the laws of mass conservation.

The interaction of thermal advection and vorticity advection at
different levels in the atmosphere ultimately determines the net vertical motion throughout the column, along with several other processes that occur in response to forcing. For example, adiabatic cooling (warming) processes associated with ascent (descent) can partially cancel the effects of thickness advection at the location of the crest (trough) in geopotential height fields. In general, however, Table 2.1 indicates the relationship between vorticity advection and the processes attributed to vertical motions within the quasi-geostrophic framework.

2.2.2 Ingredients-Based Methodology

As previously identified, the ingredients used by WM00 in the diagnosis of precipitation include: (i) forcing for ascent, (ii) moisture, (iii) instability, (iv) efficiency, and (v) temperature. The location and strength of the forcing mechanisms are the primary concern of the initial ingredient. The generation of upward vertical motion is essential to the cloud development process. Using the components of the quasi-geostrophic omega equation in Equation 1, evaluation of the vorticity advection, thermal advection, convergence/divergence patterns, and isentropic analyses at various levels in the atmosphere assists meteorologists in identifying areas conducive to upward vertical motion. The stability adjustment parameter in the QG analysis is not included in
the quantification of the forcing for ascent using QG theory. Values for the intensity of forcing are directly related to the intensity of precipitation, as indicated in Table 2.2.

To determine the amount and spatial distribution of moisture is the objective of the second ingredient. Moisture available is typically determined using thermodynamic soundings, model vertical cross-sections, and satellite imagery. Without significant atmospheric water content, even regions with strong mechanisms for lift will fail to produce precipitation. Limited moisture inhibits the condensation process normally associated with adiabatic cooling by ascent. Hence, the moisture availability is also essential to the generation of clouds and precipitation. Using the aforementioned tools, a complete horizontal and vertical profile of water vapor will differentiate areas based on instability, efficiency, and precipitation type. For example, additional moisture at low-levels increases the buoyancy of parcels and, therefore, results in decreased stability of the lower atmosphere. Co-located regions of vertical ascent and relative humidity values exceeding 80 percent are indicative of areas highly conducive to precipitation generation.

The stability is a determining factor in the response to forcing mechanisms. The spatial and temporal distributions of heat and moisture quantities in the atmosphere are considered with respect to the stability
parameter within the quasi-geostrophic framework. The stability parameter includes only the gravitational instability and not symmetric instabilities. For symmetric instability to exist, the atmosphere must be saturated at the level considered and forcing for upward vertical moisture must also be present.

The efficiency of precipitation generation during the winter season depends heavily on the generation of ice crystals through microphysical processes. Errors in the precipitation rate can result from misdiagnosis of the active phase conversions and in-cloud thermal environment. As described in WM00, the maximum range of cloud temperatures for a significant chance that ice would be present in the cloud was –12 to –14 C with the greatest potential at temperatures below –15 C and lesser odds above –10 C.

Temperature distribution within the atmosphere is of great importance as well in the determination of precipitation type. Heat transferred from water droplets to ice crystals is proportional to the temperature difference between the ice crystal (~ 0 C) and the surrounding environmental air, and the time spent at the environmental temperature during the phase change from liquid to ice. The vertical profile of temperature is affected by thermal advection, adiabatic temperature changes due to vertical displacement, and non-adiabatic
heating/cooling. The predominant source of temperature variability is the thermal advection by the horizontal wind.
3. MODELS

3.1 The NCEP Eta Model

As of 2 December 2000, the Eta model was operationally available from NCEP four times daily with model integrations beginning at 0000, 0600, 1200, and 1800 UTC. Each model run contained forecasts out to 48 hours and was initialized with full-cycled Eta Data Assimilation System (EDAS) data. The assimilation combined observational data with the three-hour forecasts from the Eta model to develop an analysis for initial and lateral boundary conditions. The Eta is a grid point model with the prognostic variables of temperature, pressure, and moisture computed at the center of each grid cube as a volumetric average. The horizontal and vertical wind components are prognostic variables determined at the sides of each box as the average between the centers of two adjacent boxes containing that grid point. Improvements to the Eta model prior to the event studied include changes implemented in September 2000. At that time, the resolution increased from a 32-kilometer grid with 45 vertical layers to a 22-kilometer grid with 50 vertical layers (Figure 3.1) with a 10 to 30 percent enhancement in the vertical resolution throughout the troposphere with the greatest improvement in the later from 500- to 250-hPa (Rogers et al. 2000). The domain was also extended about 900 kilometers to the west of Hawaii in the Pacific Ocean (Figure 3.2).
The Eta model used the eta (\( \eta \)) vertical coordinate system developed in the early 1980s as an alternative to the sigma (\( \sigma \)) coordinate system described later in Section 3.2 with the MM5 model. Using the sea level pressure at the bottom of the model, \( \eta \) is computed in Equation 3,

\[
\eta = \left( \frac{p - p_T}{p_{sfc} - p_T} \right) \left( \frac{p_{ref}(z_{sfc}) - p_T}{p_{ref}(0) - p_T} \right),
\]

where \( p \) is the pressure at the specified height, \( p_T \) is the pressure at the top of the model (25 hPa), and \( p_{ref} \) is a function of the distance above sea level. The bottom of the model is represented as a flat step; therefore, this system is often referred to as ‘step-mountain’ coordinates.

A modified version of the Betts-Miller (Betts and Miller 1986, Betts and Miller 1993) scheme is used to parameterize the convective processes (Janjic 1994). In the Betts-Miller-Janjic (BMJ) scheme, the atmosphere is adjusted toward a reference state, a thermodynamic profile that represents quasi-equilibrium in a post-convective environment. The initial step in the scheme involves lifting the most unstable air parcel to the lifting condensation level (LCL) that serves as the definition of the model cloud base. From the LCL, the parcel is subjected to moist adiabatic ascent to the equilibrium level (EL). If the parcel is not buoyant at any level, then there is no activation of
convection in the BMJ scheme. If the cloud depth, defined by the distance between the LCL (cloud base) and, typically, the model level immediately below the EL (cloud top) is less than 200 hPa, then the model activates shallow convection. For cloud depths greater than 200 hPa, the model checks for the activation of deep convection. To determine if deep convection will occur, the BMJ scheme requires that the enthalpy of the model grid column be conserved. Latent heat release associated with convection and condensation must be directly proportional to the net removal of water vapor from the atmosphere. Therefore in regions of deep convection, the BMJ requires ample cloud layer moisture to initiate and sustain the model warming and drying associated with the development of precipitating convective clouds (Baldwin et al. 2002).

The precipitation scheme in the Eta model incorporates both cloud water and cloud ice into the convective and grid-scale precipitation schemes. The environmental temperature, $T$, and cloud top temperature, $T_p$, determine the cloud hydrometeor type. At temperatures over 0°C or below −15°C, the phase of cloud hydrometeors is considered to be entirely water or ice, respectively. When the environmental temperature falls between 0°C and −15°C, the cloud top temperature determines the phase state present within the cloud. At cloud top temperatures below −15°C, ice crystals form and water droplets seed to ice. For warmer cloud
top temperatures, the water is considered to be super-cooled liquid water (SLW). Large scale condensation over land-based areas occurs when the relative humidity equals or exceeds 75 percent and over water when greater than or equal to 80 percent. Precipitation is calculated from the cloud water-to-ice mixing ratio that is explicitly determined by its own prognostic equation. Once condensation occurs, the precipitation either falls to the ground or evaporates in lower model layers with relative humidity values below the thresholds within one time step. The convective microphysics scheme includes auto-conversion of cloud water to rain, collection of droplets by falling rain, auto-conversion of ice to snow, collection of ice particles by falling snow, melting snow at the 0 C line, and evaporation of precipitation below the base of the cloud.

3.2 The Local, Real-Time MM5 Model

The NCAR/PSU fifth-generation, mesoscale modeling system (MM5) is a community model available for download to educational institutions, government agencies, and scientific researchers. As of 2 December 2000, version 2.12 of the modeling system was configured locally at the North Carolina Supercomputing Center (NCSC) to run twice daily at 0000 and 1200 UTC with forecasts out to 48 hours with a horizontal grid spacing of 45-kilometers. The model contained a multiple nest with the 45-kilometer
outer domain over the eastern half of the United States, primarily east of the Mississippi River. The inner domain was positioned over the Southeast with a horizontal grid spacing of 15-kilometers (Figure 3.3). The inner nest initialized at forecast hour 9 and ran through forecast hour 33, yielding 24 hours of high-resolution model guidance for the region. Like the Eta model, the MM5 was designed as a grid point model with prognostic variables computed at grid box center as volumetric averages and wind vectors determined at the grid box sides as centered averages of adjacent grid boxes. The MM5 was initialized using the EDAS analysis at forecast hour 0, the same used to initialize the Eta model. If the Eta was unavailable, the AVN model fields were used in place of the EDAS. This was not the case in the 2 December 2000 event. The EDAS served as lateral boundary conditions for the MM5 model, as well.

The MM5 model configuration used at the time of this event had 31 vertical layers based on the $\sigma$-coordinate system in non-hydrostatic form (Figure 3.4). Sigma coordinates avoid pressure or height surfaces that intersect the ground in mountainous regions; therefore, $\sigma$ is often referred to as the ‘terrain-following’ coordinate. The values of $\sigma$ range from 1 at the surface to 0 at the model top (50 hPa) and can be
determined by Equation 4,

\[
\sigma = \left( \frac{p - p_t}{p_t - p_s} \right).
\] (4)

Since \(\sigma\) is related directly to pressure, the equations of motion are simplified in the boundary layer.

The cumulus parameterization scheme used operationally in the local, real-time version of MM5 in December 2000 was the Kain-Fritsch (Fristch and Chappell 1980, Kain and Fritsch 1993). The Kain-Fritsch (KF) scheme is specifically designed for high-resolution, mesoscale models with horizontal grid spacing of 10 to 30 kilometers. The convective available potential energy (CAPE) determines the extent of cloud development by the ability of air parcels to reach the level of free convection with a sustained upward vertical velocity. The removal of all CAPE within a single time step assures the closure assumption is fulfilled. Separate updrafts and downdrafts are calculated with parcel entrainment into the updrafts and detrainment at the cloud top and base in the downdrafts. The KF scheme parameterizes a thorough set of in-cloud physical processes including downdrafts to allow better representation of mesoscale response to convective processes.

The Dudhia simple ice (SI) microphysical scheme (Dudhia 1993) served to parameterize the precipitation development process in the MM5 model.
Ice phase processes occurs at temperature below 0 C at which point cloud water and rain immediately converts to cloud ice and snow, respectively. Separate predictive fields in the MM5 model also included a field for water vapor; however, no calculation of super-cooled liquid water (SLW) was performed in the SI scheme.
4. CASE STUDY

4.1 EDAS Analysis

The proper representation of the atmospheric conditions across the eastern United States from the model initialization time of concern at 1200 UTC 2 December through 1200 UTC 3 December 2000 is crucial to the diagnosis of model forecast errors. Indications of the initial differences in both models from the EDAS analysis at initialization time will provide substantiation for the potential model errors throughout the depth of the atmosphere, including those proposed by the National Weather Service offices in Wakefield, Virginia, and Raleigh, North Carolina. Using the EDAS analysis, instead of an independent objective analysis such as RUC, allows the evaluation to favor the Eta over the MM5 model analysis due to the inclusion of the Eta forecasts in the EDAS analysis. To evaluate how well the EDAS analysis represents the observed conditions across the Southeast, Figures 4.1, 4.3, and 4.4 provide overlays of upper air and surface observations with EDAS analyzed contours at the surface and standard pressure levels of 850, 500, and 250 hPa. At 1200 UTC 2 December, the radar and satellite imagery are also included (Figure 4.2).

At 1200 UTC 2 December, relative to the observational data, the EDAS sea level pressures are higher by 1 to 2 hPa (Figure 4.1a) over the
southern Appalachian Mountains from North Carolina into Virginia. At 850 hPa, the temperatures over central and northern portions of Georgia and Alabama are not consistent and differ by several degrees. At Peachtree City, Georgia, for example, the observed upper air sounding provided a 0 degree C reading while EDAS produced temperatures between 0 and 5 degrees C (Figure 4.1b). At Birmingham, Alabama, on the other hand, actual observations were 5 degrees C with EDAS analysis between 0 and 5 degrees C. The winds and heights at 500 hPa appear reasonable at first glance over the Southeast. Further inspection of the temperature fields in Nashville, Tennessee (BNA), suggests that the observed temperatures are much cooler than EDAS represents with the –25 degree C contour well into Kentucky at 1200 UTC 2 December (Figure 4.2c). The heights in EDAS analysis over northern Georgia and Alabama at the 250-hPa level (Figure 4.1d) also differ with height differences of 10 to 20 m compared to observations. Even at BNA, the 10320 m EDAS contour is analyzed within 100 kilometers of the 10320 m observed height from the sounding. Despite the differences at the initial time considered, the overall synoptic pattern is well represented by the EDAS analysis at 1200 UTC 2 December 2000. The location of the upper level vorticity maximum across northern GA can be seen through investigation of the radar (Figure 4.2a) and satellite imagery (Figure 4.2b).
The 0000 UTC 3 December 2000 analysis coincides with the next standard observation time for upper air soundings. At the surface, several suspicious surface observations from ship reports off the coast of Virginia exist where sea level pressure readings of 1020.5 and 1028.4 hPa are observed adjacent to one another (Figure 4.3a). The EDAS analysis of sea level pressure continues to indicate differences of 1 to 3 hPa to the west of the Appalachians with the 1026 hPa EDAS contour passing through an observation at Knoxville, Tennessee of 1024.8 hPa. Similar differences are identified at Greensboro, North Carolina, where 1029.4 hPa is observed but EDAS analyzed pressures between 1030 and 1032 hPa. At 850 hPa (Figure 4.3b), no major variations in the temperature or height fields could be identified and the winds appear representative of the EDAS analyzed pressure fields. Ridging at the 500-hPa is amplified in the EDAS analysis with differences of 10 m in FFC and BHM (Figure 4.3c). The inconsistencies continue through central Tennessee at BNA and into Ohio where the observations and the EDAS analysis exhibit additional differences. At Columbus, Ohio (CVG), the EDAS temperature is above –30 degrees C while actual observations indicate a reading of –31 degrees C. At the jet-level (Figure 4.3d), the position and strength of the jet streak over the lower Mississippi River Valley are not coincident with a wind of 120 knots observed at BHM. The EDAS analyzed winds of 120
knots appear to the northwest over northern Mississippi and central Arkansas. EDAS-analyzed winds at FFC, however, differ by at least 5 knots when compared to observational data. Comparison of EDAS pressure fields and wind directions inland indicate that the winds are well represented.

At the last observation time of 1200 UTC 3 December, similar differences exist near and west of the Appalachian Mountains with the EDAS sea level pressure analyzing a 1030 hPa contour passing again through Knoxville where the observed sea level pressure is 1028.5 hPa. Pressure variations over central North Carolina near Greensboro continue to be on the order of 1 hPa with EDAS too high compared to observations (Figure 4.4a). The placement and intensity of the coastal cyclone is reasonable with a slight deviation to the southwest based on the Buoy 41002 report of 1017.3 hPa and west-southwest winds of 10 knots east-northeast of Charleston, South Carolina. At 850-hPa (Figure 4.4b), the EDAS −10 degrees C contour isolated to southwest Virginia is not consistent with the −10 degrees C observation at BNA. Geopotential heights over the Carolinas and Georgia are observed to be 5 to 10 m lower than surrounding areas and winds at Charleston, Morehead City, Roanoke, and Atlanta, suggest that a closed low may exist over the region that is not analyzed in the EDAS. Differences on the order of
several meters are observed from BNA to ROA where the EDAS 1530 m contour passes through observations of 1526 m and 1528 m, respectively. At 500-hPa, the slight variations in the northern portions of Georgia and Alabama continue with EDAS analysis indicating observed geopotential heights may differ by 10 m at FFC (Figure 4.4c). The closed low over eastern Tennessee in the EDAS analysis also appears questionable with the northwest winds and approximate 10 m height differences at CVG, and in the South, not supportive of a closed circulation center. The isotachs at 250 hPa are relatively symmetrical and parallel to the geopotential height contours (Figure 4.4d). Variations in the wind speeds in the EDAS of 10 to 20 knots exist at FFC and BHM compared to observations. Otherwise, the EDAS is representative of the general synoptic conditions.

Taking into account the observed differences between the EDAS and observations at these pressure levels, the EDAS analysis will be used as a general representation of the synoptic environment and to describe the evolution of the 2-4 December 2000 snowstorm over central and eastern North Carolina.
4.2 Synoptic Overview

At the surface (Figure 4.5), the center of a 1040-hPa anticyclone moved south into the northern Plains states on 2 December 2000. Despite the location of the high-pressure system, pronounced ridging over the Northeast and Great Lakes regions established geostrophic northeasterly flow over the Mid-Atlantic coastline. West of the Appalachians, the remnants of a dissipating low were evident as an inverted trough that shifted east from the lower Mississippi River Valley to the Tennessee River Valley from 1200 UTC 2 December through 1200 UTC on 3 December (Figure 4.5b). To the east of the Appalachians, the initial stages of cold air damming were already evident at 1200 UTC 2 December as the inverted ridge signature in sea level isobars extending from central Pennsylvania southward into northeastern Georgia. The development of a coastal trough coincided with the evolution of cold air damming east of the Appalachians with a weak 1016-hPa cyclone analyzed to the east of Charleston, South Carolina, by 1200 UTC 3 December 2000. The coastal cyclone moved generally northeastward during the event from approximately 150 kilometers southeast of Savannah, Georgia, at 1200 UTC 2 December to well offshore of Cape Hatteras, North Carolina, by 0000 UTC 4 December. By this time, the coastal trough had lost its identity to the expanding circulation of the
cyclone as pressures fell to 1014 hPa.

At the 850-hPa level, light westerly winds over central North Carolina at 1200 UTC 2 December shifted to the northeast then east by 0000 UTC 3 December (Figure 4.6). The winds gradually increased in intensity through 0000 UTC 4 December when the maximum of 20 to 30 knots were observed. Cold advection, provided by the low-level northeasterly winds over central North Carolina, forced the 0 degree C isotherm farther south as time progressed. At 1200 UTC 2 December, the 0 degree C isotherm stretched from near Cape Hatteras westward to Asheville in North Carolina. By 0000 UTC 4 December, the freezing line had progressed southeastward to a position across Georgia and South Carolina. The 850-hPa low over central Tennessee slowly weakened as it moved southeast into central South Carolina on 3 December before redeveloping offshore on 4 December in association with the developing surface cyclone. Warm advection and moisture transport into North Carolina were limited to the hours surrounding 1200 UTC on 3 December that coincided with closest approach of the surface cyclone. Between 1200 UTC 3 December and 0000 UTC 4 December, low-level flow became more northerly and inhibited onshore flow and isentropic lift by reducing moisture and thermal transport into central North Carolina. Cold advection by north-northeast flow continued to support low-level stability and limit upward vertical motion.
In Figure 4.7, a positively tilted trough extended from the Canadian Maritimes southwestward to the Upper Midwest region at 1200 UTC 2 December. Confluent flow to the east of the trough persisted through 1200 UTC 3 December over New England before shifting offshore into the northwest Atlantic Ocean by 0000 UTC 4 December. A lead shortwave trough associated with a vorticity maximum over Tennessee at 1200 UTC 2 December passed across central North Carolina around 0000 UTC 3 December. The main lobe of cyclonic vorticity intensified briefly during the 12 hours preceding 1200 UTC 3 December, but sheared as the disturbance moved over and east of the Appalachian Mountains. The main trough continued to move eastward through 0000 UTC 4 December and the trough axis was aligned north-to-south over eastern North Carolina by this time.

At the 250-hPa level (Figure 4.8), the polar and subtropical jets were both evident in the eastern half of the United States throughout the period. The subtropical jet extended from the southern Gulf of Mexico across Florida and northeast into the western Atlantic. The polar jet was positioned to the south and downstream of the trough over the Great Lakes region. A 120-knot polar jet streak over the Central Plains moved east as the trough progressed southward into the Tennessee Valley from 1200 UTC 2 December through 0000 UTC 3 December. The amplitude of the trough decreased after this time as a new cyclonic vortex develops
near the Great Lakes and no significant wind maximum enters the western side of the trough to continue the amplification process. The trough also lifts northward during the 12 hours preceding 0000 UTC 4 December. This results in westerly flow over North Carolina, impeding the development of precipitation over the region.

At 1200 UTC 2 December (Figure 4.9), three distinct areas of precipitation were observed across the eastern United States including scattered showers in the far western Atlantic Ocean off the Southeast Coast, light to moderate precipitation in Georgia and Upstate South Carolina, and a broad east-west oriented band across the Ohio River Valley. The evolution of each individual area is considered separately due to the hypothesized impact of the related synoptic features on the quantitative precipitation forecasts. The offshore precipitation was aligned with the position of the coastal trough (Figure 4.10d) where 850-hPa warm advection developed (Figures 4.6b and 4.6c) over the cold dome immediately inland over the Carolinas. The second region of precipitation over southwestern North Carolina was displaced from the coastal trough and developing coastal cyclone, and was southeast of the cyclonic vorticity advection at 500 hPa ahead of the lead vorticity maximum. Cyclonic flow at the 850-hPa level over eastern Tennessee supported warm advection over the cold dome as southwesterly flow developed in
the lower troposphere. By 0000 UTC 3 December, the precipitation shield over the Ohio River Valley became better organized and intensified as a closed low formed over Tennessee at the 500-hPa level. As the precipitation approached western North Carolina, a rapid decrease in coverage and intensity was observed as energy transfer to the coastal low occurred. The main 500-hPa vorticity maximum crossed the Appalachians around 1200 UTC 3 December and sheared east-to-west over central North Carolina. As the transfer of energy occurred, a new region of precipitation developed westward over parts of central and eastern North Carolina. The precipitation then slowly shifted to the east and northeast through 0000 UTC 4 December with lingering effects only along the Outer Banks of North Carolina.

4.3 Atmospheric Response to 500 hPa Troughs

The initial disturbance crossed central North Carolina around 0000 UTC 3 December 2000 (Figure 4.7) with cyclonic vorticity advection at 500-hPa assisting in the generation of precipitation across the region prior to that time. Cyclonic (anti-cyclonic) vorticity advection at the 500-hPa level implies regions of upward (downward) vertical motion and, therefore, precipitation processes and thermodynamic response within the damming region. Prior to the arrival of the main vorticity maximum,
dynamically-driven atmospheric changes over central North Carolina may have contributed to the evolution and enhancement of topographically-induced flow over North Carolina and, hence, limited the precipitation associated with the second trough on 3 December 2000.

As precipitation fell into the cold, dry air at low-levels, evaporational cooling increased static stability over central North Carolina. Latent heat release due to condensation processes at cloud level may have further supported increased static stability within the damming region. Reinforced damming, therefore, continued to produce an environment conducive to subsidence, low-level cold advection, and transport of dry air into North Carolina. The strengthening cold air damming episode should result in an eastward and southward progression of the surface ridge and the attendant gradients of temperature and moisture to a location along the coast.

The following sections will evaluate the effects of the lead vorticity maximum on the atmosphere over central North Carolina. The initial disturbance at 500 hPa passed over North Carolina around 0000 UTC 3 December. The values for cyclonic vorticity advection reached a maximum over central North Carolina around 2100 UTC 2 December and moved east by 0600 UTC 3 December. To consider the effects of upper-level forcing separately from diabatic processes, the evaluation of the
event compares the initial analyses with each subsequent times to quantify the actual atmospheric response as the disturbance crosses North Carolina. Using the observational frame of reference, an evaluation of the model forecast skill provides an initial assessment of the meteorological factors potentially responsible for the precipitation errors in model guidance.

4.3.1 PRE-VORT: 1200 UTC 2 December – 2100 UTC 2 December

At 1200 UTC 2 December (Figure 4.10c), a shield of precipitation was entering southwestern North Carolina and extended into northwest South Carolina and northeastern Georgia. By 1500 UTC 2 December, the precipitation entered the Raleigh-Durham metropolitan area with the heaviest precipitation developing over eastern and central North Carolina between 1800 (Figure 4.11) and 2100 UTC (not shown). Cyclonic vorticity advection at 500-hPa increased across central North Carolina to a maximum between 1800 UTC 2 December and 0000 UTC on 3 December. Anti-cyclonic vorticity advection entered western North Carolina around the same time as the lead disturbance passed east of the mountains and over central North Carolina.

During the consideration period, surface pressures increased over central North Carolina by 1 to 3 hPa as temperatures remained relatively
steady or slowly increased through the day. For example, at the Raleigh-Durham International Airport (RDU), sea-level pressures rose from 1026.7 hPa to 1028.5 hPa ($\Delta p_{\text{pre}} = +1.8$ hPa) during the evaluation period (Figures 4.10 and 4.11). Temperatures in the east showed the most significant change with 5 to 10 degrees F rises evident along the coast where temperatures reached into the lower 40s. As precipitation began over central North Carolina, surface dewpoint temperatures climbed from the initial observations of upper 20s to lower 30s into the mid 30s to near 40 by 2100 UTC 2 December. Upstream in Virginia, however, low-level drying was reflected in lowering dewpoint temperatures on the order of 10 degrees F during the same time. Radar indicated little or no precipitation in Virginia associated with the lead vorticity maximum. The lack of precipitation processes in that region suggests that adiabatic processes resulting from upper-level forcing may have provided a more significant contribution to the increasing strength of the surface ridge rather than a diabatically-driven mechanism such as precipitation.

4.3.2 POST-VORT: 2100 UTC 2 December – 0600 UTC 3 December

As mentioned in the previous discussion, precipitation entered central North Carolina around 1500 UTC and was exiting to the east of
Greensboro, North Carolina, by 2100 UTC with the heaviest precipitation aligned roughly parallel to the coast between 50 and 100 kilometers inland over eastern sections of North Carolina (Figure 4.11). Temperatures prior to the precipitation and shortwave passage had risen several degrees over much of central North Carolina, except for locations near Greensboro where the effects of cold air damming initiated earlier than in Raleigh. Dewpoint temperatures exhibited a similar rise with the moistening low-levels of the atmosphere as the precipitation fell across the region.

In the wake of the lead vorticity maximum, however, temperatures fell 5 to 10 degrees with similar response in the dewpoint temperatures (Figures 4.11 and 4.12). Sea level pressures continued to rise following the passage of the upper-level disturbance and the progression of the precipitation shield into the coastal waters of North Carolina. Positive pressure changes across central North Carolina ranged from 1 to 2 hPa in the POST-VORT time frame with an observed change ($\Delta p_{\text{post}}$) at RDU of +0.6 hPa to 1029.1 hPa (Figure 4.12). By 2100 UTC 2 December, the first indications of development along the coastal baroclinic zone were noted with low pressure development offshore in response to the upper-level forcing.

As the main disturbance neared the Appalachians in western North
Carolina around 0600 UTC (Figure 4.12), the lead precipitation moved east and remained nearly stationary along the immediate coastline to approximately 50 kilometers inland. West of the Appalachians, the precipitation associated with the main vorticity maximum failed to progress over the barrier with only scattered showers developing over northwestern South Carolina and northern Georgia. Also in Figure 4.12, a small area of precipitation was developing near Greensboro, North Carolina, indicating the potential transfer of energy from the main upper-level feature to the coastal cyclone. Therefore, the 0600 UTC 3 December is used to evaluate the effects of the lead disturbance on the atmospheric environment over central North Carolina relative to the effects of the main 500-hPa trough. Anti-cyclonic vorticity advection across the northern half of North Carolina with cyclonic vorticity advection along the South Carolina-North Carolina border to the south indicated the role of upper-level forcing following the lead disturbance in maintaining and strengthening the cold air damming through adiabatic processes.

4.3.3 LEAD VORT: 1200 UTC 2 December – 0600 UTC 3 December

Total pressure changes ($\Delta p_{\text{total}}$) over North Carolina ranged from +2 to +6 hPa with observations of +2.4 hPa at RDU and +3.8 hPa at Greensboro during the 18-hour period (Figures 4.10 and 4.12).
Temperature changes of +/- 4 degrees F accompanied falls in the dewpoint temperatures of 5 to 15 degrees F. Precipitation ended around 0000 UTC 3 December in central North Carolina, however, cold air moved southward into both Carolinas with sub-freezing dewpoint temperatures as far south as Columbia, South Carolina, and as far eastward as Fayetteville and Rocky Mount in North Carolina. The contribution (C) of the upper level disturbance following the CVA to sea level pressure changes can be roughly computed using the ratio of $\Delta p_{post}$ to $\Delta p_{total}$ from the RDU observations in Equation 5, such that

$$C = \frac{\Delta p_{post}}{\Delta p_{total}} \times 100\% = \frac{0.6 \text{ hPa}}{2.4 \text{ hPa}} \times 100\% = 25\% \quad (5)$$

This would suggest that up to 25 percent of the atmospheric response in surface pressures can be attributed to adiabatic processes. Therefore, the role of diabatic processes associated with the lead precipitation must be investigated as a contributor to the model errors in quantitative precipitation forecasts over central North Carolina, especially given the recent research by Brennan and Lackmann (2005) for the 24-25 January 2000 case. Their research investigated the development of a potential vorticity maximum associated with initial precipitation southwest of North Carolina as well. Drying of surface air over Virginia as a result of convergence aloft over New England may be another key factor in limiting
precipitation in central North Carolina. As a result, a more thorough investigation of the role of upper-level forcing should be performed.

4.3.4 MAIN VORT: 0600 UTC 3 December – 0000 UTC 4 December

The small area of light precipitation over central North Carolina at 0600 UTC briefly expanded and intensified over central North Carolina; however, by 1200 UTC 3 December, the scattered showers began to dissipate from south to north across the Piedmont of North Carolina. The light precipitation that had developed in northwest South Carolina around the same time dissipated slowly and remained nearly stationary through 1800 UTC. Along the coast, the shield of precipitation began to progress westward along the Coastal Plain prior to 1200 UTC 3 December. At 1200 UTC, the westernmost edge of the precipitation shield reached the Raleigh-Durham metropolitan area before retreating to the east-northeast and establishing a mesoscale, banding feature along the same region where the maximum precipitation occurred. Cyclonic vorticity advection to the north and east of a line from Asheville to Wilmington, North Carolina (Figure 4.13) slowly progressed off to the east and became quasi-stationary over eastern North Carolina. Surface pressures continued to rise in central North Carolina through 1800 UTC with observed values of 1030 to 1034 hPa common in the region. As the low pressure along the
coast moved to the northeast, pressures began to fall as northerly winds imported dry, cold air from the Mid-Atlantic region. Dewpoint temperatures across the Carolinas and Virginia fell into the mid-teens to near 20 in regions where little or no precipitation occurred.

4.4 Model Performance Evaluation

The Eta FOUS data (Table 1.1) from 0000 UTC 2 December indicated that measurable precipitation was expected to begin at RDU between 0000 and 0600 UTC 3 December, with heavier precipitation (> 0.25”/6hr) not expected until after 0600 UTC 3 December. However, precipitation actually developed over central North Carolina between 1200 UTC and 1800 UTC 2 December, 6 to 18 hours prior to the model-generated precipitation. The initial precipitation over central North Carolina was associated with a lead shortwave trough that moved across the state ahead of the primary disturbance. With cold, dry air in place across central North Carolina, evaporational cooling produced additional support for the damming event that was underway.

As discussed in Section 4.1, the EDAS analysis used to initialize the Eta and MM5 models at 1200 UTC 2 December failed to accurately represent the lead disturbance over central Tennessee. As described in Sections 4.2 through 4.3, cyclonic vorticity advection associated with the
shortwave trough at the 500-hPa level enhanced lift over the cold dome and progressed in tandem with the initial precipitation shield across central portions of North Carolina. In addition, warm advection at 850 hPa was limited in the EDAS at 1200 UTC 2 December with southerly component winds in extreme western North Carolina, far southeastern North Carolina, and South Carolina (Figure 4.6b). Winds at 850 hPa were westerly across most of central and eastern North Carolina. The temperature gradient between FFC and BHM in Figure 4.1b was poorly identified in the initialization fields as well. In the wake of the disturbance, anticyclonic vorticity advection and confluence aloft assisted in the maintenance of the cold pool east of the Appalachians. It is hypothesized that limited moisture availability, subsidence, and cold advection in the damming region resulted in atmospheric conditions that were less favorable for the transfer of energy to the coastal trough and decreased the quantitative precipitation over central North Carolina in later periods.

Overall, the model forecasts for the event performed reasonably well with respect to the overall pattern; however, as seen in Figures 4.14 through 4.20, the Eta model did not represent the evolution of the lead shortwave trough at 500-hPa as well as the MM5. Figure 4.14 indicates the frontogenesis at 1000 hPa along the Southeast coast and provides
evidence that the shift in the coastal front seaward occurred between 0000 UTC and 0600 UTC 3 December, especially off the North Carolina coast, before strengthening at the immediate coastline by 1200 UTC 3 December. In addition, both models under-estimated the strength of the cold air damming event with errors in the Eta model compared to EDAS of -2 (Figure 4.15a) to -6 hPa (Figure 4.15d). At the coast, the cyclone tracked farther to the north and west of the EDAS analyzed storm with differences of -4 to -8 hPa observed by 1200 UTC 3 December along the coastline of North Carolina (Figure 4.15d). The MM5 exhibited similar errors with inland sea level pressure differences of -2 hPa (Figure 4.16a) to –6 hPa (Figure 4.16c), and coastal cyclone intensity errors up to –4 hPa as shown in Figure 4.16d. Both models over-predicted the strength of the coastal cyclone and timing issues exist between observations and model forecasts of precipitation and associated East Coast cyclogenesis. Comparison of the Eta and MM5 model forecasts for the lead disturbance over central North Carolina at 0000 UTC 3 December (Figure 4.17) shows the vorticity maximum of $20 \times 10^{-4} \text{ s}^{-1}$ over central North Carolina is under-predicted. The Eta model forecasts predicted values less than $12 \times 10^{-4} \text{ s}^{-1}$ while the MM5 guidance provided values of $12 \times 10^{-4} \text{ s}^{-1}$ to $16 \times 10^{-4} \text{ s}^{-1}$ in the region. The Eta and MM5 model forecasts valid six hours later (Figures 4.18a and 4.18b) fail to indicate the sheared vorticity lobe
over central North Carolina in EDAS (Figure 4.18c).

The model forecasts of coastal front location are compared to Figure 4.14. The increased resolution of the MM5 provides details not evident in either the EDAS or Eta model forecasts (Figures 4.19 and 4.20). At 0000 UTC 3 December, the 1000hPa frontogenesis in the Eta model forecast (Figure 4.19a) is immediately adjacent to the coast and shifts slightly offshore over the South Carolina coastal waters through 0600 UTC (Figure 4.19b). The MM5 model forecasts seen in Figures 4.19c and 4.19d indicate a more eastward position than the Eta and an enhanced shift to the east between 0000 UTC and 0600 UTC 3 December. The track of the cyclone can be examined through the plots of 1000 hPa frontogenesis in the 24- and 30-hour forecasts from the Eta and MM5 in Figure 4.16. With a position farther north and east than either the Eta forecasts (Figure 4.20a) or EDAS analysis (Figure 4.14d), the 24-hour MM5 model forecast (Figure 4.20c) for frontogenesis indicates a significant difference between the model forecasts and analysis. Remember from Section 4.1, however, that the displacement of the low pressure center in the EDAS was farther to the southwest with respect to the observed wind pattern and sea level pressure fields.

The following section will provide support for the main hypothesis through discussion of several additional suggestions for the source of
model errors in QPF including missing upper air soundings near and within the CAD region, model parameterization, and under-estimation of the CAD. The final section supports the hypothesis that the misrepresentation of a lead vorticity maximum at 500-hPa in the model initialization led to errors in the diagnosis of the cold air damming episode, coastal front, coastal cyclogenesis, and associated precipitation.

4.5 Proposal of Main Hypothesis

Through the assimilation of meteorological data from surface observations, upper air soundings, in situ data, and other analysis fields, mesoscale models calculate atmospheric variables on standard pressure level surfaces on the given forecast grid to represent the initial state of the atmosphere. The use of operational model data in the nudging of forecasts assists in quality control during the first several hours of integration by limiting error growth. The Eta Data Assimilation System was implemented in 1996 and provided a higher resolution objective analysis specifically designed to initialize the Eta model (Rogers et al. 1996). The EDAS used the Optimum Interpolation (OI) analysis until the spring of 1998 when the OI was replaced by the 3-dimensional variational scheme (3DVAR) (Rogers et al. 2000). The EDAS continually performs analysis steps using the 3DVAR at 3-hour intervals. By using a self-
cycling routine, the EDAS replaced the coarse-resolution Global Data Assimilation System (GDAS) in favor of consistent physics and improved horizontal and vertical resolution.

Model errors can be described as systematic or random. The total error associated with a model forecast is the sum of these errors. Systematic errors are repetitive and are caused by factors specific to a model, such as resolution or physical parameterization and can be accounted for subjectively and even corrected through model development. The random errors are generated by the inability to measure the atmosphere completely and the limitations of computational origin due to the truncation of dynamic and physical equations used in the model. Random errors are unpredictable and result from integration of the errors with time such as initialization fields that are incorrect. The rate of growth of these errors is many times dependent on the physical design of the model as well through the equations used to represent the physical and dynamic processes active in the model atmosphere.

In December 2000, the Eta and MM5 used the same initial conditions from the EDAS. Therefore, the differences between model forecasts would be due to the model design properties such as physical parameterization, dynamical processes, and resolution. Nonetheless, the accuracy and availability of these data are often limited by spatial
resolution and temporal distribution frequency. The development of cold
air damming, a coastal trough, and cyclonic weather systems along the
coast of the Carolinas and Virginia are often associated with, and driven
by, mesoscale variations in the horizontal and vertical distribution of
temperature and moisture (Fristch et al. 1992, Bosart and Lin 1984, Kuo
1987) throughout the depth of the atmosphere. Therefore, the
initialization of models used to forecast these events has been a priority
of National Weather Service offices across the Southeast (Gurka et al.
over central North Carolina suggests that models predicted excess
precipitation in a region where atmospheric conditions failed to support
the development of precipitation. For example, the Eta model is capable
of producing ice crystal growth and precipitation generation in grid
volumes that do not necessarily meet the environmental temperature
thresholds, so long as the prior time-step grid volume or vertically
adjacent grid is cold enough to support ice nucleation (Janjic 1994).
Further discussion of the microphysical considerations associated with this
process is provided in Chapter 5 as part of the sensitivity studies.

The track and intensity of low pressure along the coast has
previously been correlated in a conceptual model from the National
Weather Service in Raleigh with the western extent of the precipitation shield into eastern North Carolina. The conceptual model is based on climatology and the critical influence the track has on the influx of moisture and upward vertical motion associated with the baroclinic environment in the vicinity of the Gulf Stream. A track farther to the east and north, as presented in the MM5 model output (Figures 4.15 and 4.16), would restrict the instability and moisture available to generate precipitation as far west as the Raleigh-Durham area.

As will be evidenced in Chapter 5, the low-level drying in the MM5 model exceeded that in the Eta model forecasts over central North Carolina. Therefore, the moisture and thermal gradients in the vicinity of the coastal front and Gulf Stream are also shifted farther offshore and, therefore, would explain a portion of the deviations of the cyclone location in the MM5 versus the Eta model. Figures 4.14, 4.19, and 4.20 reflect this in a comparison of the EDAS analysis with the Eta and MM5 model forecasts valid at times from 1200 UTC 2 December through 1800 UTC 3 December. As with the model biases, the eastern shift in the position of the CAD boundary, attendant coastal front, and incipient cyclogenesis would limit the precipitation in eastern North Carolina through reduction in the available factors conducive to precipitation generation as discussed in the WM00 ingredients-based methods in
Chapter 2.

The main hypothesis of this study is that the MM5 represented the evolution of the lead 500-hPa trough better than the Eta model, resulting in improved representation of the CAD, coastal front, coastal cyclone, and precipitation. To evaluate each model’s tendency to produce precipitation, a quasi-geostrophic analysis of the atmospheric properties over central and eastern North Carolina will identify if the Eta model produced precipitation in a region that would typically limit precipitation production and whether the MM5 model properly represented the actual conditions observed over the region.
5. RESULTS

5.1 Model Sensitivity Experiments

The methods for testing the main hypothesis through sensitivity experiments with the MM5 model should provide evidence that the SST had little influence on the MM5 model-generated precipitation and topographic flow regimes through the cumulus, planetary boundary layer, or microphysics schemes. In Section 5.2, the quasi-geostrophic diagnosis of winter season precipitation using the basic principles of WM00 will also be presented to determine the likelihood that the Eta produced snowfall in an atmosphere non-conducive to precipitation development using cross-sectional analyses.

Prior sensitivity experiments conducted using the MM5 model suggest the importance of proper physical parameterization schemes for optimal performance (Zhang et al 1994). Parameterization schemes in mesoscale modeling systems represent the physical and dynamic processes responsible for the generation, dissipation, transport, and distribution of heat, moisture, and mass quantities throughout the depth of the atmosphere at scales smaller than the model grid resolution (Cortinas, Jr. and Stensrud 1995). Sensitivity experiments using a real-time, local version of the MM5 model were performed with changes to the cumulus parameterization, precipitation microphysics, and planetary
boundary layer schemes (Table 5.1). In the upcoming sections, the methodology for the several sensitivity tests is outlined using a general conceptual model and discussion of the results with respect to the model physics options that influence the synoptic pattern and precipitation fields.

5.1.1 Cumulus Parameterization Schemes: KF vs. BM

The impact of the cumulus scheme on sensible weather and precipitation forecasts has been thoroughly examined in prior studies using mesoscale models (Seaman 1999, Wang and Seaman 1999, Zhang et al 1994, Kuo et al 1996). Kuo et al (1996) found that model temperatures, the evolution of mesoscale low pressure centers, and precipitation were extremely sensitive to the CPS.

The MM5 system offers the Betts-Miller (BM) (Grell et al. 1993, Grell et al. 1994) scheme as a convection parameterization option that is nearly identical to the Betts-Miller-Janjic scheme (Janjic 1994) of the operational Eta model. If the convective precipitation associated with the coastal low led to early cyclogenesis and a track farther offshore in the MM5, the experiment with the BM instead of the Kain-Fritsch scheme should generate similar results to the Eta model precipitation forecasts across North Carolina. The MM5 sensitivity experiment did reveal
differences of up to 4 hPa in the intensity of the coastal low with the BM scheme (Figure 5.1); however, differences in the position of the cyclone and associated precipitation shield were barely remarkable with only the sub-grid scale convective precipitation totals of particular note (Figures 5.2 through 5.5). The MM5 run did not exhibit the western bias, or a definite western shift in the precipitation exhibited by the Eta model guidance.

5.1.2 Planetary Boundary Layer Scheme: BLK vs. MRF

The generation of excessive precipitation over central North Carolina was most prolific during the six hours preceding forecast hours 24 and 30 as the storm moved northeast along the North Carolina coast to the east of Cape Hatteras (Figure 1.2). Between 1200 and 1800 UTC 3 December, drying throughout the depth of the atmosphere near Raleigh caused the precipitation to dissipate and move east along the I-95 corridor. The Eta model forecasts from the 1200 UTC run show a strong easterly jet near the 850-hPa level that transports the warm, moist air over the Gulf Stream into central North Carolina over the wedge of cold air at the surface. The enhanced lower tropospheric flow in the Eta appears to originate as a diabatically-forced process near the surface that leads to isentropic transport of the thermodynamic instability at low- and
mid-levels across central North Carolina. Actual examination of these features will be incorporated in the three-dimensional analysis performed in the next section. Therefore, some additional consideration regarding the physical design of the MM5 requires investigation of the planetary boundary layer (PBL) scheme that involves turbulent mixing between the surface layer and vertically throughout the depth of the modeled atmosphere. At the surface, the PBL scheme allows the atmosphere to interact with the surface layer by transporting the quantities of moisture and heat through turbulent eddies to adjacent and/or subsequent layers above the surface. As a sensitivity experiment, the Blackadar and MRF schemes were tested. The results show little difference in either the coastal cyclone (Figure 5.1) or the precipitation (Figures 5.2 through 5.5), except for mesoscale variations in the sub-grid scale precipitation.

5.1.3 Microphysical Parameterization: SI vs. R2

The microphysical scheme determines the evolution of atmospheric water within the clouds generated through the CPS. The release and absorption of heat through phase changes is particularly important in determining the atmospheric stability used by the CPS to generate vertical motion. The final MM5 sensitivity experiment that intends to reproduce the forecast errors exhibited in the Eta forecasts involves the
microphysical parameterization scheme. As will be discussed in Section 5.2, additional heat and moisture are available in the Eta relative to the MM5 at forecast hour 12. Relative humidity was less than 40 percent above 500-hPa and limited negative values of pressure vertical velocity exist in the 950- to 850-hPa layer. Since the Eta model has shown a prior bias to forecast snow in regions without the atmospheric environment required for ice crystal growth, the sensitivities using MM5 model microphysics compared the Dudhia simple ice scheme and the Reisner-2 scheme that includes graupel and super-cooled liquid water. The seeder-feeder effect in the model atmosphere may have contributed to the heavy snowfall produced by the Eta due to the consideration of the environmental temperature in the adjacent grid volume above the level of saturation to determine the precipitation type. However, the MM5 runs again fail to provide a major difference in precipitation forecasts over central North Carolina with the change in physical parameterization (Figures 5.2 through 5.5).

5.2 Quasi-geostrophic Analysis

In the case study from Chapter 4, the displacement of the coastal trough to the east was correlated with the passage of the 500-hPa vorticity maximum. The adiabatic and diabatic processes associated with
the initial precipitation lend credence to the fact that the 500-hPa lead trough contributed to the model error. Using a cross-section (Figure 5.6) from Buoy 41001, east of Cape Hatteras (HAT) westward across the Appalachian Mountains to London, Kentucky, the vertical motion will be diagnosed using quasi-geostrophic diagnostics such as temperature advection and vorticity advection, as suggested by WM00. Lastly, the vertical temperature and moisture profiles over central North Carolina will be examined to determine if the Eta may have produced ice crystal growth in an environment with insufficient moisture and/or cloud top temperatures of −10 C or higher.

The increased surface pressures and lowering dewpoint temperatures associated with the lead disturbance substantiate the rapid development of cold air damming across central North Carolina, particularly following 1200 UTC 2 December when precipitation initially over-spread the region. Dewpoint temperatures plummeted some 15 to 20 F between 2100 UTC 2 December and 0000 UTC 4 December 2000 as winds transported cold, dry air from the Mid-Atlantic region into the Carolinas. As discussed in Chapter 4, forecasts from the Eta and MM5 models under-estimated the strength of the surface ridge throughout the event. Although the errors of −1 to −3 hPa were consistent between both models through forecast hour 18, the magnitude of model errors
increased with time and diverged between the models through forecast hour 30. The Eta model exhibited the largest errors of −5 to −10 hPa over central North Carolina compared to the −4 to −6 hPa exhibited in the MM5 model forecasts. In the following sections, the evolution of the precipitation errors relative to the lead trough are examined using the cross-sectional analyses and forecasts of the atmosphere at forecast hours 12, 18, 24, and 30. The distribution of vertical forcing, moisture, and temperature throughout the atmosphere will be examined using the WM00 methodology from Chapter 2 as a basis for determining the predisposition of the atmosphere to precipitation generation and the source of model errors in quantitative precipitation forecasts.

5.2.1 Cross-Sectional Analysis

To investigate the role of upper-level forcing in the misdiagnosis of precipitation across eastern and central North Carolina during the event, comparisons of the cross-sectional analyses and forecasts of the pressure vertical velocity, \( \omega \), are presented in Figures 5.7 through 5.10 to determine the location, magnitude, and direction, of vertical motion across the region with respect to time. Since the primary errors in precipitation occurred during the 12 to 30 hour period following 1200 UTC 2 December 2000, differences between the model guidance and EDAS
analysis at 12, 18, 24, and 30- hours will illuminate the regions of primary concern and provide a trace for the source region of any significant differences apparent between models or with the actual observed conditions.

In Figure 5.7, the EDAS analysis at 0000 UTC 3 December indicated subsidence across much of North Carolina at low-levels with positive values of $\omega$ extending from near 700 hPa to the surface over central North Carolina and weakly negative $\omega$ along the eastern escarpment of the Appalachians and offshore. Between 700 and 500 hPa, the region of maximum upward vertical motion ($\omega < -6 \mu$b/s) was positioned over eastern North Carolina with weak forcing over western and extreme eastern North Carolina above 500 hPa. A secondary maximum ($\omega < -4 \mu$b/s) was located offshore near Buoy 41001 below 850 hPa. The 12-hour forecasts from the Eta and MM5 model generally predicted the overall circulation patterns evident in the EDAS analysis with moderate to strong ($\omega < -10 \mu$b/s) upward motion evident between the 500- and 700-hPa levels over eastern North Carolina (Figure 5.7). Although both models show a more pronounced maximum than in the EDAS analysis, the higher values in the MM5 relative to the Eta are likely attributed to the increased resolution (15- vs. 40-km) in the MM5 model. Similarly, both models over-predict the vertical motion offshore with the MM5 model exhibiting
the largest errors as convective activity is evident near Buoy 41001.

By 0600 UTC 3 December, the EDAS cross-sections indicate the upward vertical velocities (UVV) had increased ($\omega < -8 \mu \text{b/s}$) over eastern North Carolina and shifted to near the 700-hPa level farther east near the coastline (Figure 5.8). A vertical extension of the $-4 \mu \text{b/s}$ contour into the 500-250 hPa layer above the Rocky Mount location was also noted in the EDAS analysis at this time. This feature is of particular note because of its absence in the 18-hour Eta forecast guidance. Instead in the Eta model, a secondary region of maximum UVV with $\omega$ values of $< -12 \mu \text{b/s}$ exists between 700 and 500 hPa and extends much farther west above the western Piedmont of North Carolina. At low-levels, weak subsidence is evident in the EDAS analysis across all of North Carolina and the adjacent coastal waters with the exception of near neutral values at the land-water interface along the coast. The boundary between the subsidence over eastern North Carolina and the convection offshore was displaced west of observations in both model forecasts, however, the Eta model at 18-hours already exhibited a western bias compared to the MM5. The MM5 again over-predicted the convective activity in the vicinity of Buoy 41001 compared to both EDAS and Eta model forecasts but showed little horizontal displacement in the dual maxima offshore.

The trend to confine the UVV near 700-hPa to the east of central
North Carolina continued in the EDAS analysis at 1200 UTC 3 December with the maximum observed values of −12 to −14 µb/s just east of the Rocky Mount location in eastern North Carolina (Figure 5.9). A secondary maximum just below 700 hPa also developed just offshore. Both regions of moderate to strong UVV at 700 hPa extended vertically to near 500 hPa. The 24-hour forecast from the Eta model significantly over-produced vertical motion in western and central North Carolina with values of $\omega$ less than −20 µb/s between 700 and 500 hPa concentrated over the Raleigh-Durham metropolitan area. The consolidation of the primary vertical motion east of the Raleigh area in the MM5 model guidance continued at 1200 UTC 3 December; however, the extension of moderate UVV ($\omega < -8 \mu b/s$) to upper-levels across the western portion of the state was slightly more pronounced than in the EDAS. Again, this reflection of higher values may be attributed to the increased resolution. Subsidence below 850-hPa in central North Carolina is less pronounced in the Eta than in MM5 model guidance at 24-hours. The region of extremely weak downward forcing ($-2 \mu b/s < \omega < 0 \mu b/s$) in the Eta forecasts is immediately adjacent to the mountains, whereas MM5 model forecasts predict the extension of the low-level subsidence as far to the east as Rocky Mount in central North Carolina.

At the end of the consideration period (Figure 5.10), 1800 UTC 3
December, the EDAS indicated the strongest vertical motion just offshore and just east of Rocky Mount between 850 and 700 hPa with the maximum $\omega$ values of −16 to −18 µb/s along the coast. The MM5 model generated several banding features evident in the pressure vertical velocity profiles across eastern North Carolina from Raleigh eastward to the coast. The UVV over the Raleigh areas, however, existed only above 700 hPa. The 30-hour forecasts from the Eta model showed limited UVV over the Rocky Mount area and significant values of $\omega$ less than −40 µb/s over central North Carolina near 700 hPa. Little vertical motion of distinction was notable in the EDAS analysis valid at 1800 UTC 3 December to the west of Raleigh from the surface to 250 hPa. The MM5 model did maintain the low-level subsidence from the mountains east to near Rocky Mount while the Eta exhibited positive values for $\omega$ only immediately at the Appalachian escarpment.

The determination of regions highly conducive to generate precipitation is limited to the locations where upward vertical velocities coincide with relative humidity (RH) values in excess of 80 percent. It is important to remember, however, that the locations with increased relative humidity are not an absolute measure of atmospheric moisture content. Instead, the relative humidity represents the degree of saturation that is forced by adiabatic lift and cooling associated with cloud
development and precipitation processes. In section 5.2.2 through 5.2.4, the thermodynamic profiles of moisture and temperatures are considered to indicate the availability of moisture.

At 0000 UTC 3 December, relative humidity values in excess of 80 percent were observed in the EDAS analysis in Figure 5.7 from the surface to 700 hPa across the most of the cross-section with the exception of areas below 850 hPa within the damming region of central and eastern North Carolina where relative humidity was generally 70 to 80 percent. From near Rocky Mount westward to the Appalachian Mountains, values for relative humidity less than 30 percent above 500 hPa showed significantly dry air in the wake of the upper level disturbance located over central North Carolina at the time. The 12-hour model forecasts from the Eta and MM5 predicted the region of substantial relative humidity relatively well. The Eta model did not represent the drier air above the 700-hPa level above the Appalachian Mountains nor the magnitude of atmospheric saturation to the 500-hPa level offshore in the vicinity of the developing low pressure area.

The drier air between 700 and 500 hPa continued to progress east through 0600 UTC 3 December (Figure 5.8) with a notable intrusion of relative humidity less than 80 percent to the east of the Appalachians and west of Rocky Mount. The Eta and MM5 both represented this area of
decreasing relative humidity with height over the Appalachian Mountains; however, the Eta model continued to over-predict the relative humidity in the 700-500 hPa layer near the Raleigh-Durham metropolitan area, an indication that model cloud and precipitation processes were active in a region where heavy precipitation was not occurring. The secondary UVV maximum ($\omega < -6 \mu$b) in the MM5 model is also outside the region of maximum relative humidity, whereas, the Eta model indicated that relative humidity values in the westward-displaced UVV maximum ($\omega < -12 \mu$b) were conducive to precipitation generation across central North Carolina. The relative humidity below 850 hPa in central North Carolina decreased in the EDAS analysis during the 6-hours preceding 0600 UTC 3 December with values now less than 50 percent near 925 hPa. The MM5 represented the low-level drying similar to the actual observations (Figure 5.8). Eta 18-hour forecasts, however, did not represent the magnitude of drying with relative humidity still near 80 percent.

Figure 5.9 shows high relative humidity over central North Carolina existed in the EDAS analysis only in the 850-700 hPa layer over central North Carolina and from the surface to near 700 hPa over eastern North Carolina at 1200 UTC 3 December. Above 700 hPa, dry air was exclusive across the entire cross section with relative humidity values below 50 percent to a minimum of less than 10 percent near 400 hPa. Moderate to
strong UVV coincided with relative humidity over 80 percent in only a shallow layer near 700 hPa from Rocky Mount westward approximately 50 kilometers. The Eta model forecasts indicated adiabatic lift and moistening of the column up to the 500 hPa level over central North Carolina with values of 80 percent, a difference from EDAS of nearly 50 percent. Strong vertical ascent (\( \omega < -20 \mu b \)) in the Eta model forecasts coincided with the over-predicted relative humidity values within a layer from 850 to near 500 hPa. The Eta produced a region of coincident UVV and relative humidity greater than 80 percent from the Foothills eastward to the coast within the 850-700 hPa layer. Although MM5 over-predicted low-level drying adjacent to the mountains, the extension of substantial relative humidity into regions of UVV was limited to the area approximately 20 to 30 kilometers west of Raleigh within a shallow layer of, perhaps, 100 hPa near the 700-hPa level (Figure 5.9). The secondary maximum in UVV (\( \omega < -6 \mu b \)) over the Appalachian Mountains was in a region of relative humidity values less than 30 percent, the same region where MM5 actually exhibited a bias of drier air compared to the Eta model. The Eta model did not predict the correct magnitude of low-level drying with relative humidity values over 70 percent from the surface to near 500 hPa.

Unfortunately, the relative humidity values at 1800 UTC 3
December were unavailable in the EDAS analysis valid at the same time. Therefore, only the evaluation of differences between the model forecasts was available and is presented in Figure 5.10. The vertical motion over central North Carolina continued to enhance the saturation of the vertical column through 1800 UTC 3 December in the Eta model forecasts at 30 hours with sufficient moisture for precipitation to the 500 hPa level across the entire region. Over the Appalachians, forecasts of significant drying in the MM5 and Eta model forecasts in the wake of the main vorticity maximum had begun below 500 hPa. Relative humidity values across eastern portions of central North Carolina increased in the MM5 forecast guidance, particularly in the 700-500 hPa layer to the west of Raleigh in the westernmost banding feature seen in the UVV profiles.

5.2.2 Skew-T Analysis at RDU

To evaluate the thermal contributions to the precipitation processes during the 2 – 4 December event, the thermodynamic (skew-T) diagrams at Raleigh-Durham (Figures 5.11 through 5.14) are examined to determine the stability, precipitation efficiency, and precipitation-type expected across central North Carolina relative to the regions of high likelihood outlined in section 5.2.1. The EDAS analysis is used to generate thermodynamic profiles of dewpoint, temperature, and wind fields with
height at the Raleigh-Durham location due to the lack of actual upper air observations at the location. Following the analysis Similar to the cross-sectional analyses, the skew-T diagrams from model forecast guidance will be compared at the interval times of 12, 18, 24, and 30 hours, as well.

Atmospheric stability is determined using the skew-T plots from Raleigh-Durham and seeking regions of dramatic differences between the EDAS thermodynamic profiles and the model forecast guidance. To avoid significant repetition in the contributions of moisture and temperature to the generation of precipitation, the following sections examine the thermodynamic profiles with respect to precipitation efficiency, precipitation type, and stability. The results of the cross-sectional evaluations and skew-T diagrams are considered in reference to the main hypothesis that the 500-hPa trough influenced the development of early precipitation in central North Carolina and limited later precipitation development through contributions to the topographically-forced flow over the Southeast.

Examination of the precipitation efficiency across central North Carolina utilized the EDAS analysis for upper air soundings at the Raleigh-Durham International Airport at six-hour intervals from forecast hour 12 through 30. Regions of likely precipitation were considered in locations
where the cloud top temperature was less than -10 \degree C. This is consistent with the methodology of WM00 where the temperature range for maximum ice crystal growth is considered to be between -12 and -15 \degree C.

In the EDAS analysis and Eta model forecast valid at 0000 UTC 3 December 2000 (Figure 5.11), the atmosphere appears to be saturated with respect to ice throughout most of the lower and middle troposphere, especially from near 850 to near 600 hPa. With similar profiles, some cloud development appears possible around 700-hPa, where cloud top temperatures approach -10 \degree C. In the MM5 model forecast, a saturated layer appears closer to the surface from 925 hPa to above 850 hPa. The cloud layer does not appear to extend vertically to the 700-hPa level and may indicate the development of low-level stratus above the cold dome. Temperatures in the cloud layer are much too warm for significant ice crystal growth at 0 to -5 \degree C in a nearly isothermal layer.

At forecast hour 18 (Figure 5.12), the models came in to much closer agreement with the EDAS analysis in the surface-850 hPa layer, although the MM5 model represented the low-level drying within the layer much better than the Eta model guidance. A deep layer of saturated air existed in the MM5 from the 850-hPa level to above 700 hPa. Cloud top temperatures were representative of an atmosphere conducive to precipitation generation with values near -15 \degree C. The Eta model
represented a cloud layer with cloud top temperatures of 0 to -5 C. The EDAS analysis at 0600 UTC 3 December showed a deeper cloud than the Eta model but less deep than the MM5 model. The cloud top temperatures also were near -10 C indicating the potential for precipitation.

By 1200 UTC 3 December, the EDAS analysis indicated a dual cloud layer with low-level clouds near 850 hPa and a second region of saturated air above the low-level stratus deck to near 700 hPa (Figure 5.13). Cloud top temperatures by this time no longer sustained ice crystal growth with values above -10 C. The Eta model 18-hour forecasts predicted a cloud layer across a similar vertical extent as the EDAS analysis; however, cloud top temperatures remained too warm for significant precipitation development in the region of saturation. The MM5 model continued the trend of extensive clouds from 925 hPa to 700 hPa. Cloud top temperatures in the MM5 also were warmer than -10 C but did match the values found in the Eta model at the same level near 700 hPa.

Although the 1800 UTC 3 December EDAS analysis was not available for consideration, comparison of the MM5 and Eta model forecasts was still considered in Figure 5.14. Remarkably, both models now exhibited the saturated layer from 925 to 700 hPa. The MM5 actually extended this layer into the 700-500 hPa level where cloud top temperatures approached -20 C. On the other end of the spectrum, the
Eta model exhibited warmer temperatures above -10 C. A nearly isothermal layer in the Eta model does suggest, however, that some snowfall may have occurred if significant evaporational cooling and melting of snowflakes continued. This is covered more succinctly in the following section on precipitation type.

As in the previous sections, chronological consideration of the model forecasts and EDAS analysis are examined using the thermodynamic profiles to determine the likely precipitation type over central North Carolina in the vicinity of Raleigh-Durham. An investigation of the Eta, MM5, and EDAS profiles indicate that the temperatures below cloud layer were below 0 C in the lowest levels and justified frozen and/or freezing precipitation reaching the surface.

5.2.3 Skew-T Analysis at RNK

Due to the potential discrepancies in the use of thermodynamic profiles generated at the RDU location by the EDAS analysis, the actual observed soundings from Blacksburg, Virginia (RNK), and Morehead City, North Carolina (MHX), are used to define the evolution of the atmosphere within the cold dome and near the coastal front, respectively. Model forecasts from the Eta and MM5 are compared to the observed soundings to evaluate the moisture and thermal properties of the atmosphere and
any potential contributing factors. Since the standard synoptic observations are taken at 12 hour intervals, the only observation times used in the analysis are 0000 UTC and 1200 UTC 3 December 2000.

At 0000 UTC 3 December (Figure 5.15), the MM5 and Eta model are similar although cooler environmental and dewpoint temperatures at the surface to 850-hPa are predicted in the MM5 forecasts. The drier air at low-levels supports the hypothesis that the MM5 model better predicted the strength and intensity of the cold air damming. Compared to the Eta model 12-hour forecasts, the MM5 model is much drier aloft between 500 and 250 hPa. The dry low-level atmosphere in the observed sounding from RNK provides evidence that MM5 better predicts the moisture availability in the surface to 850-hPa layer with similar dewpoint temperatures. Neither model is dry enough in the 850- to 700-hPa layer. The MM5 model forecasts for dewpoint temperature at 500 hPa were approximately −70 C, as were the observed readings from the 0000 UTC sounding. The Eta model, however, predicted values 10 C warmer at that level, indicating a positive bias in available moisture. The low-level jet is of particular concern as well with the MM5 and Eta model forecasts predicting winds from the east or east-northeast at 20 knots. The Eta model forecast indicated the easterly flow over a deeper layer than MM5 and observations that showed a 15 knot easterly wind at 850 hPa.
By 1200 UTC 3 December (Figure 5.16), the observed sounding from RNK indicated significant dry air at 850 hPa. The extremely low dewpoint temperature at 850 hPa in the observed sounding is likely a bad data point. Given the moist profile aloft, ice crystal growth is supported with atmospheric conditions conducive to precipitation. Given the likelihood that precipitation is occurring, the level near 850 hPa is more likely saturated. Increased moisture content of the atmosphere at 850 hPa in Eta model forecasts is indicated by dewpoint temperatures that are 25 to 30 C higher than the observation or MM5 model forecast. Perhaps, the Eta model better represents the actual conditions at RNK. The MM5 model, on the other hand, reproduces the bad data in observations and under-predicts the depth of the cloud layer at 700 hPa. Unlike the Eta model, dry air in the MM5 forecast does not support precipitation processes. Therefore, the dry air in the sounding is more likely a reflection of the lack of precipitation in the MM5 rather than a truly accurate representation of the atmospheric conditions for RNK at 1200 UTC 3 December. Dry air at low levels would have limited precipitation in the models through stabilization at the top of the planetary boundary layer. Again, the Eta 30 knot jet and MM5 25 knot jet were over-predicted at 850 hPa with observed values around 10 knots.
5.2.4 Skew-T Analysis at MHX

At 0000 UTC 3 December (Figure 5.17), the Eta model indicated moist instability near 850 hPa with an easterly jet of 30 knots near 850 hPa. No easterly flow was observed in the actual sounding at MHX. MM5 also indicated a weaker jet of 15 knots. The dry air at 500 hPa was overpredicted by both the Eta and MM5 models at MHX. Compared to the MM5 model, dry air existed at the surface in the Eta forecast sounding. This drying could be an indicator of the lack of initial precipitation in the Eta model.

By 1200 UTC 3 December (Figure 5.18), the Eta model was much too warm at low-levels with temperatures of nearly 10 C, no inversion, and isothermal profiles to 850 hPa. The 850-hPa jet was east at 45-50 knots in the Eta forecasts compared to a more northeasterly 30-40 knots in the MM5 and 25 to 30 knots from the east in the observations. Moisture availability in the Eta model was greater than the MM5 based on a comparison of dewpoint temperatures in the surface to 700 hPa later. Profiles aloft were similar between both model forecast soundings and observations.
6. CONCLUSION

6.1 Project Description

Operational numerical weather prediction models in the year 2000 exhibited a significant lack of skill in three independent East Coast snowstorms including a record-breaking storm on 24-25 January. Several case studies performed for the events proposed various sources of error including the common thread of errors in sea surface temperature (SST) analyses for each event. The role of SST fields in the development of East Coast snowstorms has been thoroughly investigated, particularly along the Southeast coastline where cold air damming events and coastal frontogenesis are commonplace during the winter season. The position and magnitude of the coastal front and associated cyclogenesis are influenced by the divergence aloft associated with the passage of upper-level disturbances.

For an eastern North Carolina snowstorm on 2-4 December 2000, the operational Eta model over-predicted the precipitation in central North Carolina on the order of 2 to 3 inches in liquid equivalent. A local version of the MM5 model, however, predicted the placement and quantity of precipitation much better than the Eta model forecasts from runs on 1200 UTC 2 December for the event. Since identical initial conditions and lateral boundary conditions were used in both the MM5 and Eta model
forecasts from that morning, the difference between the two model forecasts presents an interesting consideration to forecasters. Did the MM5 produce the right forecast for the wrong reason? One potential hypothesis suggested superior model physics in the MM5 led to an improved precipitation forecast relative to the Eta model. To investigate the source of model errors in QPF over central North Carolina, sensitivity studies and a quasi-geostrophic evaluation of model guidance were performed.

Model-generated precipitation is forced by the location of atmospheric moisture, divergence, thermal properties, and stability issues and can be diagnosed using a quasi-geostrophic framework. However, initially the model errors are diagnosed through a case study to identify the synoptic environment and to divulge the actual errors that existed in the initial conditions and in the EDAS analysis used for the synoptic overview. Sensitivity experiments with the MM5 physical parameterization schemes were used to determine if those features of the model design are a function of the precipitation errors due to the initial condition problems. The quasi-geostrophic analysis followed to discern the differences in atmospheric properties between the Eta and MM5 that may have led to errors in QPF, specifically the hindered development in the MM5 over central North Carolina and the over-prediction by the Eta.
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6.2 Major Findings

With respect to the initial conditions, a preliminary comparison of EDAS and observations suggests that the primary differences were focused from northern portions of Alabama and Georgia through Ohio with differences in the initial 500 hPa trough and 850 hPa thermal gradient. The east-west gradient of temperature at 850-hPa between FFC and BHM is not noted in the EDAS contours. Sea level pressures were also too high along and east of the Appalachian Mountains in the EDAS analysis compared to actual observations throughout the three considered time periods of 1200 UTC 2 December, 0000 UTC 3 December, and 1200 UTC 3 December. Problems over the mountainous regions could arise from a possible misrepresentation by the Eta model of terrain due to the use of step-mountain coordinates rather than the terrain-following sigma coordinate system.

The initial precipitation developed in a region of cyclonic vorticity advection over central North Carolina ahead of the lead disturbance and in response to warm advection at 850 hPa between 1200 and 1800 UTC 2 December. Sea level pressures rose across the region with the initial low-level drying indicated in central Virginia by 2100 UTC 2 December where dewpoint temperatures fell by nearly 10 °F. Radar indicated that no precipitation occurred across Virginia to enhance the cold air damming
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and that the strengthening low-level damming event was forced by adiabatic processes aloft as anticyclonic vorticity advection in the wake of the lead trough restricted the removal of mass from the atmospheric column.

The case study revealed that the Eta model forecasts from 0000 UTC 2 December developed precipitation too late over North Carolina with no indication of the early precipitation associated with the initial 500 hPa trough. Errors in the EDAS analysis at 1200 UTC 2 December were primarily in the 850 hPa temperature gradients between FFC and BHM and 500 hPa heights over the Southeast. Radar and satellite imagery confirmed the existence of the lead vorticity maximum over the region at 1200 UTC 2 December. In the model evaluation, the Eta and MM5 model both under-estimated the strength of the cold air damming. Evaluation of the 1000 hPa frontogenesis suggests that the Eta model also predicted the position of the coastal front farther to the west than the MM5 model along the coast. The MM5, however, predicted the coastal low position farther northeast than the Eta model. The displacement of the low in the EDAS analysis was observed to be southwest of the actual position based on a comparison to the ship and buoy reports at 1200 UTC 3 December. Therefore, the placement of the low pressure system by the MM5 is considered a better representation of actual conditions. Both models also
over-predicted the strength of the coastal low with largest errors in the Eta model of 8 hPa. The reflection of lower pressures in the MM5 model may be an artifact of the earlier development of the low pressure system in the MM5 model, although this was not examined in the present study. The sheared 500 hPa vorticity lobe seen in the EDAS analysis suggests that the initial disturbance decreased the potential for transfer of the upper level energy across the mountains to the coast by enhancing subsidence, decreasing the available moisture in the atmosphere, and stabilized the low-levels in the damming region.

Sensitivity experiments with the MM5 divulge the role of model physics in the development of the western edge of precipitation over North Carolina and the coastal low development. Results of the sensitivity studies indicated only mesoscale variability in the precipitation fields with no large scale synoptic differences in the surface fields. The western edge of the precipitation fields were basically unchanged between individual model runs with the MM5 cumulus parameterization, planetary boundary layer scheme, and microphysical scheme. These results indicate that the role of the SST fields on model-generated precipitation in the MM5 model was limited based on the fact that the distribution and magnitude of precipitation and the location and strength of the coastal low were remarkably similar. Positioning errors of the coastal low and 2 to 4 hPa
differences also provide evidence that the coastal cyclone was not the source of the precipitation errors. As will be discussed in the following section, future sensitivity tests should involve using two different SST fields, resolution, and interpolation of the initial conditions. The representation of these physical and dynamic processes for the evolution of moisture, mass, and temperature at the sub-grid scale in the model atmosphere failed to justify the model errors in the Eta model and further supported the hypothesis that the improved MM5 forecasts for the event were not in response to errors in the sea surface temperatures.

To further evaluate the model differences in the Eta and MM5 model-generated precipitation forecasts, and to discern if the Eta model QPF errors evolved in a region of limited moisture and improper temperatures for precipitation, the quasi-geostrophic analysis was performed. Cross-sections from Buoy 41001 to London, Kentucky, provided an analysis of the coastal front, damming region, and mountain effects. These analyses indicated that both models over-predicted the offshore values of pressure vertical velocity compared to the EDAS analysis by 0000 UTC 3 December. The potential impact of increased 15-kilometer grid spacing in the MM5 model compared to the 40-kilometer Eta and EDAS grid may explain the larger quantities in the MM5 model. An Eta model bias is evident by 0600 UTC 3 December in the 18-hour
forecasts with a definite western shift in the upward motion fields relative to the MM5 forecasts. Subsidence was enhanced in central North Carolina by 1200 UTC 3 December in the MM5 model forecasts with positive values of pressure vertical velocity extending to near Rocky Mount with the Eta producing subsidence only immediately adjacent to the Appalachian escarpment. By 1800 UTC 3 December, the Eta and MM5 shifted the primary lift to the coastal sections of North Carolina.

Using the EDAS analysis, the regions of coincident upward vertical motion and relative humidity greater than 80 percent indicated that the Eta model under-estimated the level of saturation near the 700 hPa level or offshore near the developing low pressure center at 0000 UTC 3 December. The Eta continued to over-estimate the saturation in the 700-500 hPa layer over central North Carolina and served as a preliminary indication that convective activity may have been active in the Eta model in a region where atmospheric stability was not conducive to sub-grid scale precipitation. The largest values of pressure vertical velocity in the MM5 model were not co-located with the relative humidity values in excess of 80 percent. Prior to 0600 UTC, low-level drying in the EDAS and MM5 model forecasts suggest that the strength of the cold air damming episode was indeed increasing with time throughout the event. The Eta model did not represent the low-level drying below 850-hPa with relative
humidity greater than 80 percent. The upward vertical velocities in the MM5 were located in a region of substantial atmospheric saturation just to the west of Raleigh by 1200 UTC on 3 December in a shallow layer. The Eta model, however, continued to predict large scale vertical ascent with relative humidity in excess of 80 percent across much of central and eastern North Carolina.

The lack of actual observed sounding data in central North Carolina presented a challenge to the evaluation of the vertical temperature and moisture quantities across the region of the greatest error in the model-generated quantitative precipitation forecasts. For that reason both the EDAS analyzed soundings at Raleigh-Durham International Airport (RDU) and observed soundings at Morehead City (MHX) and Blacksburg, Virginia (RNK), were used as comparison to the MM5 and Eta model forecasts for evaluation of stability, temperature, and moisture at specific locations across the state. As an indication of precipitation efficiency, the cloud top temperature values of less than –10°C are used as the critical value for indicating an environment conducive to the development of ice crystals and, therefore, precipitation.

In RDU at 0000 UTC 3 December, the MM5 actually exhibited temperatures that were too warm for precipitation, while the Eta model cloud top temperatures were near –10°C. Low-level dry air was best
predicted by the MM5 model through 0600 UTC 3 December where cloud
tops in the Eta warmed and MM5 cooled to below −10°C. Moisture at the
low-levels in the Eta model relative to the MM5 model continues to
support the impact of continued precipitation development on the large
scale and lesser amounts, if any, on the sub-grid scale.

The skew-T analyses at RNK and MHX were also compared to the
forecast thermodynamic profiles from the Eta and MM5 output valid at
0000 UTC and 1200 UTC 3 December. At 0000 UTC, the RNK sounding
confirmed that the cooler, drier air at the surface to 850 hPa was better
represented by the MM5 model, an indication of stronger cold air
damming. Drier air aloft in the MM5 model may be an indication of the
response to the initial precipitation and better representation of the 500
hPa trough evolution. This is further verified by the MM5 forecast and
EDAS dewpoint temperature at 500 hPa being 10 to 20°C cooler than the
Eta model forecasts. At RNK and MHX the overactive 850 hPa jet was
observed with errors of 10 to 30 knots in both the Eta and MM5. The Eta
model produced the largest errors of 25 to 30 knots at 1200 UTC 3
December at MHX. The cold, dry air at low-levels in North Carolina and
Virginia was poorly handled by the Eta model with positive biases evident
in the dewpoint temperatures at RNK and surface temperature errors of
+10°C with no inversion of stability evident in the predicted atmospheric
profile at the PBL top.

Better representation of the precipitation development could be related to the early development of low pressure in the MM5 model in response to better representation of the evolution of the 500 hPa vorticity lobe. The MM5 also forecast the observed coastal front development better than Eta model forecasts. Evidence was also provided for the stronger cold air damming, eastward-displaced coastal front, and lack of variation in sensitivity experiments within the MM5 model forecasts for precipitation. In comparison to the Eta model forecasts, drier mid- to upper-level atmospheres in the MM5 model at 12-hours provides some evidence of atmospheric drying in response to the anticyclonic vorticity advection, subsidence, and precipitation of available moisture associated with the lead disturbance. Eta model forecasts show that the 850 hPa jet was over-predicted, as did the MM5 model forecasts; however, the errors in the MM5 were of approximately half the magnitude and therefore warm advection at 850 hPa could have resulted in enhanced totals in the Eta model over central North Carolina compared to the MM5. The observed drying at 850 hPa at 1200 UTC 3 December in the sounding at MHX is, perhaps, the most significant finding in validating that despite the initial conditions in the Eta and MM5 model being the same, the MM5 model represented the evolution of the atmospheric thermodynamics properly
due to better representation of these physics in the model design. Based on comparison to EDAS profiles, the Eta model failed to predict values of cloud top temperatures less than –10 °C for a length of time sufficient to support the development of the significant precipitation over central North Carolina that was diagnosed in the model guidance. However, conclusive evidence is unavailable due to the lack of observed soundings in central North Carolina during the event.

6.3 Future Work

The role of errors in the initial SST analysis in the model QPF errors cannot be excluded as the source of the superior forecast in the MM5 model relative to the Eta model. Sufficient evidence is provided in the current study that the 500 hPa trough was a contributing factor in the evolution of the coastal cyclogenesis event. Sensitivity studies with the SST fields, model resolution, and initial conditions are required to absolutely determine the source of the QPF errors in central North Carolina. The optimal testing procedure to evaluate the effect of SST on precipitation generation would involve the sensitivity studies using the 50-kilometer MCSST and the RTG_SST implemented in January 2001 to run the case to discern the effects on QPF.

In addition, further in-depth analysis with respect to the role of
precipitation parameterization in the Eta model forecasts should be performed in a more quantitative, rather than qualitative sense, to evaluate the relative contributions of individual physical and dynamic processes involving errors of moisture and temperature throughout the atmosphere.

Additional studies could investigate the differences in observations and EDAS analysis through comparison with the RUC or a hand-drawn surface and upper air analysis as an independent variable. The comparison of point observations to fields would also increase the reliability of any conclusions made with regards to the EDAS representation of the synoptic conditions across the Southeast. The case study of 2 – 4 December 2000 provides an example of quantitative precipitation errors that evolve from model physical design and observational data quality issues with different forcing mechanisms evident in two independent modeling systems. The development of an operational real-time model and data assimilation scheme may provide forecasters with a location-specific tool to verify initial conditions and to provide boundary conditions for future modeling projects. The importance of testing the model performance in multiple winter season events over several years is essential to the overall performance level of the MM5 in the North Carolina cool season. This study provides the basis for further
investigation into both models but concludes that the effect of the upper-level forcing on the evolution of the low-level topographically-induced phenomena and the surface-based forcing of upper-level dynamics can be of equal magnitude and importance in the skill of winter season precipitation forecasts. The complexity of the feedback processes between topographic flow and precipitation with implications on winter season forecast skill will become clearer as we learn the limitations of modeling systems that are not currently well-represented by current technology.


Zhang, F., C. Snyder, and R. Rotunno, 2001: Sensitivity to initial state and grid resolution in the prediction of the January 2000 East Coast snowstorm. Preprints. *18th Conference on Weather Analysis and Forecasting*, Fort Lauderdale, FL.


Table 1.1 6-hourly and total 48-hour integrated precipitation totals from the Eta model for initialization times every 6 hours from 1800 UTC 1 December 2000 through 1800 UTC 3 December 2000 for the Raleigh-Durham International Airport, North Carolina (KRDU).

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<td>2.39</td>
<td>1.46</td>
<td>0.97</td>
<td>0.28</td>
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Table 2.1  Typical relationship between anticyclonic (AVA) and cyclonic (CVA) vorticity advection and downward (DVM) and upward (UVM) vertical motion provided the level of consideration is above the level of convergence.

AVA $\rightarrow$ CONVERGENCE $\rightarrow$ DVM
CVA $\rightarrow$ DIVERGENCE $\rightarrow$ UVM
Table 2.2  Forcing for ascent as categorized by Wetzel and Martin (2000) for determining regions of sufficient vertical motion to support the generation of winter season precipitation.

<table>
<thead>
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<th>Forcing</th>
<th>Character</th>
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<tr>
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<td>Moderate</td>
</tr>
<tr>
<td>&lt; -15</td>
<td>Strong</td>
</tr>
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</table>
Table 5.1  Table indicating the various sensitivity experiments with MM5.

NATURE OF THE SENSITIVITY TESTS

* Cumulus Parameterization (CPS)
  Kain-Fritsch vs. Betts-Miller

*Planetary Boundary Layer (PBL)
  Blackadar vs. MRF

*CPS and PBL (CPSPBL)
  Kain-Fritsch/Blackadar vs. Betts-Miller/MRF

* Precipitation Microphysics (MICRO)
  Dudhia Simple Ice vs. Reisner-2
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Figure 2.2 Ingredients to qualify the tendency of the atmosphere to produce winter season precipitation from Wetzel and Martin (2000).
Figure 2.3  Diagram of 500 hPa geopotential heights (solid in dam) and absolute vorticity (dashed in $\times 10^{-4} \text{s}^{-1}$) contours representative of a trough (left) and ridge (right). Indications of the anticyclonic (NVA) and cyclonic (CVA) vorticity advection are noted. Parcels entering the vorticity maximum (red) and minimum (purple) must maintain equilibrium with the increasing (decreasing) environmental vorticity. To accomplish this, the red parcel converges to increase the rate of spin (e.g. ice skater) while the parcel in the CVA diverges to lower the vorticity (Adapted from McGill University Course Notes).
Figure 2.4  Idealized representation of the circulation induced by cyclonic (CVA) and anticyclonic (AVA) vorticity advection between the surface and 500 mb using the quasi-geostrophic approximation. Indicated by arrows, the region of convergence and divergence and subsequent response in the vertical motion is based on the assumption that the level of non-divergence (dashed red line) is below the level of the vorticity advection analysis considered.
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Figure 3.2 The horizontal domain of the Eta-212 model as of September 2000 (dashed line) with 32-kilometer resolution and 2 December 2000 (solid line) with 22-kilometer resolution indicating the increased resolution incorporated with changes in September 2000 according to Rogers et al. (2000).
Figure 3.3 The horizontal domain of the MM5 model on 2 December 2000 (solid line) with 45-kilometer outer domain (D01) and 15-kilometer inner domain (inset) resolution as used at the North Carolina Supercomputing Center.
Figure 3.4 The vertical resolution of the MM5 model on 2 December 2000 (solid line) with 31-layers with increased resolution in the planetary boundary layer below 800 hPa ($\sigma = 0.812$).
Figure 4.1 Observational analyses from surface and upper air soundings with EDAS analysis plots for (a) sea-level pressures (pink lines) and (b) 850 hPa geopotential heights (red lines) and temperature (blue lines).
Figure 4.1 Observational analyses from surface and upper air soundings with EDAS analysis plots for (c) 500 hPa geopotential heights (red lines) and temperatures (blue dashed lines), and (d) 250 hPa geopotential heights (red lines) and isotachs (blue lines and shaded >80 kts) valid 1200 UTC 2 December 2000.
Figure 4.2 Plot of the (a) radar imagery and (b) infrared satellite imagery valid at 1200 UTC 2 December 2000.
Figure 4.3 Same as Figure 4.1 except valid at 0000 UTC 3 December 2000.
Figure 4.3 Same as Figure 4.1 except valid at 0000 UTC 3 December 2000.
Figure 4.4  Same as in Figure 4.1 except valid at 1200 UTC 3 December 2000.
Figure 4.4  Same as in Figure 4.1 except valid at 1200 UTC 3 December 2000.
Figure 4.5  Surface analyses of sea level pressure (solid lines) from the Eta Data Assimilation System (EDAS) valid at (a) 1200 UTC 2 December 2000 and (b) 0000 UTC 3 December 2000.
Figure 4.5 Surface analyses of sea level pressure (solid lines) from the Eta Data Assimilation System (EDAS) valid at (c) 1200 UTC 3 December 2000 and (d) 0000 UTC 4 December 2000.
Figure 4.6  As in 4.1 except indicating the EDAS analyses at 850-hPa for wind direction and speed (red barbs), temperature (blue contours, dashed below 0 C), and geopotential height (black lines every 30 m).
Figure 4.6 As in 4.1 except indicating the EDAS analyses at 850-hPa for wind direction and speed (red barbs), temperature (blue contours, dashed below 0°C), and geopotential height (black lines every 30 m).
Figure 4.7 As in 4.1 except indicating the EDAS analyses at 500-hPa for absolute vorticity (blue dashed, shading < -16/s^2) and geopotential height (black lines every 60 m).
Figure 4.7 As in 4.1 except indicating the EDAS analyses at 500-hPa for absolute vorticity (blue dashed, shading $<-16/s^2$) and geopotential height (black lines every 60 m).
Figure 4.8  As in 4.1 except indicating the EDAS analyses at 250-hPa for wind direction and speed (red barbs), isotachs (blue lines every 10 m/s shaded above 70 m/s), and geopotential height ((blue dashed, shading < -16/s^2) and geopotential height (black lines every 60 m).
Figure 4.8  As in 4.1 except indicating the EDAS analyses at 250-hPa for wind direction and speed (red barbs), isotachs (blue lines every 10 m/s shaded above 70 m/s), and geopotential height (blue dashed, shading < -16/s^2) and geopotential height (black lines every 60 m).
Figure 4.9 As in 4.1 except for the radar imagery valid at the same times courtesy of the National Climatic Data Center archives. Heaviest precipitation indicated in green and yellow. Lighter precipitation in blue.
Figure 4.9 As in 4.1 except for the radar imagery valid at the same times courtesy of the National Climatic Data Center archives. Heaviest precipitation indicated in green and yellow. Lighter precipitation in blue.
Figure 4.10 EDAS analyses of (a) 500-hPa vorticity advection (pink) and geopotential height (red) and (b) absolute vorticity (shaded) and geopotential height (green) valid at 1200 UTC 2 December 2000.
Figure 4.10 Plots of (c) NEXRAD 4-km radar, and (d) station plots from METAR and ship/buoy reports valid at 1200 UTC 2 December 2000.
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Figure 4.11 Plots of (c) NEXRAD 4-km radar, and (d) station plots from METAR and ship/buoy reports valid at 1800 UTC 2 December 2000.
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Figure 4.12 Plots of (c) NEXRAD 4-km radar, and (d) station plots from METAR and ship/buoy reports valid at 0600 UTC 3 December 2000.
Figure 4.13 EDAS analyses of (a) 500-hPa vorticity advection (pink) and geopotential height (red) and (b) absolute vorticity (shaded) and geopotential height (green) valid at 1200 UTC 3 December 2000.
Figure 4.13 Plots of (c) NEXRAD 4-km radar, and (d) station plots from METAR and ship/buoy reports valid at 1200 UTC 3 December 2000.
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Figure 4.15 EDAS analyses (red) and Eta model forecasts (pink) from 1200 UTC 2 December 2000 run of sea level pressure valid at (c) 1200 UTC 3 December and (d) 1800 UTC 3 December 2000.
Figure 4.16 EDAS analyses (red) and MM5 model forecasts (pink) from 1200 UTC 2 December 2000 run of sea level pressure valid at (a) 0000 UTC 3 December and (b) 0600 UTC 3 December 2000.
Figure 4.16 EDAS analyses (red) and MM5 model forecasts (pink) from 1200 UTC 2 December 2000 run of sea level pressure valid at (c) 1200 UTC 3 December and (d) 1800 UTC 3 December 2000.
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Figure 4.19 Surface observations (red and pink) and model forecasts of 1000-hPa frontogenesis (contours and shaded) from the MM5 (c) 12-hour and (d) 18-hour forecasts valid at 0000 UTC and 0600 UTC 3 December 2000, respectively.
Figure 4.20 Surface observations (red) and model forecasts of 1000-hPa frontogenensis (contours and shaded) from the Eta model (a) 24-hour and (b) 30-hour forecasts valid at 1200 UTC and 1800 UTC 3 December 2000, respectively.
Figure 4.20 Surface observations (red) and model forecasts of 1000-hPa frontogenesis (contours and shaded) from the MM5 (c) 24-hour and (d) 30-hour forecasts valid at 1200UTC and 1800 UTC 3 December 2000, respectively.
Figure 5.1 Sea level pressure forecasts from the MM5 model sensitivity experiments indicating MICRO (green), CPSPBL (light blue), CPS (dark blue), PBL (red), and the operational CNTRL run (pink) for forecast hours 12, 18, and 24 valid at (a) 0000 UTC 3 December, (b) 0600 UTC 3 December, and (c) 1200 UTC 3 December 2000, respectively.
Figure 5.2 3-hourly precipitation forecasts from the (a) MM5 CNTRL, (b) MM5 MICRO, (c) MM5 CPS, (d) PBL, (e) MM5 CPSPBL, and (f) Eta model guidance for forecast hour 12 valid at 0000 UTC 3 December 2000.
Figure 5.3 3-hourly precipitation forecasts from the (a) MM5 CNTRL, (b) MM5 MICRO, (c) MM5 CPS, (d) PBL, (e) MM5 CPSPBL, and (f) Eta model guidance for forecast hour 18 valid at 0600 UTC 3 December 2000.
Figure 5.4 3-hourly precipitation forecasts from the (a) MM5 CNTRL, (b) MM5 MICRO, (c) MM5 CPS, (d) PBL, (e) MM5 CPSPBL, and (f) Eta model guidance for forecast hour 24 valid at 1200 UTC 3 December 2000.
Figure 5.5 3-hourly precipitation forecasts from the (a) MM5 CNTRL, (b) MM5 MICRO, (c) MM5 CPS, (d) PBL, (e) MM5 CPSPBL, and (f) Eta model guidance for forecast hour 30 valid at 1800 UTC 3 December 2000.
Figure 5.6 Diagram of the cross-section taken from Buoy 40001 to London, Kentucky with Rocky Mount, North Carolina (RWI), (black dot) and central North Carolina (yellow stripe) indicated.
Figure 5.7 Cross-sections of pressure vertical velocity (red) and relative humidity (pink) from (a) EDAS analysis valid at 0000 UTC 3 December 2000.
Figure 5.7 Cross-sections of pressure vertical velocity (red) and relative humidity (pink) from (b) Eta 12-hour forecasts valid at 0000 UTC 3 December 2000.
Figure 5.7 Cross-sections of pressure vertical velocity (red) and relative humidity (pink) from (c) MM5 12-hour forecasts valid at 0000 UTC 3 December 2000.
Figure 5.8 Cross-sections of pressure vertical velocity (red) and relative humidity (pink) from (a) EDAS analysis valid at 0600 UTC 3 December 2000.
Figure 5.8 Cross-sections of pressure vertical velocity (red) and relative humidity (pink) from (b) Eta 18-hour forecasts valid at 0600 UTC 3 December 2000.
Figure 5.8 Cross-sections of pressure vertical velocity (red) and relative humidity (pink) from (c) MM5 18-hour forecasts valid at 0600 UTC 3 December 2000.
Figure 5.9 Cross-sections of pressure vertical velocity (red) and relative humidity (pink) from (a) EDAS analysis valid at 1200 UTC 3 December 2000.
Figure 5.9 Cross-sections of pressure vertical velocity (red) and relative humidity (pink) from (b) Eta 24-hour forecasts valid at 1200 UTC 3 December 2000.
Figure 5.9 Cross-sections of pressure vertical velocity (red) and relative humidity (pink) from (c) MM5 24-hour forecasts valid at 1200 UTC 3 December 2000.
Figure 5.10 Cross-sections of pressure vertical velocity (red) from (a) EDAS analysis valid at 1800 UTC 3 December 2000.
Figure 5.10 Cross-sections of pressure vertical velocity (red) and relative humidity (pink) from (b) Eta 30-hour forecasts valid at 1800 UTC 3 December 2000.
Figure 5.10 Cross-sections of pressure vertical velocity (red) and relative humidity (pink) (c) MM5 30-hour forecasts valid at 1800 UTC 3 December 2000.
Figure 5.11 Thermodynamic diagrams (skew-Ts) indicating the temperature (blue line), dewpoint temperature (pink line), and wind speed and direction (pink barbs at left) from the surface to 500 hPa from (a) EDAS analysis forecasts valid at 0000 UTC 3 December 2000.
Figure 5.11 Thermodynamic diagrams (skew-Ts) indicating the temperature (blue line),
dewpoint temperature (pink line), and wind speed and direction (pink barbs at
left) from the surface to 500 hPa from (b) Eta 12-hour forecasts valid at 0000
UTC 3 December 2000.
Figure 5.11 Thermodynamic diagrams (skew-Ts) indicating the temperature (blue line), dewpoint temperature (pink line), and wind speed and direction (pink barbs at left) from the surface to 500 hPa from (c) MM5 12-hour forecasts valid at 0000 UTC 3 December 2000.
Figure 5.12 Thermodynamic diagrams (skew-Ts) indicating the temperature (blue line), dewpoint temperature (pink line), and wind speed and direction (pink barbs at left) from the surface to 500 hPa from (a) EDAS analysis valid at 0600 UTC 3 December 2000.
Figure 5.12 Thermodynamic diagrams (skew-Ts) indicating the temperature (blue line), dewpoint temperature (pink line), and wind speed and direction (pink barbs at left) from the surface to 500 hPa from (b) Eta 18-hour forecasts valid at 0600 UTC 3 December 2000.
Figure 5.12 Thermodynamic diagrams (skew-Ts) indicating the temperature (blue line),
dewpoint temperature (pink line), and wind speed and direction (pink barbs at
left) from the surface to 500 hPa from (c) MM5 18-hour forecasts valid at 0600
UTC 3 December 2000.
Figure 5.13 Thermodynamic diagrams (skew-Ts) indicating the temperature (blue line), dewpoint temperature (pink line), and wind speed and direction (pink barbs at left) from the surface to 500 hPa from (a) EDAS analysis valid at 1200 UTC 3 December 2000.
Figure 5.13 Thermodynamic diagrams (skew-Ts) indicating the temperature (blue line), dewpoint temperature (pink line), and wind speed and direction (pink barbs at left) from the surface to 500 hPa from (b) Eta 24-hour forecasts valid at 1200 UTC 3 December 2000.
Figure 5.13 Thermodynamic diagrams (skew-Ts) indicating the temperature (blue line), dewpoint temperature (pink line), and wind speed and direction (pink barbs at left) from the surface to 500 hPa from (c) MM5 24-hour forecasts valid at 1200 UTC 3 December 2000.
Figure 5.14 Thermodynamic diagrams (skew-Ts) indicating the temperature (blue line), dewpoint temperature (pink line), and wind speed and direction (pink barbs at left) from the surface to 500 hPa from (a) Eta 30-hour forecasts valid at 1800 UTC 3 December 2000. The EDAS analysis was unavailable at this time.
Figure 5.14 Thermodynamic diagrams (skew-Ts) indicating the temperature (blue line), dewpoint temperature (pink line), and wind speed and direction (pink barbs at left) from the surface to 500 hPa from (b) MM5 30-hour forecasts valid at 1800 UTC 3 December 2000. The EDAS analysis was unavailable at this time.
Figure 5.15 Thermodynamic diagrams (skew-Ts) indicating the temperature (blue line), dewpoint temperature (pink line), and wind speed and direction (pink barbs at left) from the surface to 500 hPa from (a) Eta 12-hour forecasts valid at 0000 UTC 3 December 2000 for RNK.
Figure 5.15 Thermodynamic diagrams (skew-Ts) indicating the temperature (blue line), dewpoint temperature (pink line), and wind speed and direction (pink barbs at left) from the surface to 500 hPa from (b) MM5 12-hour forecasts valid at 0000 UTC 3 December 2000 for RNK.
Figure 5.15 Thermodynamic diagrams (skew-Ts) indicating the temperature (blue line), dewpoint temperature (pink line), and wind speed and direction (pink barbs at left) from the surface to 500 hPa from (c) observed sounding valid at 0000 UTC 3 December 2000 for RNK.
Figure 5.16 Thermodynamic diagrams (skew-Ts) indicating the temperature (blue line), dewpoint temperature (pink line), and wind speed and direction (pink barbs at left) from the surface to 500 hPa from (a) Eta 24-hour forecasts valid at 1200 UTC 3 December 2000 for RNK.
Figure 5.16 Thermodynamic diagrams (skew-Ts) indicating the temperature (blue line), dewpoint temperature (pink line), and wind speed and direction (pink barbs at left) from the surface to 500 hPa from (b) MM5 24-hour forecasts valid at 1200 UTC 3 December 2000 for RNK.
Figure 5.16 Thermodynamic diagrams (skew-Ts) indicating the temperature (blue line),
dewpoint temperature (pink line), and wind speed and direction (pink barbs at left) from the surface to 500 hPa (c) observed soundings valid at 1200 UTC 3 December 2000 for RNK.
Figure 5.17 Thermodynamic diagrams (skew-Ts) indicating the temperature (blue line), dewpoint temperature (pink line), and wind speed and direction (pink barbs at left) from the surface to 500 hPa from (a) Eta 12-hour forecasts valid at 0000 UTC 3 December 2000 for MHX.
Figure 5.17 Thermodynamic diagrams (skew-Ts) indicating the temperature (blue line), dewpoint temperature (pink line), and wind speed and direction (pink barbs at left) from the surface to 500 hPa from (b) MM5 12-hour forecasts valid at 0000 UTC 3 December 2000 for MHX.
Figure 5.17 Thermodynamic diagrams (skew-Ts) indicating the temperature (blue line), dewpoint temperature (pink line), and wind speed and direction (pink barbs at left) from the surface to 500 hPa from (c) observed soundings valid at 0000 UTC 3 December 2000 for MHX.
Figure 5.18 Thermodynamic diagrams (skew-Ts) indicating the temperature (blue line), dewpoint temperature (pink line), and wind speed and direction (pink barbs at left) from the surface to 500 hPa from (a) Eta 24-hour forecasts valid at 1200 UTC 3 December 2000 for MHX.
Figure 5.18 Thermodynamic diagrams (skew-Ts) indicating the temperature (blue line), dewpoint temperature (pink line), and wind speed and direction (pink barbs at left) from the surface to 500 hPa from (b) MM5 24-hour forecasts valid at 1200 UTC 3 December 2000 for MHX.
Figure 5.18 Thermodynamic diagrams (skew-Ts) indicating the temperature (blue line), dewpoint temperature (pink line), and wind speed and direction (pink barbs at left) from the surface to 500 hPa from (c) observed soundings valid at 1200 UTC 3 December 2000 for MHX.