ABSTRACT

BORKOWSKI, MATTHEW PAUL. The Significance of Boundary-Layer and Upper-Level Processes in Wintertime Extratropical Cyclogenesis. (Under the direction of Professors Sethu Raman and Gary Lackmann.)

Winter storms along the U.S. East Coast can significantly impact the nation’s economy. Despite improvement in recent decades, numerical models still produce false alarms or underestimate storm development. Operational meteorologists often seek simple analysis methods to quickly condense the wealth of meteorological information into a format that can be easily interpreted. For this reason, the Atlantic Surface Cyclone Intensification Index (ASCII) and Improved ASCII (I-ASCII) were developed. ASCII relates the pre-storm low-level baroclinicity to the deepening rate of East Coast cyclones, while I-ASCII adds a measure of upper-level forcing. Such products are meant to supplement numerical models and help forecasters quickly gauge the likelihood of a rapidly deepening cyclone.

One purpose of this study was to evaluate the performance of I-ASCII at the National Weather Service (NWS) office in Wilmington, NC for the winter of 2005-2006. Unfortunately, major storms did not occur during the study period preventing a statistically significant analysis from being performed. However, some useful analysis was still performed. It was found that ASCII outperformed I-ASCII for all observed cases due to the conflicting grid resolutions used for the original I-ASCII development and its operational implementation. Three quick fixes are proposed: the first to interpolate the analysis to the grids from which the real-time 500-mb absolute vorticity was originally obtained; the second to correlate the values of 500-mb absolute vorticity maxima between the two conflicting grids; the third to employ the ijskip variable in GEMPAK to effectively reduce grid spacing. Only the third solution will allow I-ASCII to outperform ASCII consistently suggesting that
it is the broad net effect of the upper-level divergence pattern which is important in storm development as opposed to the fine details of the pattern.

The second part of this research was to determine the source of bad I-ASCII forecasts. Two cases were chosen: December 25, 2002 in which I-ASCII performed well; and December 14, 2003 in which I-ASCII performed poorly. The Weather Research and Forecasting (WRF) model, operated with a 12-km outer domain and 4-km nested domain, was used to simulate the two cases using high resolution sea-surface temperatures (SSTs) and North American Regional Reanalysis data. The most likely source of I-ASCII error is the quantification of upper-level forcing using 500-mb absolute vorticity, which is highly dependant on grid resolution. In addition, the location, timing, and orientation of vorticity maxima are essential to the evolution of the surface low, but are not included in I-ASCII.

Lastly, two WRF simulations were performed for the February 12, 2006 winter storm. The first objective was to determine whether there are any fundamental differences in the development of Miller Type-A and Type-B snowstorms with respect to the performance of I-ASCII. While there were no significant differences in how well I-ASCII handles the two storm types, the comparison of these two cases revealed the importance of a southerly and easterly component to the surface flow prior to the development of surface lows to help destabilize the boundary layer. The second objective was to identify how sensitive WRF was to differences in the resolution of the input SST analysis. Slight differences in surface heat fluxes may result in more accurate simulations of weakly forced synoptic scale lows or meso-lows when using higher resolution SST analyses, but the impact on strongly forced events is not great enough to necessitate the usage of high resolution SST analyses in all model simulations and forecasts.
THE SIGNIFICANCE OF BOUNDARY-LAYER AND UPPER-LEVEL PROCESSES IN WINTERTIME EXTRATROPICAL CYCLOGENESIS

by

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BIOGRAPHY

Matthew Paul Borkowski was born on 19 November 1979 in Baton Rouge, LA. He spent most of his childhood and adolescence in Pittsburgh, PA, where he developed a great interest in meteorology. After graduating from North Catholic High School in 1998, he attended Penn State University where he was active in a number of activities including: the Campus Weather Service; the Newman Catholic Student Association; and the Earth and Mineral Science Interest House. He also had an opportunity to participate in an internship with AccuWeather, Inc. and the 2001 Summer Institute on Atmospheric and Hydrospheric Sciences at NASA Goddard Space Flight Center in Greenbelt, Maryland. In 2002, Matthew graduated from Penn State with a Bachelor of Science degree in Meteorology. Following graduation, Matthew spent time exploring his interests by attending a semester in the Graduate Atmospheric Sciences program at University of Maryland and working as an operational meteorologist at Fleetweather, Inc. in Hopewell Junction, NY. Ultimately he decided to pursue a Master of Science degree at North Carolina State University and enrolled in the Graduate School in fall of 2004. While at N.C. State, Matthew has had the opportunity to participate in the Raleigh National Weather Service Student Internship program. In May 2005 he began his formal research on the impacts of various surface and upper-level processes on wintertime U.S. East Coast cyclogenesis focusing especially on the effects of the Gulf Stream position and orientation.
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Also deserving of countless thanks are Aaron Sims (SCO NC) and Dr. Michael Brennan (Tropical Prediction Center). Aaron graciously donated his time and skills to assist me in generating customized high resolution SST composites and to get SST composites into a compatible format for the Wilmington NWS office. I cannot express enough thanks for the help of Dr. Brennan who was never too busy to take the time to help solve a GEMPAK or WRF related problem. His patience and willingness to help will not be forgotten.

Finally, I cannot go without giving thanks to my wonderful family. First, I express
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CHAPTER 1: INTRODUCTION

1.1 Motivation

The U.S. East Coast and adjacent coastal waters have long been known to be a favorable area for wintertime cyclogenesis (Sanders and Gyakum 1980; Zishka and Smith 1980; Roebber 1984). In many instances, storms that develop in or near the offshore waters of North Carolina and northward can be classified as meteorological “bombs”. These “bombs”, or explosively deepening cyclogenesis events, are rapidly deepening cyclones that experience surface pressure drops exceeding 1 mb hr$^{-1}$ for 24 hours at 60ºN (Sanders and Gyakum 1980). These extratropical cyclones (ETCs) are of great meteorological interest often having long lasting effects on the Eastern U.S. and posing a significant threat to life and property. Kocin and Uccellini (2004) briefly report on the impacts of East Coast snowstorms, citing the biggest impact being to transportation, as both automobile travel and aviation become dangerous and even impossible due to heavy snows. The nation's economy also suffers as the biggest storms often result in costs in the billions of dollars due to property damage, snow removal efforts, and the closure of business (Kocin and Uccellini 2004; NCDC 2003). Even without the snow, reports from the National Climatic Data Center's (NCDC) National Storm Database demonstrate that flooding and strong winds can result in significant damage and loss of life.

Naturally, given their prevalence along the East Coast, reliable forecasts for ETCs are in high demand. While operational numerical models have improved significantly in their reliability in recent decades, severe busts such as that of the January 24-25, 2000 snowstorm expose their weaknesses (e.g., Brennan and Lackmann 2005; Zhang et al. 2002). Cione et al.
(1993) detailed the impact of Gulf Stream (GS) induced baroclinicity on East Coast cyclones in hopes of finding a simple parameter for accurately forecasting the deepening rates and determining the likelihood of rapidly deepening storms off the Mid-Atlantic Coast based on the pre-storm environment. The Atlantic Surface Cyclone Intensification Index (ASCII) was introduced to accomplish this goal, demonstrating “encouraging results” in an operational evaluation of the tool for the period 1994-1996 (Cione et al. 1998). ASCII was later improved by including a measure of the upper level influence associated with the East Coast cyclones (Jacobs et al. 2005). A real-time operational evaluation and verification of the latest version of ASCII (which will be referred to as I-ASCII) needs to be performed and weaknesses addressed before it can be implemented for widespread operational use. These issues will be addressed in this thesis using a verification of real-time ASCII forecasts generated by the Wilmington, NC (ILM) National Weather Service (NWS) for the 2005-2006 winter season and by multiple case study simulations.

1.2 East Coast Cyclogenesis

1.2.1 Overview

Significant advances in the theory of ETCs began with the emergence of the Bergen school in Norway in the early 20th century. Driven largely by the research of Vilhelm Bjerknes, the focus on such storms shifted to a blend of dynamic and thermodynamic factors associated with their development (Uccellini 1990). From that time up through the present, extensive research has been conducted to identify processes essential to ETC development.

In order for significant cyclogenesis to occur, there must be a mutual interaction
between the surface and upper levels (Bjerknes and Holmboe 1944; Sutcliffe 1939; Sutcliffe and Forsdyke 1950). Such interactions typically occur when an upper-level trough approaches an area of high baroclinicity (i.e. a frontal zone) and a weak surface low or trough (Fig. 1.1a). This interaction is the essential ingredient in the Sutcliffe-Pettersen self-development theory, which details an ideal cyclogenesis scenario (Sutcliffe and Forsdyke 1950; Petterssen 1956; Uccellini 1990). In this scenario, warm advection ahead of the surface low tends to amplify and slow the upper-level ridge while cold advection behind the surface low deepens the upper-level trough (Fig. 1.1b). Upper-level diffluence and wind speed increase downstream of the upper-level trough in response to the changing flow pattern. As a result, net divergence increases, strengthening the surface low.

As the system amplifies, a low-level jet (LLJ) often develops on the southeast side of the surface low (Uccellini 1990). In the case of East Coast cyclogenesis, low-level flow over the Atlantic Ocean increases resulting in higher heat and moisture fluxes. The warm moist air flows westward towards the surface cyclone serving to further destabilize the environment and strengthen the low (Bosart et al. 1995; Bosart and Lin 1984; Bosart 1981). This “warm conveyor belt” (Carlson 1980) also serves to feed moisture into the precipitation shield, the source of abundant latent heat release which provides additional energy for the storm (Robertson and Smith 1983; Kenney and Smith 1983; Chang et al. 1984; Dare and Smith 1984). Eventually, the system occludes as the upper-level trough becomes located directly over the surface low (Fig. 1.1c) effectively ending the feedback cycle and leading to the ultimate demise of the system.

Latent heat release (LHR) also plays a major role in the development and
intensification of developing cyclones. The revival of interest in the topic of LHR was
driven by a study by Danard (1964) who found that the inclusion of LHR in model
simulations was necessary to accurately predict vertical motion and deepening rates. Most
case studies involving LHR since that time show that rapid deepening typically begins as
deep convection and heavy precipitation develop on the northwestern side of the cyclone
(Uccellini 1990). LHR not only increases the angular momentum transport and mass
circulation required to deepen a cyclone, but also impacts the upper-level trough-ridge
structure and jet streaks (Johnson and Downey 1976; Change et al. 1982).

East Coast cyclones can be classified by two different schemes. The first, proposed
by Petterssen and Smebye (1971), identifies two general types of cyclones. Type-A cyclones
are those which are driven largely by surface forcings, while Type-B cyclones are driven
largely by upper-level processes. It follows then that heat fluxes and other surface forcings
may not always be significant in the development processes, and the exact factors responsible
for each storm may vary significantly from one storm to the next. Not to be confused with
the Petterssen and Smebye (1971) classification is the Miller (1946) classification. Miller
classified Type-A cyclones as storms which form over the Atlantic Ocean (or Gulf of
Mexico) and move northeast, sometimes undergoing a period of rapid intensification. Miller
Type-B systems are those which are characterized by a redevelopment or transfer of energy
between two cyclones near the coastline. The two classification schemes from this point will
be differentiated as Petterssen Type or Miller Type storms.

Ultimately, multiple factors drive cyclogenesis. Often, complex feedbacks develop
which contribute to the rapid strengthening of East Coast cyclones, often making it difficult
to determine what forces are directly responsible for storms development and intensification as opposed to those which only affect the storm indirectly (Kocin and Uccellini 2004). Therefore, examining more than a few such factors can be an overbearing and complex task. The main focus for this thesis will be on the Gulf Stream induced baroclinicity, surface heat fluxes, and upper-level forcing, which are most relevant to ASCII, and are discussed in more detail below.

1.2.2 Role of the Gulf Stream on Pre-storm Baroclinicity

Sanders and Gyakum (1980) were among the first to perform a detailed climatology of rapidly deepening cyclones on the East Coast. They found that unlike tropical cyclones, which require a minimum threshold for SSTs to develop, ETC development is not dependent on the magnitude of SSTs and typically occurs near the strongest SST gradients. It is here that baroclinic instability, which has been linked to rapidly deepening cyclones (Rogers and Bosart 1986; Nuss and Anthes 1987; Uccellini et al. 1987), is the greatest. A natural baroclinic zone exists during the wintertime off the U.S. Mid-Atlantic Coast due to the presence of the Gulf Stream. The near-surface horizontal gradient between cold continental air and the warm air above the Gulf Stream (GS) can be sharp, and is dependent on two factors.

First and foremost, the near-shore temperature gradient is dependent on the surface temperatures at the coast. Cold-air outbreaks (CAOs), characterized by cold Canadian air advected over the Eastern U.S. by strong northwesterly winds, frequent the East Coast during the winter season and serve to reduce the coastal surface temperatures, while the SSTs over
the GS remain relatively constant. Typical wintertime SSTs at the Gulf Stream Front (GSF) offshore of North Carolina range from 22ºC to 25ºC, while onshore temperatures range from -20ºC to 10ºC during CAOs (Cione et al. 1993). Initially, during these CAOs, the East Coast is under broad anti-cyclonic flow unfavorable for cyclogenesis. However, as cold air is modified by the GS, an inverted trough typically forms at the surface near the GS where the near-shore temperature gradient (and likewise, the baroclinicity) is enhanced. This combined with an approaching upper-level trough creates a favorably unstable environment off the North and South Carolina Coastlines ripe for cyclogenesis.

The second factor influencing the temperature gradient between the coast and the Gulf Stream is the exact position of the GSF. Variations in GS path are often a result of the presence of an underwater topographic feature, the Charleston Rise, located at 32ºN and 79ºW off of the South Carolina Coast (Pietrafesa et al. 1979; Brooks and Bane 1978). As GS water flows northward, it is deflected to the right by the rise. Warm-core eddies frequently develop downstream of the deflection point (Fig. 1.2), with an average of 22 occurring each year (Hogg and Johns 1995). Here, eddy kinetic energy values for the GS reach their peak values, resulting in a high degree of variability in the Gulf Stream position (Richardson 1983). The position of the Gulf Stream Front (GSF) can vary from 15 to 120-km offshore of Hatteras, NC (HSE) and from 100 to 300-km offshore of Wilmington, NC (ILM) (Cione et al. 1993).

Doyle and Warner (1993) have discussed the influence of GS features and SST gradients on the low-level processes that affect cyclogenesis. Due to regular CAOs and lateral meandering of the GSF position, the near-surface temperature gradient offshore of the
North Carolina coast varies greatly throughout the winter season. Warm-core eddies, in particular, serve to increase the low-level baroclinicity, which has been linked to the development of closed low-level circulations in the presence of the eddy (Reddy and Raman 1994; Raman and Reddy 1996; Cione and Raman 1995). In addition, Cione et al. (1993) linked the increase of offshore baroclinicity to stronger cyclonic storm development. However, it is the magnitude of the surface heat fluxes, discussed next, that ultimately determine how efficiently surface baroclinicity translates into low-level instability.

1.2.3 Surface Heat Fluxes

Sensible and latent heat fluxes from the surface to the atmosphere are proportional to wind speed and the temperature gradient between the surface and atmosphere. Over the ocean, these fluxes typically serve to transport heat and moisture from the ocean surface into the marine boundary layer (MBL). Sensible and latent heat flux ($Q_s$ and $Q_l$ respectively) can be estimated using the bulk aerodynamic closure scheme,

$$Q_s = \rho C_p \left( \overline{w'\theta'} \right)_b$$  \hspace{1cm} (1.1)

$$Q_l = L_v \left( \overline{w'q'} \right)_b$$  \hspace{1cm} (1.2)

where $\rho$ is the density of air, $C_p$ is the specific heat of air, $L_v$ is latent heat of vaporization, $\overline{w'\theta'}$ is potential temperature flux, and $\overline{w'q'}$ is specific humidity flux. The potential temperature and humidity fluxes are proportional to wind speed and the vertical gradient of potential temperature and humidity respectively.

There is still some uncertainty about the significance of the contribution of surface heat fluxes to cyclogenesis in light of other significant forcings such as upper-level
divergence. For instance, a modeling study by Danard and Ellenton (1980) found that sensible and latent heat fluxes from the ocean had little impact on a cyclone during its rapidly deepening phase. It is more likely that the latent and sensible heat fluxes are most important during the pre-conditioning and early development of cyclones. Kuo et al. (1991) note a decrease in storm development when surface heat fluxes are turned off in the 24-48 hours prior to storm development, with fluxes occurring in the within 24 hours of development and through the remainder of the simulation having little impact on storm intensification. Cione and Raman (1995) site an overall reduction in intensity of a meso-low when sensible heating is removed but point out that most of the influence of sensible heating was in the first 24 hours of development. Similar results are suggested by Reed and Albright (1986) and Fantini (1991). It is likely that the influence of surface heat fluxes is highly case dependent (Uccellini 1990). Nuss and Anthes (1987) explain that the phase of heat fluxes relative to the low-level temperature gradient is important, as heat fluxes will only serve to enhance storm development if they are positive in the warm sector of the cyclone.

Early work done on air-sea interaction hypothesized the significance of heat and moisture fluxes in cyclogenesis but with little insight into its exact role (Winston 1955; Petterssen et al. 1962; Pyke 1965). Later studies by Bosart (1981) and Bosart and Lin (1984) concluded that heat fluxes were important in the establishment and maintenance of the low-level baroclinic zone associated with the GSF and the coastal front (Bosart 1981; Bosart and Lin 1984). However, the most important function of heat fluxes is likely their modification of the MBL.

Observational and modeling studies have shown that total surface heat fluxes can
approach 1500 Wm$^{-2}$ during CAOs across the eastern U.S. prior to cyclogensis and have a significant impact on the structure of the MBL (Raman and Riordan 1988; Bane and Osgood 1989; Vukovich 1991). These fluxes serve to decrease the low-level static stability (Wash 1988) in the MBL, which in turn impacts cyclone development and strengthening in a couple of ways. First, Staley and Gall (1977) showed that decreased static stability results in increased baroclinic growth rates. Also, decreased static stability increases the potential for increased vertical motion, vortex stretching (as dictated by quasi-geostrophic theory), and rapid deepening which ultimately contributes to the development of a LLJ (Wash et al. 1988; Uccellini et al. 1987).

In addition to the decreased static stability, sensible and latent heat fluxes serve to heat and moisten the air mass to the east of the developing cyclone which feeds into the storm via the warm conveyor belt or LLJ (Carlson 1980; Bosart 1981; Nuss and Anthes 1987; Mailhot and Chouinard 1989). A simulation of the 1979 Presidents’ Day storm by Whitaker et al. (1988) show that potential temperature and specific humidity can increase by 7 K and 4 g-kg$^{-1}$ for a parcel in the warm conveyor belt before it begins to ascend into the core of the storm. This transport of warm moist air feeds the diabatic processes (discussed previously) which further help to strengthen the developing storm. This is verified by Uccellini et al. (1987), who demonstrated that removal of sensible and latent heat fluxes from the MBL for the 1979 Presidents’ Day Storm resulted in a reduction of precipitation rates associated with the deepest convection.

It appears that sensible and latent heat fluxes are most crucial in the period just prior to cyclogensis, or during the very early stages of development. Higher values for heat
fluxes will result in a greater modification of the boundary layer, reducing static stability and increasing the temperature and humidity which may ultimately enhance convection if this air is effectively fed into the developing cyclone. A sharp gradient in heat fluxes (which result from a strong temperature gradient) will help enhance the baroclinicity within the MBL.

1.2.4 Upper-level Forcing

Early pioneers in the field of meteorology had a good handle on the necessary processes essential for cyclogenesis by the early 20th century, recognizing that the continuity equation dictates that convergence and pressure falls at the surface require upper-level divergence to remove air from the column of air below (Palmén and Newton 1969). It was also clear that divergent upper-level flow was often found above strengthening surface lows. Jacob Bjerknes (son of Vilhelm) was the first to make practical use of the tendency equation, derived by Margules in 1904, and related lower-level pressure changes to upper-level divergence (Bjerknes and Holmboe 1944; Palmén and Newton 1969). Combined with Bjerknes’ 1937 observation that upper-level divergence typically lies downstream of upper-level troughs, he proposed that the optimal location for surface development would be in this downstream region, a theory later confirmed by Sanders and Gyakum (1980). Surface disturbances cannot deepen if they are located directly underneath the center of an upper-level trough, where divergence is zero, thus requiring a westward tilt of the trough axis with height which increases the baroclinicity. Work done by Charney (1947) and Eady (1949) “identified preferred wavelengths of trough-ridge systems that were likely to produce cyclonic development” and emphasized the need for baroclinic instability (Kocin and
Uccellini 2004). Ultimately, the westward phase tilt into the shear flow results in baroclinic conversion providing the energy for wave growth and subsequently cyclogenesis.

Several factors help to create and enhance upper-level divergence. Sutcliffe (1939) pointed out the need for vertical wind shear and ageostrophic motions to generate upper-level divergence, a finding later expanded upon by Bjerknes and Holmboe (1944). Essentially, winds accelerate from sub-geostrophic at the base of the trough to super-geostrophic at the crest of the ridge increasing divergence downstream of the trough (Kocin and Uccellini 2004). The ageostrophic wind component is also significant in the interaction of jet streaks (Palmén and Newton 1969; Kocin and Uccellini 2004). Jet streaks help enhance upper-level divergence due to the deceleration and acceleration of winds and cyclonic wind shear that results as air passes through the streak. Diffluence, the “rate at which adjacent flow diverges along an axis oriented normal to the flow at the point in question” (Glickman 2000), is frequently linked to increasing upper-level divergence (Palmén and Newton 1969; Kocin and Uccellini 2004). Diffluence increases in response to deepening and negatively tilted troughs as well as the propagation of jet streaks to the base of a trough (Kocin and Uccellini 2004).

Many researchers found that scarce upper-level wind measurements made it impractical to measure upper-level divergence directly (Petterssen 1956; Kocin and Uccellini 2004). Instead, it was found that upper-level divergence could be more easily approximated by the vorticity advection given that divergence is typically associated with areas of cyclonic vorticity advection (Sutcliffe 1939; Bjerknes and Holmboe 1944; Sutcliffe 1947; Petterssen 1956). Sutcliffe (1947) found a link between 500 mb vorticity advection and surface cyclogenesis. Work done by Sanders and Gyakum (1980) and Sanders (1986) verified that
deepening rate at the surface is highly correlated to the rate of positive vorticity advection (PVA) downstream of the upper-level trough.

1.2.5 Impact of Broad Atmospheric Circulations

Winter weather patterns and the frequency of winter cyclones in the Eastern U.S. are influenced by two atmospheric patterns: the El Niño Southern Oscillation (ENSO) and the North Atlantic Oscillation (NAO). ENSO is the interaction between the ocean and atmosphere in the tropical Pacific and Indian Ocean basins. El Niño and La Niña refer to the warming and cooling of the equatorial Pacific waters respectively, while the Southern Oscillation (SO) is the pressure difference between Tahiti and Darwin (Philander 1990). The SO is measured by the SO Index (SOI). A negative SOI indicates the negative phase of the SO and typically results in an El Niño event. The SOI is positive for the positive phase of the SO and typically associated with La Niña events. While the complex interactions are not fully understood, some basic correlations to winter weather conditions in the Eastern U.S. have been identified associated El Niño and La Niña (Fig. 1.3). Generally, El Niño events are associated with cool, wet, and overall stormy conditions in the southeastern U.S. La Niña events often result in warmer conditions across the Eastern U.S. and dry conditions in the Southeast due to the development of an upper-level ridging pattern in the East (Kocin and Uccellini 2004). A study by Shapiro et al. (2000) suggests an increase in negatively tilted troughs (and therefore more intense storms) during El Niño with positively tilted troughs (and weaker storms) being dominant during La Niña.

The second pattern is the NAO, which is the normalized pressure difference between
the subtropical high in the Mid-Atlantic Ocean and the polar low over Greenland (Wanner et al. 2001; CPC (Climate Prediction Center) 2006b). The positive (negative) phase of the NAO is characterized by below-normal (above-normal) heights and pressure around Greenland and above-normal (below-normal) heights and pressure in the Mid-Atlantic. The resulting atmospheric circulations manifest themselves in mild and wet conditions in the Eastern U.S. during the positive phase and colder and drier conditions in the northeastern states during the negative phase (Fig. 1.4) (Kocin and Uccellini 2004; Visbeck 2006). According to Kocin and Uccellini (2004), a negative NAO correlates well to snowier conditions in the Northeast. However, this does not necessarily imply stronger or more frequent storms, as the main reason for the increased snowfall over the positive NAO is colder temperatures, not increased precipitation. In fact, the flow pattern is typically more zonal (west to east) during the negative phase of the NAO (Visbeck 2006) which does not promote rapid development due to the lack of forcing associated with upper-level troughs (as discussed previously). The more amplified pattern observed during the positive phase is more conducive to rapid cyclogenesis.

1.3 Atlantic Surface Cyclone Intensification Index

The combination of recognizing the many factors that contribute to cyclogenesis, anticipating the influence of these factors from numerous weather models and observation networks, and compiling all the information to make a single accurate forecast can be a daunting task. Having tools that automatically survey certain parameters and generate a forecast based on the state of those parameters can be very useful to operational
meteorologists to get a quick overview of conditions and an idea of what kind of
development can be expected. ASCII was developed to relate the 12-hour deepening rate
\((dP/dt)\) of winter cyclones to the pre-storm low-level baroclinicity off the East Coast of the
United States (Cione et al. 1998). In order to adequately capture changes in low-level
temperature gradient caused by Gulf Stream meanders, the area of focus for prior studies was
limited to a defined ASCII domain (Fig. 1.5).

Cione et al. (1998) originally developed ASCII using data from 1982-1990, excluding
the warm-season months of May-September. This climatology was later expanded to include
1991-2002 (Jacobs et al. 2005). The authors examined the National Center for
Environmental Prediction’s (NCEP) North American surface weather maps to ascertain
storm track information. All non-tropical closed surface lows that passed through the ASCII
domain and remained within the domain for a minimum of 6 hours were included in the
study. Storms that meet this criterion and develop during or within 48 hours of a CAO are
referred to as ASCII storms. Here, a CAO is said to occur when near-coastal surface winds
are from the northwest (between 270-360º) for 12 or more hours. The deepening rate for
each ASCII storm while it was within the domain was computed and normalized to 12 hours.
Pre-storm baroclinicity was quantified using the pre-storm baroclinic index (PSBI), defined
by

\[
PSBI = \frac{T_{GSF} - T}{d}
\]  

(1.3)

where \(T_{GSF}\) is the average sea-surface temperature (SST) at the western edge of the GS or
GSF east of ILM and HSE, \(T\) is the average near surface air temperature at ILM and HSE,
and \(d\) is the average distance between the two measurements at ILM and HSE. PSBI was
averaged between HSE and ILM to obtain a single PSBI representative of the entire region. Values for $T_{GSP}$ and $d$ were obtained from a bi-weekly digitized SST composites and are measured where the SST gradient drops significantly (typically below 0.25-0.5°C/km) (Rick Neuherz 2006, personal communication). $T$ was taken to be the coldest 24 hour average temperature during the 48 hours preceding each storm. PSBI is typically expressed in units of °C/10km.

Fig. 1.6 shows the 12-hour deepening rate plotted with respect to PSBI for the 231 ASCII storms that occurred during the period 1982-2002 and fit the ASCII criteria. A linear regression performed by Jacobs et al. (2005) yields a regression coefficient of 0.55 between PSBI and $dP/dt$ that explains 30% of the total variance. These values closely match those computed by Cione et al. (1998) with the 1982-1990 dataset which included 116 storms. But while ASCII documents a clear linear relationship between low-level baroclinicity and storm development, a significant portion of the variance, 70%, is left unexplained by neglecting other significant factors in cyclogenesis, such as upper-level forcing.

Given the wealth of research regarding the importance of upper-level forcing, as discussed above, Jacobs et al. (2005) recognized ASCII had the potential to be improved significantly by including a measure of the upper-level forcing. Using 500-mb absolute vorticity advection as a gauge of upper-level forcing, Sanders (1986) found the deepening rate of intense storms to be strongly linked to upper-level forcing, with a regression coefficient of 0.872. Therefore, Jacobs et al. (2005) introduced upper-level forcing into ASCII using absolute vorticity as a proxy. Although the use of vorticity advection or Q-vectors would be the best gauge of upper-level divergence, due to the complexity of
computing such values in an operational setting, the magnitude of 500-mb absolute vorticity maxima were chosen in place of the actual vorticity advection to represent upper-level forcing.

Higher mandatory levels, such as 300-mb, which may be more indicative of true upper-level divergence, are not used here for a couple of reasons. First, the 500-mb level has a greater ability to resolve wave trains. Also, forcing at the 500-mb level is more likely to reach and interact with surface disturbances. Lastly, it is differential vorticity advection with height which is correlated to divergence. Because wind speeds nearly always increase between the surface and 500-mb, one can assume that the vorticity advection at 500-mb is directly related to the differential vorticity advection between the surface and 500-mb. Meanwhile, wind speed becomes more variable above 500-mb, and the same assumption cannot be applied at 300-mb.

To recalculate the ASCII regression equation, Jacobs et al. (2005) quantified the upper-level forcing for each ASCII event during the period 1991-2002. Absolute vorticity data were taken from the 2.5º×2.5º National Centers for Environmental Prediction-National Center for Atmospheric Research (NCEP-NCAR) 6-hour reanalysis data. For each event, the value of the nearest 500-mb absolute vorticity maximum within 1000-km of the storm center was recorded. Storms were categorized into three bins based on the value for absolute vorticity: strongly forced (greater than 19×10^{-5} s^{-1}), moderately forced (between 15×10^{-5} s^{-1} and 19×10^{-5} s^{-1}), and weakly forced (less than or equal to 15×10^{-5} s^{-1}). The storm data were plotted and a separate regression line was calculated for each vorticity bin (Fig. 1.7). The resulting correlation coefficients were 0.84, 0.86, and 0.59 respectively, all of which were an
improvement on the original ASCII. The new method also explains as much as 74% of variance in deepening rate for strongly forced cases, with lower percentages for weakly and moderately forced cases (35% and 71% respectively). Due to its proven improvement over the original version of ASCII, the new version of ASCII will be referred to as I-ASCII, for improved-ASCII.

The original ASCII method was tested in an operation setting at the Raleigh (RAH) NWS by Cione et al. (1998). ASCII demonstrated a 28.6% reduction in the root mean square forecast error over the then prevalent Nested Grid Model (NGM), with an even greater reduction in error observed when only storms with weak-to-moderate intensification rates were included. Evidence suggested the method was most useful for storms with modest deepening rates which often occur suddenly with little anticipation from forecasters. However, I-ASCII has yet to be extensively tested in an operational setting. While the time constraints of the project in conjunction with the quiet winter of 2005-2006 prevented a statistically significant test of the improved method from being conducted, one of the goals of this research is to perform a preliminary evaluation and make suggested improvements.

It should be noted that a storm does not need to pass over ocean waters to be classified as an ASCII storm. Such land-locked storms may still be impacted by the offshore waters since it is unlikely that a storm is influenced solely by the area directly below the surface low, but rather a broader region. Some of the variability in deepening rate associated with storms passing over land is taken into account in the development of ASCII since it includes both land-locked and oceanic storms. However, it is still likely that when a storm’s surface low remains entirely over land that it behaves differently than ocean storms. Given
the relative area of the ASCII domain covered by water, it is likely much more of the 
variability associated with ocean storms is explained by ASCII compared to land-locked 
storms resulting in an increased likelihood of ASCII error for land-locked storms.

1.4 Objectives and Hypotheses

The primary objective of this research is to evaluate how well synoptic and mesoscale 
forcings driving wintertime East Coast cyclogenesis are captured by I-ASCII. We will begin 
with a preliminary assessment of the performance of I-ASCII for the 2005-2006 winter 
season and identification of potential areas of improvement. Following this, we will attempt 
to identify the source(s) of observed I-ASCII error by comparing a case in which I-ASCII 
performed well and one in which it performed poorly. Specifically the role of sensible and 
latent heat fluxes on the MBL will be investigated and related to the PSBI as computed for 
ASCII. A Miller Type-A cyclone will be examined in detail to determine inherent 
differences between Miller Type-A and Miller Type-B systems that may impact I-ASCII 
performance. Lastly, an examination of the sensitivity of the Weather Research and 
Forecasting (WRF) model to SST resolution will be conducted.

We expect the performance of I-ASCII will be difficult to ascertain given the lack of 
events for the 2005-2006 winter season. However, the subjectivity associated with 
classifying the upper-level forcing for each case may be a concern. Generally, given the 
results of Jacobs et al. (2005), we anticipate that surface and upper-level processes will be 
fairly well represented in I-ASCII. However, some of the spread in deepening rate may be 
explained by considering surface heat fluxes. Lower values for heat flux may prevent the
low-level baroclinicity from penetrating deep into the MBL serving to slow deepening rates of cyclones relative to what would be anticipated by I-ASCII. However, upper-level processes will likely be the dominating factor for cyclogenesis events given that they explain a larger portion of the variance in deepening rates than the baroclinicity alone. Therefore, storms that are better forecasted by I-ASCII are expected to more closely resemble Petterssen Type-B storms. In addition, the timing and location of vorticity maxima, which is not taken into account in I-ASCII, may also explain some of the errors associated with I-ASCII. While we anticipate some differences in surface and upper-level patterns between Miller Type-A and Type-B storms, we do not expect any differences that would significantly impact I-ASCII forecasts. Finally, we do expect WRF to be somewhat sensitive to the resolution of the input SST analysis, with the biggest impact near the surface and over the GS.

Chapter 2 gives an overview of the modeling system and data which will be used in all simulations. In Chapter 3, the operational I-ASCII and its performance in the 2005-2006 winter season will be discussed along with three methods of improving I-ASCII forecasts will be tested. Two events, one in which I-ASCII performed well and one in which I-ASCII performed poorly will be simulated with the WRF model and discussed in Chapter 4. A detailed comparison of the simulated sea-level pressure, upper-level vorticity and heights, surface heat fluxes, and surface convergence will follow in an attempt to investigate the possible reasons for I-ASCII error. A Miller Type-A case from 2006 will be simulated using WRF followed by a detailed analysis and discussion in Chapter 5. The results will be compared with the Miller Type-B cases discussed in Chapter 4. Chapter 5 will also include a SST sensitivity study for the 2006 storm using high and low resolution SST analyses. The
storm will be simulated using high and low resolution SST analyses and the results will be compared. Lastly, Chapter 6 will provide a summary, conclusions, and suggestions for future research.
Figure 1.1. Schematic 500-mb contours (heavy solid lines), 1000-mb contours (thin lines), and 1000-500 mb thickness (dashed), illustrating the “self-development” process during growth of a cyclone. The stages displayed are a developing cyclone (a), a mature cyclone (b), and an occluded cyclone (c) (from Palmén and Newton, 1969).
Figure 1.2. An image of SSTs in the Gulf Stream on a typical day. The image shows the location of the Charleston Rise (black arrow) and the resulting warm-core eddies downstream (figure taken from Jacobs (2005)).
Figure 1.3. General winter weather conditions associated with El Niño (top) and La Niña (bottom) (from CPC (Climate Prediction Center) 2006a).
**Figure 1.4.** General winter weather conditions associated with the positive phase the NAO (top) and the negative phase of the NAO (bottom) (from Visbeck 2006).
Figure 1.5. A map of the U.S. East Coast showing the locations of ILM and HSE. The black box denotes the boundaries of the ASCII domain.
Figure 1.6. The full ASCII dataset (1982-2002) of dP/dt vs. PSBI plotted with corresponding linear regression fit (figure reproduced with data courtesy of Neil Jacobs and Joe Cione).
Figure 1.7. ASCII data (1991-2002) of dP/dt vs. PSBI and sorted by strength of 500mb absolute vorticity. The corresponding linear regression fit for each vorticity bin is displayed (figure reproduced with data courtesy of Neil Jacobs).
CHAPTER 2: MODEL SYSTEM AND DATA DESCRIPTION

2.1 Description of the Model System and Setup

2.1.1 Model System Overview

The numerical simulations performed in this research were conducted using version 2.1.1 of the Weather Research and Forecasting Model (WRF) Advanced Research Core (ARW). WRF-ARW (hereafter referred to simply as WRF) has been developed through the collaboration of numerous groups including: National Center for Atmospheric Research (NCAR) Mesoscale and Microscale Meteorology (MMM) Division, the National Oceanic and Atmospheric Administration’s (NOAA) National Centers for Environmental Prediction (NCEP) and Forecast System Laboratory (FSL), the Department of Defense’s Air Force Weather Agency (AFWA) and Naval Research Laboratory (NRL), the Center for Analysis and Prediction of Storms (CAPS) at the University of Oklahoma, and the Federal Aviation Administration (FAA). WRF is more user-friendly, flexible, and efficient as compared to the Penn State-NCAR fifth-generation Mesoscale Model (MM5). The non-hydrostatic meso-model core (WRF-NMM), which uses different grid architecture and physics options, was implemented operationally by NCEP on June 20, 2006.

Polar stereographic, Lambert-conformal, and Mercator map projections are supported by WRF. Users have the option of one-way, two-way, and moving nests. An overview of the Modeling System components is shown in Fig. 2.1a. The two main programs are the WRF Standard Initialization (WRFSI) and the ARW Model Solver. The optional 3DVAR program, used for ingesting observations, was not used here.

WRFSI is used to set up the model domain and interpolate the necessary data through
the use of several sub programs (Fig. 2.1b). The program gridgen is used to define the model
domain and interpolate terrestrial datasets (i.e. terrain, landuse, etc.) to the domain. Gridded
binary (GRIB) files containing meteorological data are decoded and converted to an
intermediate format by the grib_prep program. The meteorological data is interpolated
horizontally by hinterp and vertically to the model’s η coordinate by vinterp.

The ARW Model Solver performs the actual simulation using the fully compressible,
Euler nonhydrostatic equations in flux form. The model operates on an Arakawa staggering
C-grid using a 3rd order Runge-Kutta time-split integration scheme. A forward-backward
time integration scheme is used for horizontally propagating acoustic modes while a
vertically implicit scheme is used to for vertically propagating acoustic modes and buoyancy
oscillations (Skamarock et al. 2005).

2.1.2 Model Setup

Model simulations were performed using version 2.1.1 of WRF with a two-way
nested grid configuration. A 12-km outer grid with a 4-km nested grid centered over the
ASCII region was employed for each simulation (Fig. 2.2). All simulations were performed
with 31 vertical η levels between the surface and 100mb with 11 levels between the surface
and 850mb. The η coordinate is different than the traditional σ in that it is hydrostatic and is
defined with respect to the dry air mass (Skamarock et al. 2005). Specified (also known as
relaxation or nudging) lateral boundary conditions are used for both the outer grid and the
nested grid. A list of the physics options employed for each simulation is shown in Table
2.1. A description of each option follows.
The Lin et al. microphysics scheme was employed due to its reputation and widespread use over the past two decades. The inclusion of snow processes improves the realism of the model by delaying rain development and reducing the amount of cloud ice (Lin et al. 1983). This is especially important for the wintertime simulations performed here. Improvements to the original scheme are inclusion of saturation adjustments and ice sedimentation (Skamarock et al. 2005).

The NOAH land-surface model (LSM) was the first choice land-surface option due to its documented success in the MM5 and known operational use in the previously operational Eta model (Chen and Dudhia 2001, Skamarock et al. 2005). It is so named because it is the culmination of effort from four major organizations: NCEP (N), Oregon State University (O), the Air Force (A), and the Hydraulic Research Lab (H). NOAH is built on the Oregon State University (OSU) LSM which was chosen due to its ability to accurately represent surface fluxes, surface skin temperature variations, evaporation, and soil moisture (Chen and Dudhia 2001). The NOAH LSM is an improvement over the OSU LSM in its coupling to diurnal and annual variations in soil parameters and vegetation cover.

Unfortunately, due to irresolvable compatibility issues between the NOAH LSM and the NARR dataset in WRF v2.1.1 (which have since been resolved by the WRF developers), the Rapid Update Cycle (RUC) LSM was implemented as an alternative. Like the NOAH LSM, the RUC LSM includes evapotranspiration, soil drainage, runoff, and coupling of heat fluxes to the boundary layer (Smirnova et al. 2000; Skamarock et al. 2005). The RUC LSM relies on explicitly predicted values for soil moisture and temperature as opposed to climatological values. The primary difference between the NOAH and RUC LSM option is

30
the addition of a sixth sub-soil layer and a multi-layer method for handling snow cover in the RUC LSM (Smirnova et al. 1997; Smirnova et al. 2000). The RUC LSM is used operationally in NCEP’s RUC model.

For convective parameterization, the recently updated Kain-Fritsch (KF) scheme (Kain 2004) was chosen to implicitly infer convection on the 12-km domain, but was turned off for the 4-km domain, which is fine enough to resolve cumulus convection explicitly (Weisman et al. 1997; Skamarock et al. 2005). The original KF scheme (Kain and Fritsch 1993) uses a simple cloud model that includes moist updrafts and downdrafts and the effects of entrainment and detrainment. Modifications to the original KF scheme serve to: suppress convection in marginally unstable and dry environments; limit convection in weakly convergent flow; allow for deep convection from shallow clouds when ice-phase processes are present; and allow for shallow convective (non-precipitating) clouds (Kain 2004).

Numerical weather models are known to be highly sensitive to the choice of convective parameterization scheme (Kuo et al. 1996; Warner and Hsu 2000; Mahoney and Lackmann 2006). The KF scheme is chosen over other schemes due to its use and evaluation in the previously operational Eta model (Skamarock et al. 2005).

The Mellor-Yamada-Janjic (MYJ) physics scheme was chosen for the planetary boundary layer (PBL) parameterization scheme, which also includes the effect of turbulence in the free atmosphere. It is a one-dimensional scheme based on the Mellor-Yamada Level 2.5 turbulence closure model (Mellor and Yamada 1982). The top of the PBL is determined using the turbulent kinetic energy (TKE) (Skamarock et al. 2005). Like most PBL schemes, the surface layer and LSM provide the surface fluxes. The scheme was chosen for its use of
TKE in the PBL and once again because of its success in the previously operational Eta model.

Selection of the MYJ PBL scheme in WRF requires that the Janjić Eta surface layer scheme be used to account of effects of surface fluxes. It is based on the Monin-Obukhov similarity theory. The effects of the viscous sublayer are accounted for by molecular diffusion over water and by varying the roughness height for temperature and humidity profiles over land (Janjić 1994; NCAR 2005). Surface fluxes are computed using an iterative method.

For atmospheric radiation, the Rapid Radiative Transfer Model (RRTM) and the Dudhia schemes are used to account for longwave and shortwave radiation respectively. The RRTM, a spectral-band scheme, uses the correlated-$k$ method, while the simple integration of solar flux in the Dudhia scheme accounts for clear-air scattering, water vapor absorption, and cloud albedo and absorption (Skamarock et al. 2005). Both schemes are taken from the MM5 model.

2.2 Description of Data Sources

2.2.1 North American Regional Reanalysis Data

North American Regional Reanalysis (NARR) data were used for model initialization and boundary conditions. According to Mesinger et al. (2006) the NARR is a regional improvement to previous global reanalysis datasets. NARR uses the NCEP-DOE (Department of Energy) Global Reanalysis for lateral boundaries, the NOAH LSM, and various other datasets not utilized by global reanalyses. All data are assimilated and
produced using the NCEP regional Eta model and data assimilation system. NARR data are available at 32-km horizontal grid spacing and 45-layer vertical grid spacing in 3-hourly increments.

There is one issue with the data that had to be resolved prior to performing model simulations. The NARR provides specific humidity, while the relevant V-table in WRFSI expects relative humidity as input. As a result, GRIB files for each time were converted to General Meteorological Package (GEMPAK) grids where relative humidity ($RH$) was computed for each grid point and for each pressure level from temperature ($T$) and specific humidity ($q$). For the following equations, temperature must be given in Kelvin. First, vapor pressure ($e$) is computed by rearranging the equation,

$$q = \varepsilon \frac{e}{p - (1 - \varepsilon)e},$$

(2.1)

to

$$e = \frac{qp}{q - q\varepsilon + \varepsilon},$$

(2.2)

where $\varepsilon = 0.622$ and $p$ is pressure in Pascals. Then, saturation vapor pressure ($e_s$) is computed with the equation,

$$e_s(T) = 0.6112 \exp \left( \frac{17.67T}{T + 243.5} \right),$$

(2.3)

which is accurate to within 0.3% (Bolton 1980). Finally, $RH$ is computed using,

$$RH = \frac{e}{e_s}.$$ 

(2.4)

The data was converted back to GRIB and then ingested into the WRFSI program.
2.2.2 Sea-Surface Temperature Data

SST data for WRF simulations were obtained from two sources. The 0.5º real-time global SST (RTG_SST) analyses were obtained from NCEP, and 1.47-km regional SST images were obtained through the NOAA CoastWatch program. Both datasets are from the Advanced Very High Resolution Radiometer (AVHRR) on the NOAA series of Polar-orbiting Operational Environmental Satellites (POES) which was primarily designed to retrieve SST and cloud data (Li et al. 2001). The 10-day composite SST images were produced from the CoastWatch SST (CW_SST) analyses and overlaid on the RTG_SST analyses for ingestion into the WRF simulations. In the interest of creating manageable file sizes and keeping computation time to a minimum, the CW_SST data were reduced to 3-km grid spacing prior to generating the composites. This procedure should not affect the simulations as the finest grid spacing used for the WRF simulations was 4-km.

The RTG_SST analysis averages satellite retrieved SST values within 0.5º grid boxes. Each analysis also incorporates buoy and ship SST measurements, which are averaged separately. The satellite-derived SST values are corrected based on the in situ data in using the bias calculation technique as described by Reynolds (1988) and Reynolds and Marsico (1993). Essentially, the in situ observations are used to define “benchmark” SST values while the satellite data dictates the shape of the SST field between the benchmarks.

The CoastWatch SST (CW_SST) datasets are provided by the National Environmental Satellite Data Information Service (NESDIS). Analyses are available for a number of regional nodes including the northeast and southeast nodes, which were utilized in this study. The technique used to derive SST values is described by Li et al. (2001a, 2001b).
A non-linear SST algorithm is used to obtain SST and cloud estimates. A split window algorithm is used for both daytime and nighttime SST estimates, with a multichannel SST algorithm applied during the day and a nonlinear SST algorithm used at night. The difference between the 11 and 12-μm channels, difference in albedo on channel 1 and 2, and a spatial uniformity test are used for cloud detection. SST estimates are only made for satellite zenith angles greater than 53º due to the likelihood of cloud contamination and attenuation of surface infrared emissions at lower zenith angles. SST estimates have been shown to be accurate within 0.2ºC (Li et al. 2001a, 2001b).

CW_SST data were downloaded beginning with the initialization day for each model simulation and going back ten days. The composites are generated beginning with the earliest dated file by writing valid SST values to a final template. These data are overwritten by valid SST data from the next time. However, in the presence of cloud cover, the most recent valid SST value is retained. This ensures the final composite will always contain the most recent SST reading for each pixel. An additional constraint placed on the composites was that the value for SST at any one pixel could not change by more than 7K between files, which prevents unrealistic changes in the data due to contamination and misinterpreted cloud masking (Aaron Sims 2006, personal communication). The process is repeated until all data files are processed. The result is a GrADS (Grid Analysis and Display System) data file which is converted to a GEMPAK grid file.

For each model simulation, a single RTG_SST data file for the initialization day is obtained. Using GEMPAK, the CW_SST composite and RTG_SST analysis are interpolated to a 3 km grid covering the WRF domain using a cylindrical equidistant (CED) map.
projection. The CW_SST composite is then overlaid upon the RTG_SST analysis. This fills in gaps in the CW_SST composite in spots where no acceptable SST readings could be ascertained in the 10-days prior to model initialization due to excessive cloud cover. In addition, the northeast and southeast regional nodes only cover an area between 25º-40ºN and 70-92ºW (Fig. 2.3b). The overlay allowed the WRF model domain to be extended beyond this CoastWatch domain. A nine point smoother was applied to the final high resolution overlay to reduce any unnatural SST gradients generated at the edges of the two analyses. The nine point smoother was chosen over a five points smoother (which would have retained more detail) in an effort to make the continuity at the edge of the two analyses as realistic as possible. Fig. 2.3a is an example of original RTG_SST analysis. The additional detail provided by overlaying the CW_SST composite upon the RTG_SST is shown in Fig. 2.3b. The area over which the high resolution data are used is outlined by white lines. Once completed, the overlays were converted into lambert-conformal map projection and converted to WRF intermediate file format, by-passing the grib_prep WRFSI step for the SST data.
Figure 2.1. Flow charts of the: (a) WRF ARW modeling system with components relevant to this study highlighted in red; and (b) WRFSI and its component programs (Wang et al. 2004).
Figure 2.2. Model domain used for all model simulations. The outer domain (outlined in white) has 12km resolution and the inner domain (outlined in yellow) has 4km resolution.

Table 2.1. A summary of physics options used for all WRF model simulations.

<table>
<thead>
<tr>
<th>Physics Option</th>
<th>Scheme Used</th>
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<tbody>
<tr>
<td>Microphysics</td>
<td>Lin et al.</td>
</tr>
<tr>
<td>Longwave radiation</td>
<td>RRTM</td>
</tr>
<tr>
<td>Shortwave radiation</td>
<td>Dudhia</td>
</tr>
<tr>
<td>Surface-layer</td>
<td>Monin-Obukhov (Janjic Eta)</td>
</tr>
<tr>
<td>Land-surface</td>
<td>RUC LSM</td>
</tr>
<tr>
<td>Boundary-layer</td>
<td>Mellor-Yamada-Janjic</td>
</tr>
<tr>
<td>Cumulus</td>
<td>Kain-Fritsch 2</td>
</tr>
</tbody>
</table>
Figure 2.3. Comparison of the original RTG_SST analysis from 23 Dec 2002 (a) and the 10-day CW_SST composite (ending 23 Dec 2002) overlaid upon the RTG_SST analysis (b). Temperatures are in Kelvin. The area covered by the CW_SST dataset is outlined in white.
3.1 Real-Time ASCII Implementation

The driving motivation behind the continuing study of ASCII is the interest in using the tool in an operational setting to supplement numerical model data. The original ASCII methods were tested operationally between February 1994 and February 1996 at the RAH NWS forecast office and exhibited an improvement in deepening rate over the NGM (Cione et al. 1998). ASCII was not tested against the Eta model at the time due to the lack of historical data. However, I-ASCII (as developed by Jacobs et al. (2005)) has yet to be tested in operations. The time constraints of this research prevented an extensive testing of I-ASCII from being performed. In addition, the calm winter of 2005-2006 resulted in only one rapidly deepening cyclone. Therefore, what follows is a preliminary evaluation of I-ASCII as implemented at the ILM NWS.

Richard Neuherz of the ILM NWS modified a previously existing script, which computed the original ASCII forecast, to generate an I-ASCII forecast. The script was set to run at approximately 05Z and 17Z daily from October through April. The first step was to detect whether or not a cold-air outbreak (CAO) had occurred. Events in which no CAO is observed are excluded from this study, because original ASCII theory required the presence of a CAO to pre-condition the atmosphere to explosive cyclogenesis. METARs (Meteorological Terminal Air Reports) from HSE and ILM from the previous seven days were used to screen for CAOs. To help filter out instances where surface winds decouple from nighttime synoptic flow, the data were supplemented by observations from buoys located at Diamond Shoals and Frying Pan Shoals (stations 41025 and 41013 respectively).
where winds are less likely to decouple. If a wind direction between 270º and 360 º (inclusive) persisted for 12-hours or greater, a CAO was said to have occurred. Temperatures were not considered in the detection of CAOs. The end of the CAO occurred when the wind direction shifted outside of the aforementioned range for more than three hours. If a CAO was not detected or if a CAO was detected but was older than 48-hours, a null ASCII forecast was generated. Otherwise, the procedure for generating a forecast continued to run as follows.

Surface temperatures from HSE and ILM were averaged for the duration of the CAO or for the coldest 24-hours during the CAO if it lasted more than 24-hours. This yielded a mean land temperature, \( T \). The average temperature at the GSF \( (T_{GSF}) \) and offshore distance \( (d) \) was determined manually from the most recent CoastWatch SST analysis. Using the CoastWatch Utilities software from the project website (http://coastwatch.noaa.gov) the SST at the GSF east of HSE and ILM was determined to be where the SST gradients dropped below 0.25-0.5ºC/km (Rick Neuherz 2006, personal communication). The distance between HSE and ILM and the GSF was also obtained from the CoastWatch Utilities software. One caveat was that the GSF be further offshore and cooler at HSE as compared to ILM. \( T_{GSF} \) and \( d \) were taken and averaged at these two points. PSBI can then be computed using the equation,

\[
PSBI = \frac{T_{GSF} - T}{d}.
\]  

(3.1)

As an example, assume the average surface temperature at HSE during the CAO is 10ºC. The temperature at the GSF is 24ºC and the GSF is 60 km offshore of HSE. For ILM, the same values are 7ºC, 25ºC, and 180 km respectively. The values from HSE and ILM are
averaged to obtain values for $T (8.5^\circ C)$, $T_{GSF} (24.5^\circ C)$, and $d (120$ km). Plugging these values into Equation 3.1 yields a PSBI of 1.3$^\circ$C/10km (keeping in mind PSBI is expressed in units of $^\circ$C/10km).

From this, an ASCII forecast was generated based on the original ASCII theory. Here, a single forecast was issued which was valid up to 48 hours from the time it was issued or 48 hours after the end of the CAO. A forecast of “likely to bomb” was issued for PSBI values that exceeded 1.7 $^\circ$C/10 km. A forecast of “not likely to bomb” was issued for PSBI values less than 1.0$^\circ$ C/10 km. A forecast of “indeterminate” was issued for PSBI values ranging from 1.0 $^\circ$C/10 km to 1.7 $^\circ$C/10 km. In other words, a forecast of “likely to bomb” was issued when the PSBI value indicated a deepening rate in excess of 12mb/12hr. When the PSBI value indicated a significant deepening rate greater than 8mb/12hr but below bomb criteria, an “indeterminate” forecast was issued. Otherwise, no more than modest deepening (8mb/12hr or less) was likely. The definition for a bomb was relaxed here (from 24mb/24hr to 12mb/12hr to increase detection of rapidly deepening events). The actual forecast deepening rate was listed on the product for reference, calculated from the original ASCII regression equation:

$$y = -0.975 - 6.697x.$$  \hspace{1cm} (3.2)

Forecasts were valid for any storm meeting ASCII criteria during the forecast period.

In order to gauge whether or not I-ASCII, taking into account upper-level dynamics, performed better than the original, an ASCII with dynamics forecast was generated. I-ASCII
requires a value for 500-mb absolute vorticity while the surface low is within the ASCII domain. As a result, forecasted values for absolute vorticity had to be obtained. The most recent model data available at time of the ASCII forecast was used, typically the 00Z (12Z) model data for the 05Z (17Z) ASCII forecast. NCEP’s Global Forecast System (GFS) model was chosen here due to its availability on a 190 km grid. This grid was the coarsest model grid available and most comparable to the 2.5º × 2.5º NCEP grid used in developing the regression equations for I-ASCII. A semi-circle (extending from 180º to 360º) with a radius of 1000-km and centered on the ASCII domain was defined as a search area. The maximum value for 500-mb absolute vorticity found within this search area for a particular forecast period was used to classify the upper-level support for each forecast. This method is slightly different than the method developed by Jacobs et al. (2005) in which the value of the vorticity maximum closest to the surface low is used as opposed to the highest value within a specified range from the low. Classifications were made based on the original bins: strongly forced (greater than 19×10⁻⁵ s⁻¹), moderately forced (between 15×10⁻⁵ s⁻¹ and 19×10⁻⁵ s⁻¹), and weakly forced (less than or equal to 15×10⁻⁵ s⁻¹) (Jacobs et al. 2005).

Finally, forecasted deepening rates were computed using the I-ASCII regression equation corresponding to the amount of upper-level forcing.

\[
\text{Strong: } y = -2.4551 + 15.094x
\]

\[
\text{Moderate: } y = -2.2673 + 10.302x
\]

\[
\text{Weak: } y = -0.79465 + 6.0919x
\]

A forecast deepening rate was generated for six-hourly increments out to 36-hours, where the forecast hour is relative to the GFS product on which the forecast is based and not the time at
which the ASCII forecast is issued. For example, the 12-hour forecast from the 05Z ASCII product would be valid at 12Z, not 17Z. It should also be noted that forecasts were only valid for up to 48 hours from the end of the CAO, so that all forecasts were not necessarily valid through the 36-hour forecast period. Like the original ASCII forecast, each forecast is applicable to any surface low that meets ASCII criteria during the forecast period. The maximum value for absolute vorticity was determined separately for each forecast hour based on the GFS output of absolute vorticity for that forecast hour. An example of an operational ASCII product with the ASCII and I-ASCII forecast is shown in Fig. 3.1.

3.2 Summary of the 2005-2006 Winter Season

Fig. 3.2 shows a time series of the SOI and NAO indices from January 2005 through March 2006 generated from data from the Climate Diagnostics Center (CDC) website. From December 2005 through March 2006 the SOI was near neutral to positive (indicative of a weak La Niña) and the NAO was slightly negative, except for January, in which it was positive. The resulting 500-mb geopotential height anomalies were near zero for most of the Southeast with slightly negative anomalies across the coastal Northeast (Fig. 3.3). The lack of a persistent trough resulted in warmer than normal conditions across the Northeast and drier than normal conditions across the majority of the Eastern U.S. (Fig. 3.4). These observations are in agreement with the expectations of a positive SOI and negative NAO. Of the storms that passed through or near the ASCII domain, about half moved in an east or east-northeast direction, consistent with the zonal wind pattern observed with the negative
NAO pattern. The resulting setup was rarely favorable for the development of rapidly deepening storms, and as a result, there were only a handful of ASCII storms observed.

### 3.3 Real-Time ASCII Verification

Archived RUC initialization analyses were used to identify ASCII storms and their deepening rates. The 40-km RUC grid was obtained in three hour increments. RUC analyses incorporate many sources of data in addition to surface data, such as aircraft and profiler data. The RUC analyses were chosen over NARR due to their availability for the entire winter season at the time of verification. Sea-level pressure was plotted every three hours for December 1, 2005 through April 30, 2006. A total of 18 storms passed through and spent at least 6 hours in the ASCII domain. However, only half of these storms were officially ASCII storms due to the CAO requirement. Of the nine observed ASCII storms, only one storm, February 11, exhibited rapid deepening. Another modestly deepening storm on March 21 had a normalized deepening rate of 7.2 mb/12hr. The remaining seven storms underwent little deepening with one storm occurring late in the season undergoing no deepening. The lack of significant storms prevents any statistically significant analyses from being performed, but a brief overview of ASCII and I-ASCII performance follows. Multiple ASCII/I-ASCII forecasts were issued at different times prior to each event; however, the forecast for 12-hour deepening was rarely changed by more than 1 mb/12hr between forecasts. Therefore, for simplicity, only the most recent forecast is considered for each event, typically providing about six-hours lead time prior to the event.

Table 3.1 summarizes the ASCII, I-ASCII, and Eta forecasts for each event compared
against the observed deepening rate. These results are also shown in Fig. 3.5. Both ASCII and I-ASCII over-predicted deepening rates, although ASCII was more accurate than I-ASCII for each event. The error associated with both methods is shown in Table 3.2 along side the predicted deepening rate by the NCEP Eta model. Fig. 3.6 displays the errors graphically. The average of the absolute value of ASCII error was 5.6 mb/12hr as compared to I-ASCII which had an average error of 12.9 mb/12hr. Neither ASCII method was able to outperform the formerly operational Eta whose average error was only 1.8 mb/12hr.

The excessive error generated by I-ASCII is not a result of the PSBI computation, as it uses the same PSBI as used in the original ASCII forecast. Instead, the error is a result of the vorticity forecasts being used to classify the upper-level forcing associated with each event. Table 3.1 shows the values for 500-mb absolute vorticity used for each event. It should be noted that the vorticity for events occurring after March 1 were estimated manually as the 190-km GFS grid was no longer available to the ILM NWS which created an error in the automated script. It is immediately obvious that every event was classified as a strong event, with vorticity far exceeding the strong threshold of $19 \times 10^{-5} \text{ s}^{-1}$ for most cases.

There are a couple of reasons for the excessive vorticity forecasts. The most obvious is the difference in grid spacing between the $2.5^\circ \times 2.5^\circ$ NCEP-NCAR reanalysis datasets from which vorticity was obtained for the I-ASCII regression and the 190-km GFS grid used for the operational ASCII products. The coarser grid spacing of the NCEP-NCAR dataset results in a reduction of detail and overall magnitude of absolute vorticity values. Finer grid resolution means more detail is retained in the analysis, and the more difficult it is to associate any one vorticity maxima to the surface low. In reality, it is the net effect of
multiple maxes that is having effect on storm development, which may be why the coarser resolution dataset does a better job performing in this case. Another possible cause of error is the method of obtaining the vorticity. The original method used by Jacobs et al. (2005) was to use the closest vorticity maximum to the surface low, while the automated operational method simply searches for the highest value of vorticity within 1000-km upstream of the surface low.

3.4 Suggested Improvements

3.4.1 Grid Interpolation Method

One potential solution for obtaining vorticity values more in-line with the NCEP-NCAR reanalysis data is to artificially reduce the resolution of the GFS grid by interpolating the data to the NCEP-NCAR grid via a graphical analysis program such as GEMPAK. This method will still provide greater detail than the actual NCEP-NCAR reanalysis because the data were originally obtained on a higher resolution grid, but it is hoped that the reduction in resolution will improve the I-ASCII forecast by bringing vorticity values more in line with the NCEP-NCAR reanalysis. The vorticity values from the interpolated GFS grid are shown in Table 3.3. Despite the interpolation, the vorticity values for seven of the ten cases remain in the strongly forced category, with only two cases being reduced enough to fall into the moderately forced category. Ultimately, this makes very little difference in the actual I-ASCII forecast. A comparison of the new I-ASCII and the original forecasts along with the updated errors are also shown in Table 3.3. Overall, the average I-ASCII error is reduced to
12.0 mb/12hr, an improvement of just 7.3%. The reclassified cases are an improvement, but still fall short of the ASCII and Eta forecasted deepening.

3.4.2 Correlation Method

Another possible improvement to the vorticity forecast is to correlate the values of vorticity from the NCEP-NCAR reanalysis to values from another grid. In this case, values of absolute vorticity for 38 cases between 2002 and 2003 were obtained both from the NCEP-NCAR reanalysis datasets and the 32-km NARR dataset. The 190-km GFS is not readily available from any archived data sources, so it could not be used to perform the correlation. The dataset was composed of a total of nine strongly forced cases, 18 moderately forced cases, and 11 weakly forced cases as classified from the NCEP-NCAR reanalysis. The scatter plot of data is shown in Fig. 3.7 along with the regression equation. The regression coefficient is 0.54 suggesting a modest correlation.

To test whether the correlation can be used to improve the vorticity forecast, the largest 500-mb absolute vorticity within 1000-km of the center of the ASCII domain were obtained for the nine 2005-2006 cases from NARR analyses. The regression equation was used to approximate the corresponding value of vorticity on the NCEP-NCAR grid. The estimated values for vorticity, shown in Table 3.4, all were reduced to the moderately forced category. The I-ASCII forecasts were updated using the new vorticity values and compared to the original forecast. The results are shown in Table 3.4. The average error was reduced by 43.4% to 7.3 mb/12hr, which while an improvement to the operational I-ASCII forecast, is still above that of ASCII. However, this new I-ASCII method does outperform ASCII on
three of the nine forecasts.

Seeking a further method of improvement using this correlation method, the way the values of absolute vorticity are obtained from the NARR data is examined. While the original method of Jacobs et al (2005) was to choose the closest vorticity maximum to the surface low, when examining the high resolution NARR data, it becomes apparent that the nearest vorticity maximum may not always be having the greatest influence over the surface low. This is illustrated in Fig. 3.8, where a stronger maximum over western Kentucky may be having a greater influence on the surface low than the closest vorticity maximum co-located with the surface low. Therefore, the absolute vorticity for the 38 cases were obtained again, but this time the largest maximum within 1000-km of the surface low was used as opposed to the closest maximum. This also matches the method employed by the ILM NWS operational method.

The new scatter plot is shown in Fig. 3.9 along with the regression equation and correlation coefficient. The correlation coefficient has been improved to 0.69. Table 3.5 shows the new estimated vorticity values, all of which remain categorized as moderately forced except for one strongly and two weakly forced cases. When the I-ASCII forecast is generated based on the new values, there is still an improvement over the original operational method, but no improvement over the previous method, which used the nearest vorticity maximum to compute the correlation.

3.4.3 ijskip Method

The ijskip variable in GEMPAK allows the user to effectively reduce the grid spacing
of an analysis by skipping grid points in the plotting procedure. Unlike the interpolation method where values at the skipped grid points are averaged and plotting, the ijskip command skips grid points completely so that all information about the value of parameters at that point are lost. Setting ijskip to one plots every other grid point, setting the variable to two plots every third grid point, and so on.

To examine the impact of employing the ijskip variable, the 500 mb absolute vorticity pattern from the NCEP/NCAR analysis at 12Z Dec 25, 2002 is compared to the NARR analysis at the same time using different values for ijskip. The results are shown in Fig. 3.10. Fig. 3.10a shows the NCEP/NCAR analysis, while Fig. 3.10b-e shows the NARR analysis with differing values for ijskip. The original NARR analysis is shown in Fig. 3.10b. Note the value of the largest vorticity maximum approaches $28 \times 10^{-5} \text{ s}^{-1}$ while the largest value for the NCEP/NCAR analysis is less than $19 \times 10^{-5} \text{ s}^{-1}$. As ijskip is increased to three (Fig. 3.10c) note the significant loss in detail, with a single vorticity maximum analyzed. When ijskip is set equal to six (Fig. 3.10d) the analysis very closely resembles the NCEP/NCAR analysis in Fig. 3.10a, with the value of the vorticity maximum between $15 \times 10^{-5} \text{ s}^{-1}$ and $19 \times 10^{-5} \text{ s}^{-1}$. The aerial extent of the vorticity maximum is reduced when ijskip is increased to nine (Fig. 3.10e). In this particular case, an ijskip value of six applied to the NARR analysis seems most effective in helping to replicate the NCEP/NCAR analysis with the NARR data. In theory however, an ijskip value of eight may be more realistic as it effectively reduces the grid spacing of the 32-km NARR analysis to 288-km which is comparable to the NCEP/NCAR analysis which has $2.5^\circ$ (278-km) grid spacing. On the other hand, an ijskip value of six which reduces the analysis to 224-km. After testing several
other cases (not shown) it was found that an *ijskip* value of six is indeed the most effective, and is used for the below reclassification.

The nine cases from 2005-2006 were reclassified using NARR 500-mb absolute vorticity analyses with *ijskip* set to six. Results of this reclassification are shown in Table 3.6. All the cases are reclassified as moderately forced or weakly forced, which reduces the average error to 5.9 mb/12hr (an improvement of 54%), which is more in line with the original ASCII method. And if *ijskip* is reduced to nine (Table 3.7), the average error is reduced to 3.8 mb/12hr (a 71% reduction in error), which is much closer to the average error of the operational Eta model and an improvement upon the original ASCII. However, as was seen in Fig. 3.10e., setting *ijskip* to nine may result in an analysis which reduces the resolution of the NARR analysis below that of the NCEP/NCAR analysis. In order to determine the best value of *ijskip*, a larger sample of cases needs to be examined. But the *ijskip* method does demonstrate the largest improvement of the three methods discussed in this section, with a potential reduction in error of 71%.

### 3.5 Conclusions

Despite the lack of cases for the 2005-2006 winter season, some useful conclusions can be drawn from this analysis. First, the 190-km GFS grid is not coarse enough to obtain 500-mb absolute vorticity values in line with those obtained from the 2.5° × 2.5° NCEP-NCAR reanalysis. Due to the added detail on the 190-km grid, the majority of cases will be classified as strongly forced. While typically in meteorology, added resolution is considered an improvement, with respect to I-ASCII it actually serves to worsen the forecast. The reason for this is two-fold. First, the grid on which I-ASCII was originally developed offered
only coarse detail. The upper-level forcing was classified using values obtained from this coarse grid. Classifying the same vorticity maxima using a grid with finer grid spacing will result in greater resolution and values that are skewed higher. Another reason for the worsening forecast with increased resolution is that the processes occurring in response to vorticity may only be operating on a broad scale. In other words, the net effect of a particular vorticity maximum may be best captured when looking at the broader picture (coarser resolution) as opposed to fine detail (greater resolution). It is the net effect of the absolute vorticity pattern that is important in the development of the surface low, and the net effect is best captured by a coarse grid spacing which smooths out less important mesoscale features.

Three quick fix methods to resolve the problem with the classification of upper-level forcing were tested: interpolating the GFS grid to the NCEP-NCAR grid; correlating vorticity values from the NCEP-NCAR reanalysis to vorticity values from a high resolution grid; and employing the ijskip variable in GEMPAK. The ijskip method proves to generate the most improvement to the operational I-ASCII method, but still cannot outperform the operation Eta model. The performance of the ijskip method is dependant on the value of ijskip used in the analyses. Additional research is needed to determine the best value for ijskip. However, it appears unlikely that I-ASCII could outperform the Eta model using this method.

Even if one of the suggested improvements was able to bring the I-ASCII forecasts more in line with the Eta forecast, there is still an additional uncertainty being introduced to the I-ASCII forecast. The only sure-fire way to resolve the problems with I-ASCII is to recompute the regression equations by using absolute vorticity values obtained on a finer grid scale. This method, however, also has its problems, in that the higher the resolution the more
noisy the absolute vorticity analysis. It can be difficult to identify the vorticity maximum having the most influence on development, and the method becomes very subjective, which will make it difficult to automate the process in the future. It would be beneficial to investigate a more objective approach, perhaps by introducing a spatial average of vorticity or vorticity advection to estimate upper-level forcing. Estimating upper-level forcing using Q-vectors may also prove to be a more accurate approach. If I-ASCII can be altered to be less dependent on grid resolution and the subjectivity associated with it be removed, it could be a potentially useful operational tool. It may be most useful as a real-time analysis tool to make forecasters aware of the deepening potential of storms moving into or near the ASCII domain, as opposed to a long-term forecasting tool for which it may offer little improvement over current operational models.
ATLANTIC SURFACE CYCLONE INTENSIFICATION INDEX (ASCII)

THE BOMB INDEX IS INDETERMINATE.

THE PRE-STORM BAROCLINIC INDEX IS 1.5 DEGREES C/10 KM.

A STORM IN THE DOMAIN IS FORECAST TO DEEPEN AT A RATE OF 10.3 MB/12 HOURS.

FORECAST VALID UP TO 48 HOURS FROM 14Z 02/10/2006.

ASCII WITH DYNAMICS (EXPERIMENTAL) USING 12Z 180 KM GFS

Forecast Hour: 0  GFS ABS VORT USED: 30.90  Forecast Deepening: 20.19
Forecast Hour: 6  GFS ABS VORT USED: 28.60  Forecast Deepening: 20.19
Forecast Hour: 12 GFS ABS VORT USED: 30.40  Forecast Deepening: 20.19
Forecast Hour: 18 GFS ABS VORT USED: 37.90  Forecast Deepening: 20.19
Forecast Hour: 24 GFS ABS VORT USED: 42.00  Forecast Deepening: 20.19
Forecast Hour: 30 GFS ABS VORT USED: 32.20  Forecast Deepening: 20.19
Forecast Hour: 36 GFS ABS VORT USED: 28.40  Forecast Deepening: 20.19

FORECAST VALID UP TO 48 HOURS FROM 14Z 02/10/2006.

FORECAST UPDATED: Sat Feb 11 17:26:06 GMT 2006

FOR MORE INFORMATION SEE:

http://www.erh.noaa.gov/ilm/science/ascii/ascfcst.html

Figure 3.1. An example of an operational ASCII product from 11 February 2006. The product displays both the original ASCII forecast and the updated I-ASCII forecast.
Figure 3.2. Monthly time series of the SOI (a) and NAO (b) indices from January 2005 through March 2006.
Figure 3.3. 500-mb geopotential Height anomalies (m) from December to March 2006.
Figure 3.4. Temperature (top) and precipitation (bottom) anomalies for December 2005 through March 2006. Temperature anomalies are in °F and precipitation anomalies are in inches.
Table 3.1. A list of ASCII storms with the forecasted deepening rates from ASCII, I-ASCII, and the Eta model as compared to the observed deepening rate.

<table>
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<tr>
<th>Date</th>
<th>Julian Day</th>
<th>Original Vorticity ($\times 10^{-5}$ s$^{-1}$)</th>
<th>Upper-Level Forcing Classification</th>
<th>ASCII Forecast (mb/12hr)</th>
<th>I-ASCII Forecast (mb/12hr)</th>
<th>Eta Forecast (mb/12hr)</th>
<th>Observed Deepening (mb/12hr)</th>
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</thead>
<tbody>
<tr>
<td>31-Dec-05</td>
<td>365</td>
<td>31.6</td>
<td>Strong</td>
<td>12.2</td>
<td>24.7</td>
<td>4</td>
<td>2.0</td>
</tr>
<tr>
<td>3-Jan-06</td>
<td>3</td>
<td>37.5</td>
<td>Strong</td>
<td>11.9</td>
<td>23.2</td>
<td>0</td>
<td>3.0</td>
</tr>
<tr>
<td>6-Jan-06</td>
<td>6</td>
<td>36.4</td>
<td>Strong</td>
<td>8.9</td>
<td>17.2</td>
<td>6.7</td>
<td>4.0</td>
</tr>
<tr>
<td>11-Feb-06</td>
<td>42</td>
<td>28.6</td>
<td>Strong</td>
<td>10.3</td>
<td>20.2</td>
<td>11</td>
<td>12.0</td>
</tr>
<tr>
<td>6-Mar-06</td>
<td>65</td>
<td>20</td>
<td>Strong</td>
<td>9</td>
<td>17.2</td>
<td>5.3</td>
<td>3.0</td>
</tr>
<tr>
<td>17-Mar-06</td>
<td>76</td>
<td>30</td>
<td>Strong</td>
<td>5.2</td>
<td>8.1</td>
<td>4</td>
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<tr>
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<td>Strong</td>
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<td>15.7</td>
<td>9</td>
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<tr>
<td>28-Mar-06</td>
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<td>Strong</td>
<td>8.3</td>
<td>15.7</td>
<td>0</td>
<td>0.0</td>
</tr>
</tbody>
</table>

Figure 3.5. A time series of the forecasted deepening rate for ASCII (dark blue), I-ASCII (blue) and Eta model (green) compared to the observed deepening rate (red).
Table 3.2. A list of ASCII storms and the associated error from the ASCII, I-ASCII, and the Eta model forecasts.

<table>
<thead>
<tr>
<th>Date</th>
<th>Julian Day</th>
<th>ASCII Error (mb/12hr)</th>
<th>I-ASCII Error (mb/12hr)</th>
<th>Eta Error (mb/12hr)</th>
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</thead>
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<td>31-Dec-05</td>
<td>365</td>
<td>10.2</td>
<td>22.7</td>
<td>2.0</td>
</tr>
<tr>
<td>3-Jan-06</td>
<td>3</td>
<td>9.9</td>
<td>20.2</td>
<td>-3.0</td>
</tr>
<tr>
<td>6-Jan-06</td>
<td>6</td>
<td>4.9</td>
<td>13.2</td>
<td>2.7</td>
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<td>11-Feb-06</td>
<td>42</td>
<td>-1.7</td>
<td>8.2</td>
<td>-1.0</td>
</tr>
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<td>65</td>
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<td>2.3</td>
</tr>
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<td>2.2</td>
<td>5.1</td>
<td>1.0</td>
</tr>
<tr>
<td>21-Mar-06</td>
<td>80</td>
<td>4.5</td>
<td>6.9</td>
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<td>24-Mar-06</td>
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Figure 3.6. Error bars showing the error associated with the ASCII, I-ASCII, and Eta model forecasts.
Table 3.3. Updated upper-level vorticity values, classifications, I-ASCII forecasts, and I-ASCII error based on the grid interpolation method. The original I-ASCII forecasts and errors are shown for comparison. Average I-ASCII error is listed in the last column.

<table>
<thead>
<tr>
<th>Case</th>
<th>Reclassified Vorticity ($\times 10^{-5}$ s$^{-1}$)</th>
<th>Upper-Level Forcing Classification</th>
<th>Original I-ASCII Forecast (mb/12hr)</th>
<th>Reclassified I-ASCII Forecast (mb/12hr)</th>
<th>Original I-ASCII Error (mb/12hr)</th>
<th>Reclassified I-ASCII Error (mb/12hr)</th>
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<td>13.2</td>
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</tr>
<tr>
<td>11-Feb-06</td>
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<td>Strong</td>
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<td>20.2</td>
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</tr>
<tr>
<td>6-Mar-06</td>
<td>20</td>
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<td>17.2</td>
<td>17.2</td>
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<tr>
<td>17-Mar-06</td>
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<td>14.1</td>
<td>6.9</td>
<td>6.9</td>
</tr>
<tr>
<td>24-Mar-06</td>
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<td>10.1</td>
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<td>4.8</td>
</tr>
<tr>
<td>28-Mar-06</td>
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<tr>
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<td>12.9</td>
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</table>
Figure 3.7. Scatterplot of the 500-mb absolute vorticity as obtained from the NCEP and NARR analyses using the closest vorticity maximum to the surface low. The best fit line is displayed along with the associated regression equation and correlation coefficient.

Table 3.4. Updated vorticity values, I-ASCII forecasts, and I-ASCII errors compared to the original values. Updated vorticity values are based on the regression equation in Fig 4.8.

<table>
<thead>
<tr>
<th>Case</th>
<th>Estimated Vorticity ($\times10^5$ s$^{-1}$)</th>
<th>Upper-Level Forcing Classification</th>
<th>Original I-ASCII Forecast (mb/12hr)</th>
<th>Updated I-ASCII Forecast (mb/12hr)</th>
<th>Original I-ASCII Error (mb/12hr)</th>
<th>Updated I-ASCII Error (mb/12hr)</th>
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<td>24.7</td>
<td>16.3</td>
<td>22.7</td>
<td>14.3</td>
</tr>
<tr>
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<td>18.1</td>
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<td>23.2</td>
<td>15.2</td>
<td>20.2</td>
<td>12.2</td>
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<tr>
<td>6-Jan-06</td>
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<td>17.2</td>
<td>11.1</td>
<td>13.2</td>
<td>7.1</td>
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<tr>
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<td>8.2</td>
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<td>1.9</td>
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<td>21-Mar-06</td>
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<td>9.1</td>
<td>6.9</td>
<td>6.1</td>
</tr>
<tr>
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<td>Moderate</td>
<td>15.7</td>
<td>10.1</td>
<td>10.3</td>
<td>4.8</td>
</tr>
<tr>
<td>28-Mar-06</td>
<td>17.8</td>
<td>Moderate</td>
<td>15.7</td>
<td>10.1</td>
<td>15.7</td>
<td>10.1</td>
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<td>Average</td>
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<td></td>
<td></td>
<td></td>
<td>12.9</td>
<td>7.3</td>
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Figure 3.8. NARR plot for 15Z Dec 5, 2003 showing sea-level pressure in mb (solid) and 500-mb absolute vorticity in $\times 10^{-5}$ s$^{-1}$ (shaded).
**Figure 3.9.** Scatterplot of the 500-mb absolute vorticity as obtained from the NCEP and NARR analyses using the strongest vorticity maximum within 1000 km of the surface low. The best fit line is displayed along with the associated regression equation and correlation coefficient.

**Table 3.5.** Updated upper-level vorticity values, classifications, I-ASCII forecasts, and I-ASCII error based on the correlation method. Updated vorticity values are computed from the regression equation in Fig 4.9. The original I-ASCII forecasts and errors are shown for comparison. Average I-ASCII error is listed in the last column.
Figure 3.8. A comparison of 500-mb absolute vorticity (shaded) and heights (solid) for the NCEP-NCAR reanalysis data and NARR data using different values for ijskip in GEMPAK. Analysis time is 12Z Dec 25, 2002. Units for absolute vorticity are $\times 10^{-5} \text{ s}^{-1}$. Plots shown are the NCEP/NCAR reanalysis (a), NARR analysis with ijskip = 0 (b), NARR analysis with ijskip = 3 (c), NARR analysis with ijskip = 6 (d), and NARR analysis with ijskip = 9 (e).
### Table 3.6. Updated upper-level vorticity values, classifications, I-ASCII forecasts, and I-ASCII error based on the ijskip method with ijskip=6. The original I-ASCII forecasts and errors are shown for comparison. Average I-ASCII error is listed in the last column.

**IJSKIP = 6**

<table>
<thead>
<tr>
<th>Case</th>
<th>Reclassified Vorticity (×10⁻⁵ s⁻¹)</th>
<th>Upper-Level Forcing Classification</th>
<th>Original I-ASCII Forecast (mb/12hr)</th>
<th>Updated I-ASCII Forecast (mb/12hr)</th>
<th>Original I-ASCII Error (mb/12hr)</th>
<th>I-ASCII Error (mb/12hr)</th>
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</thead>
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<td>16.3</td>
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<td>14.3</td>
</tr>
<tr>
<td>3-Jan-06</td>
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<td>23.2</td>
<td>15.2</td>
<td>20.2</td>
<td>12.2</td>
</tr>
<tr>
<td>6-Jan-06</td>
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<td>Moderate</td>
<td>17.2</td>
<td>11.2</td>
<td>13.2</td>
<td>7.1</td>
</tr>
<tr>
<td>11-Feb-06</td>
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<td>8.2</td>
<td>1.2</td>
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<td>8.1</td>
</tr>
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<td>24-Mar-06</td>
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</tr>
<tr>
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<td>6.5</td>
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<tr>
<td>Average</td>
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<td></td>
<td>12.9</td>
<td>5.9</td>
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</table>

### Table 3.7. Updated upper-level vorticity values, classifications, I-ASCII forecasts, and I-ASCII error based on the ijskip method with ijskip=9. The original I-ASCII forecasts and errors are shown for comparison. Average I-ASCII error is listed in the last column.

**IJSKIP = 9**

<table>
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<tr>
<th>Case</th>
<th>Reclassified Vorticity (×10⁻⁵ s⁻¹)</th>
<th>Upper-Level Forcing Classification</th>
<th>Original I-ASCII Forecast (mb/12hr)</th>
<th>Updated I-ASCII Forecast (mb/12hr)</th>
<th>Original I-ASCII Error (mb/12hr)</th>
<th>I-ASCII Error (mb/12hr)</th>
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</thead>
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<td>31-Dec-05</td>
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</tr>
<tr>
<td>3-Jan-06</td>
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<td>Weak</td>
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<td>9.6</td>
<td>20.2</td>
<td>6.6</td>
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<td>6-Jan-06</td>
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<td>Weak</td>
<td>17.2</td>
<td>7.1</td>
<td>13.2</td>
<td>3.1</td>
</tr>
<tr>
<td>11-Feb-06</td>
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<td>Moderate</td>
<td>20.2</td>
<td>13.2</td>
<td>8.2</td>
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<td>8.1</td>
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<tr>
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<td>3.5</td>
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<td>0.5</td>
</tr>
<tr>
<td>21-Mar-06</td>
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<td>5.9</td>
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<td>24-Mar-06</td>
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<td>6.5</td>
<td>10.3</td>
<td>1.2</td>
</tr>
<tr>
<td>28-Mar-06</td>
<td>11</td>
<td>Weak</td>
<td>15.7</td>
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<tr>
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<td>12.9</td>
<td>3.8</td>
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</tbody>
</table>
CHAPTER 4: THE REPRESENTATION OF SURFACE AND UPPER-LEVEL PROCESSES IN ASCII

4.1 Introduction

Two case studies are presented in this chapter in an effort to diagnose for some of the errors in deepening rate encountered with I-ASCII. The storms on 25 December 2002 (hereafter referred to as D-02) and on 14 December 2003 (hereafter referred to as D-03) were chosen from 52 observed winter storms meeting ASCII criteria during the period 2002-2004. Both storms were simulated using the WRF model. The motivation for choosing these two cases was the sharp contrast in I-ASCII error. I-ASCII is used here due to its demonstrated improvement over the original ASCII.

Upper level forcing for both events was classified as strong, with the associated absolute vorticity value for each case being $19 \times 10^{-5}$ s$^{-1}$ (determined from the 2.5º NCEP-NCAR Reanalysis). Plugging the computed PSBI values (shown in Table 4.1) into the regression equation for strongly forced events ($y = -2.4551 + 15.094x$) yields an I-ASCII predicted deepening rate of 15.8 mb/12hr for D-02 and 28.4 mb/12hr for D-03. The observed deepening for D-02 was 16 mb/12hr, yielding an error of only -0.2 mb/12hr, or essentially a perfect ASCII forecast. On the other hand, the observed deepening for D-03 was 10 mb/12hr, an error of 18.4 mb/12hr, a significant I-ASCII bust. D-02 and D-03 have been plotted in Fig. 4.1 along with the I-ASCII dataset and corresponding regression lines. Note that the point representing D-02 lies exactly along the regression line for strongly forced events, while D-03 lies well off the line. In fact, had the upper level forcing been classified as weak for D-03, the I-ASCII predicted deepening would have been very accurate. For comparison, ASCII predicted a deepening rate of 9.1 mb/12hr for D-02 and 14.7 mb/12 hr for
D-03, yielding errors of 6.9 mb/12hr and 4.7 mb/12hr respectively. So I-ASCII outperformed ASCII for D-02, but not for D-03. However, given that I-ASCII has, in general, been shown to be an improvement on ASCII (Jacobs et al. 2005) it will be the focus of this chapter.

While the low-level baroclinicity and upper level forcing have been taken into account with I-ASCII, there are a number of additional factors which are investigated here. PSBI is essentially a measure of the surface temperature gradient off the Carolina coastline. However, a strong surface temperature gradient alone does not immediately result in a low-level temperature gradient and storm development. Rather, it creates a favorable area for storm development by promoting the potential for large values of sensible and latent heat flux which have been shown to be of importance in cases of rapid cyclogenesis (Bosart 1981; Bosart and Lin 1984; Fantini 1991; Kuo et al. 1991; Mak 1998; Reed and Albright 1986; Roebber 1989; Sanders and Gyakum 1986). It is this exchange of heat and moisture from the surface into the atmosphere that generates static instability in the lower atmosphere, which is realized with the assistance of an external forcing mechanism (Bosart 1981). The horizontal gradient of fluxes helps generate baroclinic instability, which promotes the conversion of potential energy into kinetic energy. In northwest flow, a sharp horizontal gradient in surface temperatures will allow cold continental air to reach the warm waters of the GS without significant modification and thereby maximize the temperature difference between surface and atmosphere and increasing sensible and latent heat fluxes. But in addition to a large vertical temperature gradient, a large wind speed is also necessary to maximize heat flux.

We hypothesize that the large error in D-03 is a result of light wind speeds which
would result in lower heat fluxes, less modification of the boundary layer, and a lesser influence of surface baroclinicity on the development of the surface low. If this is true, then an inclusion of heat flux or heat flux gradient in I-ASCII may help improve overall performance. In addition, the actual location of the surface low with respect to land and the largest heat flux gradients may be vital to development, with deepening rates being greatest for storms nearest to the gradient. Problems with the classification of upper-level forcing are also anticipated given the sensitivity to grid spacing demonstrated in the previous chapter. The timing and pattern of upper-level vorticity may be important in explaining I-ASCII error. Lastly, the warm-core eddy is expected to have an impact in the heat flux patterns but not a significant impact on I-ASCII performance despite the eddy having an impact on the temperature gradient which may not be well represented by the PSBI.

In this chapter, a synoptic overview of the 25 December 2002 and 14 December 2003 case studies will be presented. Some information about the model setup and experimental design will follow. Several model output fields will be examined and discussed including: sea-level pressure (SLP), 500-mb heights and absolute vorticity, surface sensible and latent heat fluxes, and near-surface (10-m) convergence. Selected cross sections will also be compared to diagnose any differences in the vertical structure in the boundary layer for each case.

4.2 Synopsis of 25 December 2002 and 14 December 2003 Case Studies

4.2.1 25 December 2002

On 25 December 2002, a rapidly strengthening low pressure system developed near
the coast of South Carolina and moved northward along the U.S. East Coast. According to
the National Climatic Data Center (NCDC) Storm Events Database, the same storm system
brought severe weather, tornadoes, and flooding rains to the Gulf Coast States on 24
December 2002. Once the coastal low developed on 25 December 2002, the biggest impacts
from the storm were heavy snowfall and gusty winds. The heaviest snowfall occurred in an
area extending from portions of Pennsylvania northeastward through southern Maine. Public
Information Statements issued by the National Weather Service (NWS) reported widespread
amounts in excess of 15-cm, with many areas receiving over 30-cm. Some selected snowfall
amounts are highlighted in Table 4.2. In addition to heavy snow, gusty winds associated
with the system caused sporadic damage and power outages from Georgia to Maine. The
combination of wind and snow resulted in blizzard conditions along the coast of Maine. One
fatality was attributed to the storm in western North Carolina. Wind gusts in the area were
estimated to be near 110-kph. Winds reportedly gusted to 80-kph across high elevations of
upstate New York. Official recorded wind gusts include: 105-kph at Milton, MA; 102-kph at
Belmar, NJ; and 90-kph at Cape May, NJ.

Surface analyses showing the evolution of sea-level pressure, 2-m temperature, and
surface station plots every six hours for D-02 are shown in Fig. 4.2 (00Z Dec 25-18Z Dec
25, 2002). Prior to the storm moving into the ASCII domain, a 1016-mb high pressure
system was situated over southern New England. A weak cold-air damming signature,
evident by the ridge of high pressure along the spine of the Appalachian Mountains, is seen
extending down into northern Georgia. A 1002-mb low pressure system moved out of the
Gulf of Mexico and west of the mountains and dissipated as the upper-level support (Fig.
4.3) moved towards the Southeast Coast. A precursor disturbance developed along an old frontal boundary off the Carolina coastline and moved northeastward and dissipated. The main ASCII low developed behind the precursor disturbance with the assistance of the arriving upper-level energy. The associated upper-level trough became negatively tilted around 12Z Dec 25, 2002 as the ASCII low began to deepen rapidly and move up the U.S. East Coast. The 500-mb absolute vorticity associated with the system closely trailed the surface low with the magnitude of the closest maximum between $28-32 \times 10^{-5} \text{ s}^{-1}$ at 18Z Dec 25, 2002.

From 1-mb contoured plots of sea-level pressure at 3-hour intervals (not shown) it is deduced that the surface low entered the southwestern portion of the ASCII domain around 0900Z Dec 25, 2002 at 998-mb. The storm moved north-northeastward exiting the northern portion of the domain around 15Z Dec 25, 2002 after deepening 8-mb to 990-mb. The normalized deepening rate was 16 mb/12hr. After exiting the domain, the storm continued to deepen rapidly ultimately reaching a minimum sea-level pressure of 973-mb at 03Z Dec 26, 2002 near Cape Cod, MA. Of the six hours spent in the ASCII domain, the surface low spent approximately half of that time over land, and half over water.

Fig. 4.4 shows base radar reflectivity at six hour increments from 00Z 25 Dec, 2002 through 18Z 25 Dec, 2002. Initially, there was an area of precipitation across Illinois and Indiana associated with the dissipating surface low in Tennessee. Another area of overrunning precipitation associated with the precursor disturbance was affecting portions of Maryland, New Jersey, and Pennsylvania. This area of precipitation moved offshore as the area of precipitation to the west moved east into Ohio and weakened. Precipitation began to
regenerate over northern Virginia up through Pennsylvania as the main ASCII low began
gathering strength off the coast. The shield of precipitation moved northeastward into New
England along the northwest side of the low.

The 24-hour analyzed precipitation computed from the NARR dataset for Dec 25,
2002 is shown in Fig. 4.5. The heaviest precipitation was associated with scattered
convection in the southeastern states and the solid area of overrunning stratiform
precipitation north and west of the main low. A large swath of amounts in excess of 10-mm
was observed down the entire eastern seaboard with some localized areas receiving over 25-
mm. Note the lighter amounts over Ohio where the precipitation field weakened before
redeveloping as the ASCII low took over.

It should be noted that the NARR precipitation is not a true analysis of observed
precipitation totals. Gauge observations are assimilated into NARR by converting it to latent
heat which is used by the Eta regional model to predict the precipitation field. It is generally
very accurate over land, but less accurate over the ocean where the data are scarce and data
assimilation is not performed north of 42.5ºN (Mesinger et al. 2006). Due to discrepancies in
observation times, the 24-hour precipitation cannot be easily validated against real-time
observations, but NARR total storm precipitation from D-02 is verified with observations in
Fig. 4.15.

High resolution sea-surface temperature (SST) composites generated from
CoastWatch data show the approximate position and orientation of the Gulf Stream (GS) on
00Z Dec 24, 2002 (Fig. 4.6). A warm-core eddy was present southeast of Wilmington, NC.
The average distance of the Gulf Stream Front (GSF) east of ILM and HSE was 122-km,
with an average temperature at the GSF of 24°C as determined from 1.47-km CoastWatch SST analyses. Coastal waters off the North and South Carolina averaged around 12°C. SSTs over the core of the GS averaged about 26°C south of Wilmington and 24-25°C north of Wilmington.

4.2.2 14 December 2003

On 14 December 2003, a low pressure system moved out of the Gulf of Mexico, across the Florida panhandle and began to strengthen as it reached the South Carolina Coast. The storm continued to strengthen as it moved north along the eastern seaboard. Prior to the storm reaching the East Coast, impacts were minimal. Wintry precipitation preceding the storm was reported in the higher elevations of North Carolina, Virginia, and West Virginia on 13 December 2003. Snow, sleet, and freezing rain continued to be the primary impact of the storm on 14-15 December 2003 as the storm moved up the East Coast, with gusty winds also being reported along the coast. Frozen precipitation fell from North Carolina through Maine and as far west as Ohio, with the hardest hit area extending from the mountains of West Virginia through eastern New York and into Maine. Many areas received in excess of 25-cm of snowfall with some locales exceeding 50-cm. Selected snowfall amounts are displayed in Table 4.3. At least one death was attributed to the storm in Virginia. Ice accumulations of 0.6-1.25-cm resulted in downed trees and power lines in western North Carolina. Parts of New Jersey received over 2.5-cm of rain on top of already saturated soils resulted in sporadic flooding problems. A quick summary of wind gusts associated with the storm are shown in Table 4.4.
Sea-level pressure, 2-m temperature, and surface station plot evolution for D-03 are shown in Fig. 4.7 (06Z Dec 14- 00Z Dec 15, 2003). A strong 1032-mb high was situated over New England with a cold-air damming signature evident down through northern Georgia. A 1014-mb low, initially over the Gulf of Mexico, moved northeast across northern Florida. The low began to strengthen slowly as it reached the Carolina coastline and moved north along the East Coast. Unlike D-02 case, the upper-level trough does not become negatively tilted (Fig. 4.8). In addition, 500-mb absolute vorticity associated with the system remains spread out within the upper-level trough and several hundred kilometers away from the surface low. The value of the nearest vorticity maximum lies between 24-28 ×10⁻⁵ s⁻¹ by 00Z Dec 15, 2003. However, strong diffluence observed over the surface low suggests strong upper-level forcing has developed by the time the storm is exiting the ASCII domain (Fig. 4.7d).

Sea-level pressure plots contoured at 1-mb intervals (not shown) reveal that the surface low entered the southwest corner of the ASCII domain sometime before 15Z Dec 14, 2003 at about 1009-mb. The low moved predominately north and exited the domain shortly after 21Z Dec 14, 2003 after deepening 5-mb to 1004-mb. The normalized deepening rate was about 11mb/12hr. After exiting the domain, the low continued to move northward along the coast eventually bottoming out around 982-mb near the southern coast of Nova Scotia at 00Z Dec 16, 2003. The surface low spent about six hours in the ASCII domain, the majority of which was spent over land or the shallow waters of the Chesapeake Bay. In other words, the December 2003 case spent less time over open ocean waters than the December 2002 case, which may be responsible for some of the observed I-ASCII error.
Base radar reflectivity at six hourly increments from 06Z 14 Dec, 2003 through 00Z 15 Dec, 2003 is shown in Fig. 4.9. Widespread precipitation associated with the system initially covering much of the eastern US gradually pushed northeast throughout the period. The area of precipitation initially covering portions of Tennessee, Kentucky, Indiana, and Ohio weakened as it moved up into Pennsylvania. The heaviest radar returns became oriented along the stalled frontal boundary near the East Coast and the developing ASCII low. As the low strengthened, overrunning precipitation redeveloped over Pennsylvania northward into New England. 24-hour analyzed precipitation from NARR is shown in Fig. 4.10. Convective precipitation resulted in heavy rainfall in Florida, North Carolina, and Virginia while stratiform precipitation northwest of the ASCII low produced heavy amounts in the northeast. Precipitation amounts in excess of 10-mm were common with many areas exceeding 25-mm. NARR analyzed precipitation for the entire storm is verified against actual observations in Fig. 4.19.

A high resolution CoastWatch SST composite displayed in Fig. 4.11 shows the approximate position and orientation of the GS on 00Z Dec 13, 2003. In this case, no distinct warm-core eddy was present, although there was a prominent bulge in the GS east of Charleston, SC. The average distance of the Gulf Stream Front (GSF) east of ILM and HSE was 82-km, with an average temperature at the GSF of 23ºC determined from 1.47-km CoastWatch SST analyses. Coastal waters off the North and South Carolina averaged around 14ºC. SSTs over the core of the Gulf Stream averaged about 26ºC up to around Hatteras, NC.
4.3 Model Overview

Due to the lack of real-time heat flux measurements, reliable calculations, and the desire for increased resolution, model simulations are used to evaluate surface sensible and latent heat fluxes. Each case was simulated using version 2.1.1 of the WRF-ARW core with a 12-km outer grid covering the eastern US and western Atlantic Ocean (Fig. 2.3). To better capture Gulf Stream features and mesoscale processes occurring within and near the ASCII domain, a 4-km nested grid centered on eastern North Carolina was utilized. A two-way nest configuration was used to make use of better temporal resolution of boundary conditions for the nested domain and to allow feedback from the nested domain to the outer domain. The model contained 31 vertical \( \eta \) levels up to 100-mb, with 11 levels below 850-mb. A summary of the physics options selected are shown in Table 2.1. For a more an explanation of each option and reasoning behind their selection, refer to Chapter 2.

NARR data were used as initial and boundary conditions for each simulation. Boundary conditions were updated every three hours. For SSTs, 1.47-km CoastWatch data was sampled at 0.108° (approximately 12-km) grid spacing. A 10-day SST composite image (ending the day corresponding to model initialization) was generated for each case. The images were overlaid upon 0.5° global SST analyses provided by the National Center for Environmental Prediction (NCEP) to fill in any areas obscured by clouds for the entire 10-day period and to extend the image to fill the entire WRF domain. SSTs were held constant through the duration of the model runs.

Each case was initialized 48-hours prior to the movement of the closed low of interest into (or development within) the ASCII domain to allow for analysis of the pre-conditioning
period. This time was determined by examining NARR sea-level pressure (SLP) plots contoured at 1-mb intervals in three hour increments. The initialization times for D-02 and D-03 were 09Z Dec 23, 2002 and 15Z Dec 12, 2003 respectively. The simulations were 72-hours in length to capture the storms’ full evolution within the ASCII domain.

### 4.4 Model Verification

#### 4.4.1 25 December 2002

Fig. 4.12 shows WRF SLP output for D-02 for the four output times corresponding to the NARR SLP analyses in Fig. 4.2 (simulation hours 39, 45, 51, 57). The precursor disturbance observed in panels Fig 4.2a and Fig. 4.2b prior to the development of the main ASCII low is absent from the WRF simulation. This is likely a result of the surface high, observed over central NY in Fig. 4.12a, being simulated stronger and farther to the southeast, serving to suppress the low pressure from tracking up the Carolina coastline. It is possible the absence of the low is also just an artifact of grid spacing and SLP calculation differences between the NARR and WRF analysis. However, as the surface high is displaced, the simulated SLP pattern begins to resemble the observed by 06Z Dec 25, 2002. The simulated ASCII low develops and moves northward, following a slightly further inland track than what was observed (Fig. 4.12c,d). By 18Z Dec 25, 2002, the simulated 986-mb low is located near the Delaware Bay, very closely matching the strength and position of the NARR analyzed low at that time.

Comparing 1-mb contoured SLP plots for D-02 plotted at 3-hour increments (not shown), the simulated surface low moves into the ASCII domain at approximately the same
time as was observed in the NARR analysis, around 09Z Dec 25, 2002. The simulated storm deepens from 999-mb to 987-mb as it moves out of the domain six hours later, also matching the approximate time the NARR analyzed storm exited the ASCII domain. The normalized deepening rate is 24mb/12hr, exceeding the observed rate of 16mb/12hr. After exiting the domain, the simulated storm takes a farther east track than the observed storm, reaching a pressure of 973-mb well southeast of Cape Cod, MA by the end of the simulation.

Examining the simulated upper-level pattern (Fig. 4.13), WRF does a good job of capturing the trough location and tilt. The trough does appear slightly weaker, with the 534 dam contour not penetrating deep into the trough until 18Z Dec 25, 2002. Quantitatively, the 500-mb vorticity cannot be fairly compared to the analysis given the difference in resolution of the NARR dataset and WRF output. However, looking at vorticity qualitatively, it can be seen it is simulated well with the orientation and location of the strongest features matching up well.

Atmospheric soundings were plotted from the surface to 100-mb to verify the WRF model’s consistency with the NARR analysis in the vertical structure of the environment and the planetary-boundary layer (PBL). Fig. 4.14 shows a comparison of temperature and dewpoint as observed from the 00Z Dec 25, 2002 sounding at Morehead City, NC (MHX) and Charleston, SC (CHS) with the WRF simulated temperature and dewpoint at the same time (simulation hour 39). At both locations, there is complex structure within the lower troposphere with multiple inversion layers present. A moist layer is also present near the surface for both locations, extending to 800-mb at MHX and about 850-mb at CHS. A larger dewpoint depression is present through the mid-levels before another moist layer is
encountered above 500-mb. Much of the detail is lost in the WRF simulation, although the
general shape of each sounding is retained. The average temperature is simulated well with
the WRF simulation being generally within 5°C of the observed sounding. The comparison
of dewpoint is less successful. At both locations, WRF simulates a rather moist layer up 500-
mb, missing out on the drier mid-levels. For MHX, the simulated dewpoint deviates
significantly from the observed above 500 mb, with a 15-20°C dewpoint depression as
compared to the 2-4 °C depression observed.

Plots of analyzed 72-hour precipitation corresponding to the time of the D-02
simulation are compared to the WRF simulated precipitation totals for the duration of the
model run in Fig. 4.15. While WRF is not expected to perfectly simulate precipitation
distribution or amounts due to limitations in the convective parameterizations, the plots can
be used to determine whether the distributions and amounts are at least reasonable and help
to identify any possible convective feedback problems. The D-02 simulation does an
excellent job picking up on the overall distribution of precipitation. Amounts are generally
reasonable and comparable to the analysis, with a few obvious exceptions such eastern
Maryland up through Massachusetts.

4.4.2 14 December 2003

There are more noticeable differences in the D-03 simulation when compared to
observations. Looking at the simulated SLP pattern (Fig. 4.16) we find that the surface low
which moved into the Ohio Valley has already dissipated. Two separate closed surface lows
are analyzed, one of which already sits off the South Carolina Coast over the Gulf Stream.
This low crosses into the ASCII domain approximately around 09Z Dec 14, 2003 (not
shown) a full six hours earlier than the observed ASCII low crossed into the domain. The
second low, initially near the Florida panhandle, more closely resembles the observed ASCII
low in its trek across northern Florida and its time of arrival in the ASCII domain, but this
low eventually dissipates by 00Z Dec 15, 2003. At this time, the first low closely matches
the location and strength of the analyzed ASCII low, sitting near the Delmarva Peninsula at
1000-mb, about 4-mb deeper than the analyzed low.

The 1-mb contoured WRF SLP analyses (not shown) reveal a pressure of 1009-mb
for the simulated low as it crosses into the southwest corner of the ASCII domain around 09Z
Dec 14, 2003. By the time the low enters the domain is about six hours earlier than the time
of the analyzed low, the simulated low exits the ASCII domain around 00Z Dec 15, 2003
which is more in line with the analyzed low. By 00Z Dec 15, 2003, the simulated low
depens to 999-mb, yielding a normalized deepening rate of 8mb/12hr, shy of the observed
rate of 11mb/12hr. After exiting the domain, the simulated low follows a track similar to the
analyzed low, albeit slightly farther offshore, before reaching 988-mb at the end of the
simulation.

The simulated upper-level forcing for D-03 more closely matches the reanalysis than
the simulated surface features. The position and strength of the upper-level trough (Fig.
4.17) compares closely with the NARR analysis. As in the D-02 case, differences in
resolution between the NARR and WRF analyses prevent a fair comparison of the strength of
the 500-mb vorticity, although the overall pattern can be assessed. In general, the simulated
500-mb vorticity generally lines up with the axis of the trough as seen in the analysis,
although it does appear to cover a broader area.

Observed and modeled dewpoint and temperature were plotted from the surface to 100-mb at 00Z Dec 14, 2003 (simulation hour 33) for MHX and CSH (Fig. 4.18). The WRF simulation does a better job of picking up on the details of the atmospheric sounding for the D-03 simulation than the D-02 simulation due to the less complex structure of the temperature profile. An inversion layer is present at both MHX and CHS just above the surface, extending to near 900-mb. From there, temperature generally decreases with just a few small inversion layers, not nearly as pronounced as in D-02. The surface moist layer for both locations is deeper than D-02, with the layer extending to 700-mb at MHX and 500-mb at CHS. The simulated temperature soundings match the observed very well, generally being within 2°C of the observations. The simulated dewpoint soundings are also reasonable, both picking up on a dry layer between 700-mb and 500-mb. The simulation underestimates the dry layer while it is over done at CHS. The simulation also misses out on a more significant dry layer at CHS between 500-mb and 400-mb. Otherwise, the simulated soundings are much more reasonable for D-03 than D-02.

Fig. 4.19 shows the 72 hour simulated precipitation for D-03 compared against the NARR analyzed precipitation from the same time period. The simulation picks up the precipitation distribution reasonably well. The most noticeable problem is over the Mississippi River Valley region where a break in precipitation was observed. Precipitation amounts are also generally reasonable with two exceptions being the Mississippi River Valley region and eastern North Carolina.
4.4.3 Verification Summary

There are some noticeable differences between the WRF simulation and the analysis. In D-02, these differences are most noticeable when looking at the vertical structure of the atmosphere, where the differences are more pronounced for D-03 when examining horizontal parameter fields. Some of the differences are likely an artifact of the difference in resolution of the two grid files associated with each case. For instance, in D-03 where the simulated surface low crosses into the ASCII domain 6 hours earlier than the analysis, it is entirely possible that a closed low did actually form at the earlier time but was not analyzed in the NARR analysis. The normalized deepening rate simulated for D-02 is 8mb/12hr greater than observed, while for D-02 to the simulated deepening rate is 2mb/12hr less than observed. The difference between the predicted deepening rate (shown in Table 4.1) and the simulated is 8.2mb/12hr for D-02 and 20.4mb/12hr for D-03. Much of this disparity is likely a result of the decreased grid spacing in the WRF simulation as opposed to the NARR analysis.

Overall, the simulation results are sufficient for addressing the questions at hand. The WRF simulation of upper-level patterns in each case appears to have a good handle on key features. The sea-level pressure patterns, while not perfect in strength, do a good job with simulating the observed storm track and timing. And the general vertical structure of the atmosphere is captured, although some minor details such as the exact location and depth of various dry and moist layers are missed. While there is an 8mb/12hr disparity in the normalized deepening rate between the simulation and analysis for D-02, the error between the ASCII predicted deepening rate and the simulated rate is still lower for D-02 as compared to D-03 which will still allow the main objectives to be addressed.
4.5 Discussion

4.5.1 Sea-level Pressure Evolution

Fig. 4.12 shows the SLP and 2-m temperature at 39, 45, 51, and 57 hours into the D-02 simulation, matching the analysis times of Fig. 4.2. The storm begins with the ASCII low developing along an old frontal boundary as a low pressure system treks out of the Gulf of Mexico. The ASCII low actually forms inland across eastern South Carolina and North Carolina. The low enters the western side of the domain remaining primarily over land until reaching the southeast Virginia Coast, and spending about 6-hours in the ASCII domain. Ultimately, the simulated D-02 storm deepened at a rate of 12 mb/12hr.

SLP and 2-m temperatures at 39, 45, 51, and 57 hours into the D-03 simulation are shown in Fig. 4.6, with the analysis times matching that of Fig. 4.2. Like the D-02, a surface low moving out of the Gulf of Mexico transfers its energy to the developing ASCII low which forms along an old frontal boundary off of the East Coast. In the D-03 simulation the ASCII low forms farther south over the coastal waters of Florida and Georgia. The low enters the southwest corner of the ASCII domain and parallels the coastline, spending about three hours more over water than the D-02 low. A strong surface high to the north of the D-03 low impeded its motion; as a result, the surface low spent about 15-hours in the ASCII domain, nine hours more than the D-02 storm. The simulated deepening rate for D-03 was 8 mb/12hr.

While the two storms had similar origins, there were some significant differences in their overall evolution. It is a bit counter-intuitive to think that the D-02 storm would deepen more rapidly than the D-03 storm given its minimum time over water and within the ASCII
domain, displaced far from the SST gradient. From the comparison of SLP, it becomes quite apparent that the exact location of the surface low does not dictate the accuracy of an ASCII forecast. Or as mentioned earlier, the behavior of storms spending the majority of their time in the ASCII domain over land may not be fully accounted for by ASCII, and therefore ASCII may not be valid for such storms. In any case, there is little doubt that the surface low is influenced by factors beyond the area immediately under the surface low. It will be important therefore, to focus on the environment within the entire ASCII domain as opposed to the small area immediately surrounding the closed surface low.

4.5.2 Sensible and Latent Heat Flux

Model generated values for sensible and latent heat fluxes (SHF and LHF respectively) are compared for the D-02 and D-03 simulations in Fig. 4.20-4.23 with the purpose of determining the amount of BL modification occurring prior to storm development. In order to ensure that the simulations are being compared at similar times along the life cycle of the surface low, a reference time, T-00, is defined as the time at which the simulated surface low forms or moves into the ASCII domain. All other times use this as a base time, so that T-24 is 24 hours prior to T-00, and T-12 is 12 hours prior to T-00, and so on. For D-02, T-00 is 09Z Dec 25. For D-03, T-00 is 09Z Dec 14 for D-03. In addition to SHF and LHF, SLP and 10-m wind vectors are included on heat flux plots.

Fig. 4.20-4.23 show side by side comparisons of SHF for D-02 and D-03 at various times throughout the simulations. SHF is generally less off the North Carolina Coast and over the Gulf Stream in D-02. Light and variable winds at T-24 result in SHF less than 100-
Wm$^{-2}$ over the southern half of the domain, while stronger westerly winds in the northeastern part of the domain result in values in excess of 200-Wm$^{-2}$. At the same time in D-03, a brisk northeasterly wind results in a broad area of SHF in excess of 100-Wm$^{-2}$ with several areas exceeding 200-Wm$^{-2}$. As time progresses, northeasterly wind increases in D-03 resulting in increasing values for SHF, approaching 400-Wm$^{-2}$ in some locations by T-12. Values for SHF have also increased in D-02 at this time, with values in excess of 100-Wm$^{-2}$ over the Gulf Stream. A convergence of winds has also setup due east of Charleston, SC where southerly and easterly winds converge. Little change is noted between T-12 and T-06 with the exception of the area of greatest SHF moving north in D-02 and being eroded on the southern end in D-03 as southerly flow becomes a more dominate component in the winds.

Differences between the two simulations are most extreme at T-00 in terms of SHF magnitude and 10-m wind direction. Over the water, southerly flow dominates the southern two-thirds of the D-02 domain, with SHF values generally less than 150-Wm$^{-2}$. Larger values for SHF can be found in the northeastern part of the domain where flow becomes more easterly. A southeast to easterly flow dominates much of the ocean waters in the D-03 domain with values of SHF over the GS in excess of 200-Wm$^{-2}$ and approaching 400-Wm$^{-2}$. In D-02 the surface low is over northeastern North Carolina over an area of negative SHF. The surface low in D-03 lies off the coast of Wilmington, NC over a strong SHF gradient.

In addition to the values for SHF, the magnitude of the SHF gradient is examined. To quantitatively examine the magnitude of the SHF gradient off of the North Carolina Coast, the gradient was computed between the first offshore grid point east of ILM and HSE to the GSF adjacent to each location. These values were averaged at various times throughout the
simulation from T-36 through T-00. Generally, the strongest SHF gradient was observed in D-03. The average SHF gradients for T-36 through T-00 are 1.1 Wm$^{-2}$km$^{-1}$ and 2.6 Wm$^{-2}$km$^{-1}$, which correspond to D-02 and D-03 respectively.

Considering that each storm is being influenced by a much broader area than captured by the above calculations, the average SHF and horizontal SHF gradient within the entire ASCII domain was computed. The results, however, still show the greatest values present for D-03. The average SHF for the entire ASCII domain for T-36 through T-00 is 65-Wm$^{-2}$ and 118-Wm$^{-2}$ for D-02 and D-03 respectively. Horizontal heat flux gradients are approximated using a partial derivative function in GEMPAK defined by:

$$ DDX(H_0) = m_x \left( \frac{\partial H_0}{\partial x} \right), $$

(4.1)

where $H_0$ is sensible heat flux, $m_x$ is the horizontal map scale factor, and $x$ is the distance between gridpoints. Average horizontal heat flux gradients in the ASCII domain for D-02 and D-03 are $1.8 \times 10^{-4}$ Wm$^{-2}$km$^{-1}$ and $2.1 \times 10^{-4}$ Wm$^{-2}$km$^{-1}$ respectively. Despite D-03 having higher heat flux gradients through the ASCII domain, D-02 actually has higher gradients at T-36 and T-24 suggesting the biggest modifications to the pre-storm environment for D-02 occurred at and prior to T-24. This is also significant because Kuo et al., 1991, showed that heat fluxes within 24-hours of rapid deepening played virtually no role in the magnitude of the deepening, whereas fluxes in the 48-24 hour period prior to deepening were significant.

Fig. 4.24-4.27 show the LHF for T-36 through T-00 for D-02 and D-03. Because the results are similar to those observed with the SHF analysis, the LHF analyses will not be reviewed in great detail. To summarize, it is noted that the magnitude of LHF is generally less in D-02 and increases as times approaches T-00. The initial location of the surface low
at T-00 is in an area of negative to near zero LHF for D-02 and over a LHF gradient for D-
03. A quantitative analysis of LHF gradients reveals an even larger disparity with average
gradients as was seen in the SHF gradients. The average LHF gradient from T-36 through T-
00 is 3.0 Wm^{-2}km^{-1} and 7.1 Wm^{-2}km^{-1} for D-02 and D-03 respectively. Average LHF within
the ASCII domain for the same time period is 197-Wm^{-2} for D-02 and 289-Wm^{-2} for D-03.
The average horizontal LHF gradients for D-02 and D-03 are 4.5\times10^{-4} Wm^{-2}km^{-1} and 4.7\times10^{-4} Wm^{-2}km^{-1} respectively. Again, the average LHF gradients for T-36 and T-24 are actually
greater for D-02.

The results for the SHF and LHF analyses do not come completely as a surprise given
that surface temperatures prior to the D-03 case were much colder than those prior to the D-
02 case. This resulted in a very strong temperature gradient between land observation
locations and the GSF, and the very high value of PSBI computed for the D-03 case. Only a
strong difference in wind speeds would’ve resulted in higher SHF and LHF values for D-02.
Heat flux gradients prior to T-24 may possibly be a better gauge of the pre-storm
environment, although this needs to be tested with greater scrutiny. However, despite this,
surface conditions have been represented by ASCII fairly well in both D-02 and D-03, and
are likely the only cause for ASCII’s poor performance in the D-03 case.

4.5.3 Surface Convergence

Convergence of 10-m wind is briefly analyzed here as it is directly related to vertical
motion. Large values of surface convergence correspond to rising air and the formation of
clouds and precipitation. It also enhances vorticity aloft through the stretching term in the
quasi-geostrophic vorticity equation. But at the same time, excessively large values of surface convergence can impede a storm’s development if divergence aloft is not large enough to counteract it, resulting in a build up of air in the vertical column above the surface low and preventing deepening of the storm.

Plots of surface convergence (computed from 10-m winds) are shown for T-12 and T-00 in Fig. 4.28 and 4.30. At T-12 in D-02, a line of convergence oriented west to east has developed from the South Carolina Coast eastward into the Atlantic Ocean. This feature is associated with a warm frontal boundary, with southerly winds south of the boundary. The warm front is confirmed in Fig. 4.29 which shows SLP and 2-m temperatures. The warm front sits in an inverted pressure trough on the south side of a sharp temperature gradient, co-located with the line of convergence. By T-00, several areas of convergence are present, the main one extending from coastal South Carolina North through eastern North Carolina. This line is associated with the coastal front, which separates the warm marine air from cold continental air. It is along this frontal boundary which the surface low moves. A similar feature is observed in D-03. At T-12, a line of convergence is present over the GS, and eventually propagates to the coast by T-00. Again, this is the focus for the surface low which parallels the coastline. At T-00, surface convergence is strongest in D-03, and should likewise be associated with the strongest vertical motion, which will be shown in the next section.

4.5.4 Cross Sections

To examine the influence of SHF and LHF on the low-level environment (specifically
static stability and low-level moisture) prior to each case as well as vertical motion, cross sections are plotted from the surface to 700-mb for each case. The transect chosen for the cross sections extends from 34.7N and 78.8W to 32.3N and 75.9W, approximately perpendicular to the GSF and passing over the warm core eddy present in D-02 (Fig. 4.31). Cross section plots include contours of potential temperature and mixing ratio, shaded contours of omega (vertical motion), and arrows showing the strength and direction of circulation tangential to the cross-section.

At T-36 (Fig. 4.32), mixing ratios near the surface for each case range from 4-5 g-kg\(^{-1}\) in the west, to 7 g-kg\(^{-1}\) in the east, over the GS. Static stability (characterized by a high vertical gradient of potential temperature) is also lowest over the GS. Significant vertical motion is confined to near the GS in D-03. 12-hours later, at T-24 (Fig. 4.33), the atmosphere in both cases has changed significantly. Near-surface mixing ratios still range from 4-5 g-kg\(^{-1}\) in the west for D-02, but have increased to 9 g-kg\(^{-1}\) in the east. Values remain approximately the same for D-03 as at T-36. In both cases, the higher mixing ratios near the surface have penetrated deeper into the atmosphere. Near surface static stability has decreased, with the strongest stability above the surface layer associated with D-03, as evidence by the tightly packed isotherms above the surface layer. Some generally unorganized areas of vertical motion have developed. Fig. 4.34 shows cross-sections for T-12. Near-surface mixing ratios have now increased to 5-7 g-kg\(^{-1}\) in the western portion of D-02, and to 11 g-kg\(^{-1}\) in the eastern portions. Mixing ratios of 7 g-kg\(^{-1}\) now extend to near 750-mb. Still little change is observed in the near-surface mixing ratios for D-03; although an area of 5 g-kg\(^{-1}\) mixing ratios is present between 950-mb and 750-mb, but still lower than
observed in D-02. Static stability is now similar for both D-02 and D-03, with strong static stability over land and lower static stability over the GS. Vertical motion has increased significantly in D-02. An onshore component to the flow is now present near the surface for both cases, although the magnitude of the flow is much greater for D-02. Finally, at T-00 (Fig. 4.35), mixing ratios for both cases have become more comparable. Static stability is slightly lower in D-02, especially near the surface over land and the GS. Vertical motion has increased for D-03. The vertical motion near the location of the surface low (denoted by the hash marks) is much greater for D-03, which is consistent with the stronger surface convergence in the previous section.

Overall, a favorable environment for development (with increased moisture, decreased static stability, and vertical motion) was quicker to develop in D-02 than D-03. Stronger onshore flow was also present in D-02 helping to advect the warm moist air from near the GSF into the storm track. Conditions for D-03 did not become comparable in terms of favorability to D-02 until T-00, which may have been too late to have a significant influence on the surface low. The stronger vertical motion prior to T-00 in D-02 also allowed for vertical advection of heat and moisture, further helping to destabilize the atmosphere.

4.5.5 Upper-Level Forcing

An analysis of the upper-level patterns of the two simulations reveals noticeable differences (Fig. 4.13 and 4.17). A deeper 500-mb trough characterizes D-02 with the 540-dm extending down to around Georgia by 51-hours and later deepening to 534-dm. The
trough also begins to take on a negative tilt by 51-hours. On the other hand, the 540-dm contour never makes it farther south than Ohio in D-03, with the trough in a neutral configuration through the simulation. Concerning the position of the 500-mb trough in each case with respect to the location of the surface low, it is observed the trough lags farther behind the surface low in D-03 than in D-02. At 51-hours, the surface low for both D-02 and D-03 is centered near southeastern Virginia. The 500-mb trough axis extends eastern Kentucky and Tennessee in D-02, while the axis lies back across central portions of these states for D-03. At 57-hours the lag is even more evident. Once again, the position of the surface lows is similar, near the Delmarva Peninsula. At that time the trough axis extends from western Pennsylvania down into southeastern Virginia in D-02, while the axis is only now in regions of eastern Kentucky and Tennessee in D-03.

500-mb absolute vorticity is also shown in Fig. 4.13 and 4.17. One of the most noticeable differences in the vorticity pattern is the orientation of the area of positive absolute vorticity associated with the 500-mb trough. Positive vorticity is generally aligned parallel to the height contours along the base of the trough in D-02. In D-03, the vorticity is aligned with the trough axis, perpendicular to the height contours at the base. As a result, the positive vorticity maximum at the base of the trough in D-02 is in better position to interact with the surface low as the trough becomes negatively tilted. On the other hand, in D-03, positive vorticity is not as concentrated and located farther from the surface low.

Because of the fine resolution of the model, it is not always easy to identify a single vorticity maximum associated with the surface low. In D-02, by 51-hours, we can identify the maximum stretching from Georgia into West Virginia as likely having the most impact.
on the surface low as the southern end of the max lies directly upstream of the surface low. At 51-hours into the D-03 simulation, there is no single maximum that can be identified as having the most impact on the surface low. The first significant maximum upstream of the surface low (located in northeastern North Carolina) lies near the southern Georgia/Alabama border. Not only is this maximum weaker than the one associated with the D-02 surface low, it is significantly farther away and most likely having a lesser impact on the surface low. As the 500-mb trough becomes negatively tilted, the height contours cross the vorticity gradient at a greater angle. While the contour lines do cross the vorticity gradient at in angle in the D-03 simulation, the vorticity gradient is weaker and farther away from the surface low. In addition, the spacing between height contours suggests stronger wind speeds across the gradient in D-02 as compared to D-03. Overall, positive vorticity advection (PVA) associated with the D-02 surface low is much stronger, which is associated with greater upper-level divergence (Sutcliffe 1939; Bjerknes and Holmboe 1944).

4.6 Conclusions

I-ASCII performed well for D-02 but poorly for D-03. The surface low for D-03 was expected to deepen at a much greater rate than observed. Sea-level pressure, upper-level forcing, heat fluxes, surface convergence, and cross-sections were compared for D-02 and D-03 to help determine significant differences between the two cases which might determine sources for ASCII error. Initially it was hypothesized that lighter winds, and therefore lower values for SHF and LHF may have resulted in more stability in the lower levels of the D-03 case, ultimately suppressing any rapid development. However, simulations show this is not
the case. SHF and LHF values were generally higher in D-03 as compared to D-02, as were the SHF and LHF gradients between the coast and the GSF. It was also thought that the storm farther removed from the heat flux gradient would have the lowest deepening rate. This was also not the case, as D-02 began its deepening period in northeast North Carolina over an area of negative heat fluxes, far removed from the greatest gradients. D-03 actually began deepening over an area of relatively high SHF and LHF gradients. There was also little evidence that the presence of the warm-core eddy in D-02 impacted I-ASCII performance, despite having an influence on the overall heat flux pattern.

It was, however, noted that the average SHF and LHF gradients in the entire ASCII domain were higher for D-02 than in D-03 prior to T-24. In fact, cross-sections revealed the greatest change to the atmosphere occurred between T-36 and T-24 for both cases. Kuo et al. (1991) demonstrated that the impact of heat fluxes was most significant in the 24-48 hour period before rapid development. Applying their result to the case studies presented here, one would expect a more significant storm for D-02 than D-03 simply by examining the T-24 cross-section (Fig. 4.32). At that time, deeper moisture and lower static stability was present in D-02, suggesting greater potential for significant development.

It should be noted that I-ASCII does take into account the entire 48-hours prior to storm development. The lowest 24-hour average temperature occurring at Wilmington and Hatteras, NC in the 48-hours preceding the development or appearance of a storm in the ASCII domain is used in the calculation of PSBI. Therefore, I-ASCII is already being computed during the time when the temperature gradient (and as a result, heat flux gradients) between the coast and GS is maximized. It therefore seems unlikely that the measure of
surface baroclinicity computed by I-ASCII is the source of error.

Excluding the measure of surface baroclinicity from the possible cause of I-ASCII error, the proximity of the surface low to open ocean waters and the measure of upper-level forcing become the main culprits. D-02 spent approximately three hours the open ocean, while D-03 spent significantly less time (approximately one hour). The lack of LHF and SHF directly beneath the surface low may serve to slow development of surface lows. Therefore, D-03 was in a less likely position for reaching its full potential. In addition, because the majority of the ASCII domain is covered by water, I-ASCII performance may be less accurate for storms that spend a significant portion of their time over land. Further testing is necessary to prove this and possibly correct for it. Until then, I-ASCII should be used with caution for land-locked storms.

Relative to the upper-level forcing, it was shown that surface convergence associated with the D-03 low was greater, and likewise the vertical motion. However, neither of these factors will result in a greater deepening rate if upper-level divergence is weak. Upper-level plots did indeed show that the PVA associated with D-03 was much weaker than that observed with D-02. It is likely, then, that D-02 deepened as a result of stronger upper-level divergence. While some divergence was likely present in D-03, it was counter-balanced by the strong surface convergence which prevented the surface low from deepening too rapidly.

Jacobs et al. (2005) acknowledged that using the strength of the vorticity maximum associated with the surface low was not the most accurate gauge of upper-level divergence, but chose it over vorticity advection due to its ease of use in an operational setting. Their improved ASCII method, which takes into account upper-level forcing, is an improvement on
the original ASCII. In general, it will outperform the original. But, due to its rough estimation of vertical motion by the magnitude of vorticity alone, there are bound to be times when it fails. For instance, the original ASCII method yields a forecasted deepening rate of -14.7 mb/12hr, an error of only 4.7 mb/12hr as compared to the 18.4 mb/12hr error generated by I-ASCII method. Only through a long-term implementation of both ASCII forecasts in an operational setting can it be determined whether making additional modifications to the improved method would be worthwhile. But the bottom line is that the upper-level forcing has the most impact on the development of the surface lows in the cases discussed here, and must be better represented in I-ASCII.
Table 4.1. A comparison of relevant ASCII parameters for 25 December 2002 and 14 December 2003. Here, $d$ is the average distance to the Gulf Stream Front from Hatteras and Wilmington, NC.

<table>
<thead>
<tr>
<th></th>
<th>25 Dec ‘02</th>
<th>14 Dec ‘03</th>
</tr>
</thead>
<tbody>
<tr>
<td>$d$</td>
<td>122.3 km</td>
<td>81.9 km</td>
</tr>
<tr>
<td>PSBI</td>
<td>1.21 °C/10km</td>
<td>2.04 °C/10km</td>
</tr>
<tr>
<td>Vorticity</td>
<td>$19 \times 10^{-5}$ s$^{-1}$</td>
<td>$19 \times 10^{-5}$ s$^{-1}$</td>
</tr>
<tr>
<td>Analyzed</td>
<td>16 mb/12hr</td>
<td>10 mb/12hr</td>
</tr>
<tr>
<td>ASCII</td>
<td>9.1 mb/12hr</td>
<td>14.7 mb/12hr</td>
</tr>
<tr>
<td>I-ASCII</td>
<td>15.8 mb/12hr</td>
<td>28.4 mb/12hr</td>
</tr>
<tr>
<td>ASCII Error</td>
<td>6.9 mb/12hr</td>
<td>4.7 mb/12hr</td>
</tr>
<tr>
<td>I-ASCII Error</td>
<td>-0.2 mb/12hr</td>
<td>18.4 mb/12hr</td>
</tr>
</tbody>
</table>

Figure 4.1. ASCII data (1991-2002) of $-\frac{dP}{dt}$ vs. PSBI and sorted by strength of 500-mb absolute vorticity. The corresponding linear regression fit for each vorticity bin is displayed. Two points denoting the 25 December 2002 and 14 December 2003 events have been added. 1991-2002 ASCII data courtesy of Neil Jacobs.
Table 4.2. Total snowfall amounts, in centimeters, for selected cities for the 24-26 December 2002. Amounts taken obtained from the NWS and NCDC.

<table>
<thead>
<tr>
<th>Location</th>
<th>State</th>
<th>Amount (cm)</th>
<th>Location</th>
<th>State</th>
<th>Amount (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tobyhanna</td>
<td>PA</td>
<td>53.3</td>
<td>Hackettstown</td>
<td>NJ</td>
<td>25.4</td>
</tr>
<tr>
<td>Portland</td>
<td>ME</td>
<td>50.8</td>
<td>Stroudsburg</td>
<td>PA</td>
<td>20.3</td>
</tr>
<tr>
<td>Laconia</td>
<td>NH</td>
<td>48.3</td>
<td>Allentown</td>
<td>PA</td>
<td>18.3</td>
</tr>
<tr>
<td>Monroe</td>
<td>NY</td>
<td>48.3</td>
<td>Islip</td>
<td>NY</td>
<td>15.2</td>
</tr>
<tr>
<td>Middlefield</td>
<td>MA</td>
<td>45.7</td>
<td>Boston</td>
<td>MA</td>
<td>13.7</td>
</tr>
<tr>
<td>Byfield</td>
<td>MA</td>
<td>38.1</td>
<td>New York</td>
<td>NY</td>
<td>12.7</td>
</tr>
<tr>
<td>Cold Spring</td>
<td>NY</td>
<td>38.1</td>
<td>Waterbury</td>
<td>CT</td>
<td>12.7</td>
</tr>
<tr>
<td>High Point</td>
<td>NJ</td>
<td>30.5</td>
<td>Newark</td>
<td>NJ</td>
<td>8.9</td>
</tr>
<tr>
<td>Danbury</td>
<td>CT</td>
<td>25.9</td>
<td>Philadelphia</td>
<td>PA</td>
<td>2.8</td>
</tr>
</tbody>
</table>
Figure 4.2. Sea-level pressure in millibars (solid) and 2-m temperature in °C (dashed) are shown, plotted from NARR data for December 2002. Surface station plots for various locations are also included. Plots are at 00Z Dec 25 (a), 06Z Dec 25 (b), 12Z Dec 25 (c), and 18Z Dec 25 (d).
Figure 4.3. 500-mb heights (solid), sea-level pressure (dotted), and 500-mb absolute vorticity (shaded) plotted from NARR data are shown for December 2002. Units are decameters for heights, millibars for sea-level pressure, and \( \times 10^5 \) s\(^{-1} \) for vorticity. Plots are at 00Z Dec 25 (a), 06Z Dec 25 (b), 12Z Dec 25 (c), and 18Z Dec 25 (d).
Figure 4.4. Base radar reflectivity (DBZ) for December 2002 at 00Z Dec 25 (a), 06Z Dec 25 (b), 12Z Dec 25 (c), and 18Z Dec 25 (d).
Figure 4.5. NARR 24-hour total precipitation analysis (in mm) for 00Z Dec 25 – 00Z Dec 26, 2002. Heaviest precipitation totals were observed from Florida northeastward into New York.
Figure 4.6. 10-day SST composite (in °C) for December 2002, ending 00Z Dec 24. Composite was generated using CW_SST data reduced to 3-km and overlaid on the 00Z Dec 24 NCEP 0.5° RTG_SST analysis. The arrow marks the location of the warm-core eddy.
Table 4.3. Total snowfall amounts, in centimeters, for selected cities for the 14-15 December 2003. Amounts taken obtained from the NWS and NCDC.

<table>
<thead>
<tr>
<th>Location</th>
<th>State</th>
<th>Amount (cm)</th>
<th>Location</th>
<th>State</th>
<th>Amount (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Plattsburgh</td>
<td>NY</td>
<td>76.2</td>
<td>Norfolk</td>
<td>CT</td>
<td>21.6</td>
</tr>
<tr>
<td>Caribou</td>
<td>ME</td>
<td>69.3</td>
<td>Saylorsburg</td>
<td>PA</td>
<td>20.3</td>
</tr>
<tr>
<td>Stowe</td>
<td>VT</td>
<td>58.4</td>
<td>Pittsburgh</td>
<td>PA</td>
<td>17.8</td>
</tr>
<tr>
<td>Tupper Lake</td>
<td>NY</td>
<td>55.9</td>
<td>Grandby</td>
<td>CT</td>
<td>16.5</td>
</tr>
<tr>
<td>New Haven</td>
<td>VT</td>
<td>45.7</td>
<td>New Windsor</td>
<td>NY</td>
<td>15.2</td>
</tr>
<tr>
<td>Francetown</td>
<td>NH</td>
<td>33.0</td>
<td>Mansfield</td>
<td>MA</td>
<td>15.2</td>
</tr>
<tr>
<td>Russell</td>
<td>MA</td>
<td>27.9</td>
<td>Hacketstown</td>
<td>NJ</td>
<td>11.4</td>
</tr>
<tr>
<td>Tionesta</td>
<td>PA</td>
<td>25.4</td>
<td>Allentown</td>
<td>PA</td>
<td>7.6</td>
</tr>
<tr>
<td>Westmoreland</td>
<td>NH</td>
<td>25.4</td>
<td>Philadelphia</td>
<td>PA</td>
<td>3.0</td>
</tr>
</tbody>
</table>

Table 4.4. Peak wind gusts, in miles per hour, for selected cities for the 14-15 December 2003. Amounts taken obtained from the NWS and NCDC.

<table>
<thead>
<tr>
<th>Location</th>
<th>State</th>
<th>Wind Gust (kph)</th>
</tr>
</thead>
<tbody>
<tr>
<td>West Quoddy</td>
<td>ME</td>
<td>119</td>
</tr>
<tr>
<td>Cape May</td>
<td>NJ</td>
<td>103</td>
</tr>
<tr>
<td>Atlantic City</td>
<td>NJ</td>
<td>93</td>
</tr>
<tr>
<td>Belmar</td>
<td>NJ</td>
<td>92</td>
</tr>
<tr>
<td>Milton</td>
<td>MA</td>
<td>89</td>
</tr>
<tr>
<td>Smithfield</td>
<td>RI</td>
<td>85</td>
</tr>
</tbody>
</table>
Figure 4.7. Sea-level pressure in millibars (solid) and 2-m temperature in °C (dashed) are shown, plotted from NARR data for December 2003. Surface station plots for various locations are also included. Plots are at 06Z Dec 14 (a), 12Z Dec 14 (b), 18Z Dec 14 (c), and 00Z Dec 15 (d).
Figure 4.8. 500-mb heights (solid), sea-level pressure (dotted), and 500-mb absolute vorticity (shaded) plotted from NARR data are shown for December 2003. Units are decameters for heights, millibars for sea-level pressure, and $\times10^5$ s$^{-1}$ for vorticity. Plots are at 06Z Dec 14 (a), 12Z Dec 14 (b), 18Z Dec 14 (c), and 00Z Dec 15 (d).
Figure 4.9. Base radar reflectivity (DBZ) for December 2003 at 06Z Dec 14 (a), 12Z Dec 14 (b), 18Z Dec 14 (c), and 00Z Dec 15 (d).
Figure 4.10. NARR 24-hour total precipitation analysis (in mm) for 06Z Dec 14 – 06Z Dec 15, 2003. Observed totals are shown for five locations to verify the accuracy of the NARR precipitation totals.
Figure 4.11. 10-day SST composite (in °C) for December 2003, ending 00Z Dec 12. Composite was generated using CW_SST data reduced to 3-km and overlaid on the 00Z Dec 12 NCEP 0.5° RTG_SST analysis.
Figure 4.12. WRF simulated sea-level pressure in millibars (solid) and 2-m temperature in °C (dashed) for D-02. Plots are at 00Z Dec 25 (a), 06Z Dec 25 (b), 12Z Dec 25 (c), and 18Z Dec 25 (d).
Figure 4.13. WRF simulated 500-mb heights in decameters (solid) sea-level pressure in millibars (dotted) and 500-mb absolute vorticity $\times 10^5$ s$^{-1}$ (shaded) for D-02. Plots are at 00Z Dec 25 (a), 06Z Dec 25 (b), 12Z Dec 25 (c), and 18Z Dec 25 (d).
Figure 4.14. Plot of temperature (°C) and dewpoint (°C) versus pressure at Morehead City, NC (a) and Charleston, SC (b) from WRF output and observations at 00Z Dec 25, 2002.
Figure 4.15. NARR analyzed (a) and WRF simulated (b) 72-hour total precipitation analysis (in mm) for 00Z Dec 23 – 00Z Dec 26, 2002. Observed totals are shown for five locations to verify the accuracy of the NARR precipitation totals.
Figure 4.16. WRF simulated sea-level pressure in millibars (solid) and 2-m temperature in °C (dashed) for D-03. Plots are at 06Z Dec 14 (a), 12Z Dec 14 (b), 18Z Dec 14 (c), and 00Z Dec 15 (d).
Figure 4.17. WRF simulated 500-mb heights in decameters (solid) sea-level pressure in millibars (dotted) and 500-mb absolute vorticity $\times 10^5$ s$^{-1}$ (shaded) for D-03. Plots are at 06Z Dec 14 (a), 12Z Dec 14 (b), 18Z Dec 14 (c), and 00Z Dec 15 (d).
Figure 4.18. Plot of temperature (°C) and dewpoint (°C) versus pressure at Morehead City, NC (a) and Charleston, SC (b) from WRF output and observations at 00Z Dec 14, 2003.
Figure 4.19. NARR analyzed (a) and WRF simulated (b) 72-hour total precipitation analysis (in mm) for 00Z Dec 12 – 00Z Dec 15, 2003. Observed totals are shown for five locations to verify the accuracy of the NARR precipitation totals.
Figure 4.20. WRF simulated sea-level pressure (mb), sensible heat flux (Wm$^{-2}$), and 10-m wind vectors (ms$^{-1}$) at T-24. Analysis times are (a) 09Z Dec 24, 2002 (b) 09Z Dec 13, 2003.
Figure 4.21. WRF simulated sea-level pressure (mb), sensible heat flux (Wm$^{-2}$), and 10-m wind vectors (ms$^{-1}$) at T-12. Analysis times are (a) 21Z Dec 24, 2002 (b) 21Z Dec 13, 2003.
Figure 4.22. WRF simulated sea-level pressure (mb), sensible heat flux (Wm$^{-2}$), and 10-m wind vectors (ms$^{-1}$) at T-06. Analysis times are (a) 03Z Dec 25, 2002 (b) 03Z Dec 14, 2003.
Figure 4.23. WRF simulated sea-level pressure (mb), sensible heat flux (Wm\(^{-2}\)), and 10-m wind vectors (ms\(^{-1}\)) at T-00. Analysis times are (a) 09Z Dec 25, 2002 (b) 09Z Dec 14, 2003.
Figure 4.24. WRF simulated sea-level pressure (mb), latent heat flux (Wm\(^{-2}\)), and 10-m wind vectors (ms\(^{-1}\)) at T-24. Analysis times are (a) 09Z Dec 24, 2002 (b) 09Z Dec 13, 2003.
Figure 4.25. WRF simulated sea-level pressure (mb), latent heat flux (Wm$^{-2}$), and 10-m wind vectors (ms$^{-1}$) at T-12. Analysis times are (a) 21Z Dec 24, 2002 (b) 21Z Dec 13, 2003.
Figure 4.26. WRF simulated sea-level pressure (mb), latent heat flux (Wm$^{-2}$), and 10-m wind vectors (ms$^{-1}$) at T-06. Analysis times are (a) 03Z Dec 25, 2002 (b) 03Z Dec 14, 2003.
Figure 4.27. WRF simulated sea-level pressure (mb), latent heat flux (Wm$^{-2}$), and 10-m wind vectors (ms$^{-1}$) at T-00. Analysis times are (a) 09Z Dec 25, 2002 (b) 09Z Dec 14, 2003.
Figure 4.28. WRF simulated sea-level pressure (mb), surface convergence, and 10-m wind vectors (ms\(^{-1}\)) at T-12. Analysis times are (a) 21Z Dec 24, 2002 (b) 21Z Dec 13, 2003.
Figure 4.29. WRF simulated SLP in mb (solid) and 2-m temperatures in °C (dashed) at 21Z Dec 24, 2002 (T-12).
Figure 4.30. WRF simulated sea-level pressure (mb), surface convergence, and 10-m wind vectors (ms\(^{-1}\)) at T-00. Analysis times are (a) 09Z Dec 25, 2002 (b) 09Z Dec 14, 2003.
Figure 4.31. CoastWatch SST composites for D-02 (a) and D-03 (b) showing the location of the vertical cross sections (solid white line).
Figure 4.32. WRF model cross sections for at T-36. Analysis times are (a) 21Z Dec 23, 2002 (b) 21Z Dec 12, 2003. Cross sections show potential temperature in Kelvin (red), mixing ratio in gkg\(^{-1}\) (blue), omega (ms\(^{-1}\)), and arrows showing the circulation tangential to the cross-section.
Figure 4.33. WRF model cross sections for at T-24. Analysis times are (a) 09Z Dec 24, 2002 (b) 09Z Dec 13, 2003. Cross sections show potential temperature in Kelvin (red), mixing ratio in gkg$^{-1}$ (blue), omega (ms$^{-1}$), and arrows showing the circulation tangential to the cross-section.
Figure 4.34. WRF model cross sections for at T-12. Analysis times are (a) 21Z Dec 24, 2002 (b) 21Z Dec 13, 2003. Cross sections show potential temperature in Kelvin (red), mixing ratio in gkg\(^{-1}\) (blue), omega (ms\(^{-1}\)), and arrows showing the circulation tangential to the cross-section.
Figure 4.35. WRF model cross sections for at T-00. Analysis times are (a) 09Z Dec 25, 2002 (b) 09Z Dec 14, 2003. Cross sections show potential temperature in Kelvin (red), mixing ratio in gkg$^{-1}$ (blue), omega (ms$^{-1}$), and arrows showing the circulation tangential to the cross-section.
CHAPTER 5: A COMPARISON OF MILLER TYPE CLASSIFIED STORMS AND AN EXAMINATION OF WRF SENSITIVITY TO SST RESOLUTION

5.1 Introduction

The February 12, 2006 Northeast blizzard was the only ASCII storm during the 2005-2006 winter season to be classified as a rapidly deepening cyclone. Unlike the December 2002 and 2003 cases examined in Chapter 4, the February 12, 2006 storm (hereafter referred to as F-06) was a Miller Type-A storm which developed over the Gulf of Mexico and moved up the East Coast as a single storm, without secondary redevelopment. The normalized deepening rate for the storm was 12 mb/12hr. ASCII performed well for this particular storm, under-forecasting the deepening rate by just 1.7 mb/12hr. I-ASCII over-predicted the deepening rate by 8.2 mb/12hr, but as discussed in Chapter 3, much of this error can likely be attributed to the relatively high resolution of the GFS dataset used to classify the upper-level forcing with respect to the NCEP-NCAR reanalysis grid on which I-ASCII was derived. Correlating the 500-mb absolute vorticity taken from the NCEP-NCAR reanalysis to the 32-km NARR dataset and updating the I-ASCII forecast yielded an error of 1.2 mb/12hr (see Tables 3.4 and 3.5). So overall, ASCII performed well for this case.

F-06 was simulated with high resolution (3-km) SSTs using WRF with the same model configuration as used for D-02 and D-03 in Chapter 4. Simulated results will be examined and compared to D-02 and D-03 results to identify differences in synoptic and mesoscale features associated with Miller Type-A and Type-B storms. Specifically, the objective is to determine whether there are any features exclusive to Miller Type-A or Type-B storms that would affect I-ASCII performance. In addition to the high resolution SST run, a low resolution (0.5°) SST simulation will be performed and compared to the high resolution
run to determine how sensitive the WRF simulation is to GS features. Here, the goal is to determine whether the added resolution of the high resolution run significantly affects the accuracy of the WRF simulation.

It is expected that Miller Type-A and Type-B storms will not have any significant differences that would impact I-ASCII performance. However, comparing the two types of storms may help identify synoptic and mesoscale features that are common for storms that are forecasted well by I-ASCII. Differences between the high resolution and low resolution SST simulations are expected. Distinct differences in the heat flux patterns should be visible and will have an influence on the vertical structure of the PBL prior to storm development. This should ultimately have an impact in the overall deepening rate of each system and ultimately the accuracy of the simulation.

5.2 Case Overview of 12 February, 2006

On 11 Feb 2006, a surface low moved out of the Gulf of Mexico and onshore between eastern Louisiana and the Florida Panhandle. The system moved over the Southeastern U.S., eventually reemerging over the Atlantic Ocean near the North Carolina/Virginia border on 12 Feb 2006. Snowfall associated with the storm was observed from Arkansas eastward through northern Georgia then northward through Maine. The heaviest snowfall fell from the western mountains of North Carolina through eastern Pennsylvania and coastal New England. According to the NCDC storm events database, snowfall amounts of 30 to 50-cm were common with localized locations exceeding 60-cm, which resulted in numerous traffic accidents and airport delays. Selected snowfall amounts
are displayed in Table 5.1. In addition to snowfall, gusty winds resulted in sporadic damage and near whiteout conditions. Official blizzard conditions were reported in Rhode Island, Massachusetts, and Maine with near-blizzard conditions reported across New Jersey, New York, and Connecticut. Wind gusts along the coast exceeded 80-kph in many locations. A summary of observed wind gusts is shown in Table 5.2. The gusty winds left hundreds of thousands of people without power at some point during the storm. Coastal flooding was also a problem from New Jersey up through Massachusetts. At least four deaths have been attributed to the storm in the Northeast U.S.

Surface analyses depicting the evolution of sea-level pressure, 2-m temperature, and surface station plots every six hours (12Z Feb 11 – 06Z Feb 12, 2006) are shown in Fig. 5.1. Like the December 2002 and 2003 cases analyzed in Chapter 4, the surface low spent much of its time in the ASCII domain over land. But unlike the D-02 and D-03, the February 2006 case was a Miller Type-A event characterized by a single surface low which develops or strengthens over the ocean (Miller 1946). The surface low originated over the Gulf of Mexico early on Feb 11 and moved inland by 12Z (Fig. 5.1a). At that time, a ridge of high pressure dominated the Northeast U.S. as frontal system extended from the surface low northeast along the southeast coast and into the Atlantic Ocean. The surface low moved northeast along the frontal boundary deepening more rapidly as it approached the East Coast. High pressure remained entrenched across northern New England until 00Z Feb 12, 2006 (Fig. 5.1c), then retreated into Canada in response to the approach of the surface low.

An amplifying pattern, signified by an approaching and deepening upper-level trough (Fig. 5.2), helped steer the storm more northward along the East Coast. The associated 500-
mb absolute vorticity maximum approached $32 \times 10^{-5}$ s$^{-1}$ at 18Z Feb 11, 2006 (Fig. 5.2b) then weakened slightly afterwards. Initially (Fig. 5.2a), the vorticity maximum was in poor position to influence the surface low, lying adjacent to the low relative to the 500-mb flow pattern. By 00Z Feb 12 (Fig. 5.2c), the vorticity maximum was positioned upstream from the surface low relative to the upper-level flow allowing for increased absolute vorticity advection over the low. It was at this time that the surface low began to undergo greater deepening.

According to 1-mb contoured plots of SLP at 3-hourly intervals (not shown) from the RUC analysis, the surface low entered the western portion of the ASCII domain at 21Z Feb 11, 2006 at 1006-mb. The storm moved northeast exiting the northern edge of the domain around 09Z Feb 12, 2006 at 994-mb. The normalized deepening rate was 12 mb/12hr. After exiting the ASCII domain, the storm move northeastward paralleling the East Coast and reaching 978-mb near Nova Scotia at 06Z Feb 13, 2006.

Precipitation totals for the 24-hour period from 12Z Feb 11-12Z Feb 12, 2006 are displayed in Fig. 5.3. The heaviest precipitation amounts were located from northeastern Virginia to near New York City where precipitation totals approached 30-mm. Precipitation in this area fell mostly as snowfall. Lighter amounts extended southward towards the Gulf of Mexico where totals of 5 to 10-mm were common. NARR analyzed precipitation for the storm is verified against observations in Fig. 5.7. Fig. 5.4 shows the high resolution CoastWatch SST composite ending 00Z Feb 10, 2006. Two warm core eddies are observed downstream of the Charleston Rise and are denoted by arrows. Water temperatures ranged
from around 9°C in coastal waters to around 23°C in the core of the Gulf Stream, with a pocket of 25°C waters east of HSE.

5.3 WRF Model Simulation of 12 February, 2006

5.3.1. Model Setup

F-06 was simulated using the same model setup as D-02 and D-03 in Chapter 4. Version 2.1.1 of the WRF-ARW core was used with a 12-km outer grid and 4-km inner grid (Fig. 2.3). A total of 31 vertical levels were used with 11 levels below 850-mb. Physics options are listed in Table 2.1 with a detailed description of each selection given in Chapter 2. NARR data were used as initial and boundary conditions. Two separate runs were made for F-06: one with high resolution SSTs (F-06 HR) to be compared with D-02 and D-03; and one with low resolution SSTs (F-06 LR) to be used with F-06 HR as a sensitivity study. F-06 HR used the 3-km resolution CW_SST composite overlaid on 0.5º RTG_SST analysis, while F-06 LR used the 0.5 º RTG_SST analysis over the entire simulation domain. For each case the model was initialized 48-hours prior to the movement of the ASCII low into the ASCII domain as determined from the 3-hourly RUC SLP analyses. This time corresponded to 21Z Feb 09. The duration of the model runs was 72 hours.

5.3.2 Model Verification

First, a verification of the F-06 HR simulation will be given. Results from the F-06 LR simulation will be examined in section 5.4. Fig. 5.5 shows the WRF simulated SLP and 2-m temperatures for the times corresponding to the NARR surface analysis in Fig. 5.1. At
12Z Feb 11, 2006 (Fig. 5.5a) the surface low is analyzed over western North Carolina, as opposed to southeastern Alabama in the NARR analysis. However, the location of the associated frontal boundary matches well with the NARR analysis, extending from the eastern Gulf of Mexico across the Florida Panhandle and paralleling the southeast coastline. The displacement of the surface low is likely an artifact of the influence of high terrain in the WRF model as the center of the surface low comes more in line with that of the NARR analysis once it moves into central North and South Carolina (Fig. 5.5b). From that point on, the positioning and track of the surface low is very well simulated. The magnitude of the simulated low is, on average, several millibars deeper than that of the NARR analysis, likely just an artifact of the increased resolution over the analysis domain. A precursor disturbance not present the NARR analysis is noted in the simulation results at 12Z Feb 11, 2006 (Fig. 5.5a) near HSE which moves northeast over the Atlantic Ocean in the following panels.

Plots of WRF simulated SLP contoured at 1-mb intervals (not shown) reveal the storm moves into the ASCII domain at approximately the same time as the NARR analysis indicates, 21Z Feb 11, 2006. The exit time also matches (09Z Feb 12) with the overall track being comparable. While the simulated low is deeper than the NARR analysis at that time (998-mb versus 1006-mb) the simulated deepening rate matches that of the NARR analysis (12 mb/12hr).

The WRF simulated 500-mb absolute vorticity and heights are shown in Fig. 5.6. There are no glaring discontinuities between the simulated upper-level pattern and the NARR analysis (Fig. 5.2). The phase speed of the upper-level trough is slightly faster in the WRF simulation. At 18Z Feb 11, 2006 during the simulation (Fig. 5.6b) the trough axis extends
from eastern Ontario through western Ohio into northeast Arkansas, while at the same time on the NARR analysis, the trough axis is too far west to be captured in the analysis domain. Also note that the simulated vorticity maximum on the east side of the trough at the same time has progressed slightly further north and east than is depicted in the NARR analysis. The slower movement of these features in the NARR analysis may be a result of a slight amplification in the ridge marked by the noticeable ridge axis which bulges from eastern Pennsylvania into Ontario, a feature not noted in the WRF simulation. However, this appears to have little effect on the rest of the simulation, as the trough axis and location of the vorticity maximum in the WRF analysis (Fig. 5.6c-d) match up well with the final two frames of the NARR analysis (Fig. 5.2c-d).

Analyzed and simulated precipitation totals for 21Z Feb 9 - 21Z Feb 12, 2006 are shown in Fig. 5.7. The WRF simulation does a good job picking up on the heaviest areas of precipitation generating an area in excess of 40-mm near the Gulf Coast with a secondary maximum in the Mid-Atlantic. WRF generates excess totals on the northern edge of the precipitation shield, but far away from the area of interest. Precipitation totals over the ocean are not compared due to the lack of gauges and NARR accuracy.

Overall, WRF does a very good job with the F-06 HR simulation. The track of the low and the simulated deepening rate matches up well with the NARR analysis. The upper-level pattern is sufficiently simulated, especially later in the simulation as the surface low begins to develop more rapidly. The typical disparities in magnitude of SLP and absolute vorticity are seen, most likely a result of resolution differences between the NARR analysis and WRF simulation. The speed of the upper-level trough and vorticity maximum is also
slightly faster in the WRF simulation, but this does not appear to have a major impact in the overall development of the surface low. Precipitation amounts are simulated well over land.

5.4 A Comparison of Miller Type-A and Type-B Events

As was done in Chapter 4, a reference time is defined for each case to ensure each comparison is made at a similar time in the storms’ evolution. T-00 is defined as the time at which the simulated surface low moves into or develops within the ASCII domain. The corresponding times for T-00 are: 09Z Dec 25, 2002 for D-02; 09Z Dec 14, 2003 for D-03; and 21Z Feb 11, 2006 for F-06. Additional times are based on T-00 so that T-12 is 12-hours prior to T-00, or 21Z Dec 25, 2002 for D-02. A summary of the analyzed deepening rates for each storm is shown in Table 5.3. A more detailed analysis follows.

5.4.1 Sea-level Pressure

The simulation of sea-level pressure is first compared to identify differences in the surface pattern between the D-02, D-03 and F-06 HR cases. D-02 and D-03 were both Miller Type-B storms as shown in Fig. 5.8. Note that the initial surface low moving out of the Gulf of Mexico for both cases moves west of the Appalachian Mountains, while a secondary low forms along the Southeast Coast and becomes the ASCII low. F-06 HR, on the other hand, was a Miller Type-A storm, as the low moving out of the Gulf of Mexico moves east of the Appalachian Mountains and does not generate a secondary low. Instead, the same low moves through the ASCII domain and begins a period of rapid deepening as it interacts with the baroclinic zone off of the Carolina Coast. The SLP evolution is similar once the ASCII
storms reach the East Coast, as the lows generally move north-northeastward paralleling the coastline. D-03 and F-06 HR are slightly more comparable in that the surface low forms farther south away from the ASCII domain, whereas the surface low for the D-02 simulation forms across eastern North Carolina, adjacent to the ASCII domain.

In all three cases, the surface lows were confined to the western portion of the ASCII domain with at least a portion of time within the domain spent over land. The analyzed and modeled deepening rates and central pressure are summarized in Table 5.3. F-06 HR spent twice as much time in the ASCII domain according to the analysis, although the simulated low for D-03 stayed within the domain for 12-hours. The analyzed deepening rates are comparable for each storm, although the D-03 storm was the weakest of the three in terms of central SLP. There is greater discrepancy in the simulated deepening rates, with the D-02 simulation deepening an impressive 24 mb/12hr, while the D-03 and F-06 HR cases are one-third and half of that rate respectively.

5.4.2 Upper-level Forcing

To determine whether there are large differences in the upper-level patterns for Miller Type-A and Type-B storms, a comparison of the 500-mb heights and absolute vorticity for the three case studies is shown in Fig. 5.9. At T-24 (first column), note the slightly more amplified flow for the D-02 and D-03 cases. In both cases, a weak upper-level trough is positioned over the Northeast. As time progresses, the upper-level patterns become more uniform as the deep upper-level trough over the Midwest digs into the Eastern U.S. However, the flow field for F-06 HR is a bit slower to react, as flow remains slightly more
zonal at T-12 (middle column) serving to force the surface low east of the mountains as
opposed to D-02 and D-03. By the time the surface low enters the ASCII domain at T-00
(column three) the upper-level trough in D-02 has become negatively tilted. Like D-02, the
500-mb absolute vorticity pattern for F-06 HR it is relatively easy to pick out which vorticity
maximum is having the most influence on the surface low. Although difficult to measure
quantitatively, the diffluence also appears strongest in D-02 and F-06 HR. Note the spacing
of the 500-mb height contours west of the surface low at the final analysis time as compared
to the spacing east of the surface low. The most pronounced difference between the spacing
is in D-02 and F-06 HR. The observed diffluence and organization of the absolute vorticity
in D-02 and F-06 HR suggest greater upper-level forcing associated with the events
translating to deeper surface lows and greater deepening rates.

5.4.3 Sensible and Latent Heat Fluxes

There are significant differences between the SHF and LHF off the North Carolina
Coast associated with each case. The differences between D-02 and D-03 were detailed in
Chapter 3. At T-24 (Fig. 5.10a,d,g), the significant difference in SHF between F-06 HR and
D-02 and D-03 is the southerly flow over most of the analysis domain. This due to the
anticyclonic flow associated with an offshore high pressure system and the frontal system
over the Midwest U.S. (Fig. 5.8g). In both D-02 and D-03 high pressure is situated over the
upper Midwest and Great Lakes region resulting in a more northerly and easterly flow. SHF
values are nevertheless comparable to those observed with D-03 and the northern portion of
the analysis for D-02 at T-24, with a large area exceeding 150-Wm$^{-2}$. By T-12 (Fig.
5.10b,e,h) the precursor surface low is developing near the GS warm core eddy southeast of ILM (Fig. 5.4). This low enhances the southerly flow ahead of it and generates a weak northerly component to the wind just to its north. SHF values now fall more in line with D-02 as a similar circulation pattern is observed. The strong northerly flow is enhancing SHF values in D-03. SHF values over the GS are generally near 150-Wm$^{-2}$ for F-06 HR and D-02 while values approach twice that in spots for D-03.

At T-00 (Fig. 5.10c,f,i) the ASCII low moves into the domain. In F-06 HR and D-02 surface winds are generally south-southwesterly over the GS and turn easterly on the northeast side of the low. Winds are generally more east-southeasterly over the GS in D-03 in part due to the slightly farther south position of the ASCII low. SHF values are largest in D-03 where they once again approach 300 Wm$^{-2}$ in spots. Values are also relatively high in F-06 HR as values approach 250-Wm$^{-2}$. More modest values for SHF are observed with D-02 due to slightly weaker winds. SHF values are generally near or less than 150-Wm$^{-2}$ expect in the northeast corner where values exceed 200-Wm$^{-2}$.

The pattern for LHF follows the same general pattern with the highest values eventually observed in D-03 and F-06 HR (Fig. 5.11). Values for LHF are generally two to three times that as SHF, which is in agreement with earlier studies (Bosart 1981). Otherwise, there are no additional significant results from the LHF analyses.

The inclusion of a third case study provides some additional details on the evolution of East Coast cyclones. There appear to be no fundamental differences in the SHF and LHF patterns associated with Miller Type-A and Type-B storms. The presence of an anticyclone off the East Coast and a frontal boundary to the west of the ASCII domain is noted for F-06
HR resulting in strong southerly flow at T-24. Whether this is a characteristic difference between Miller Type-A and Type-B cyclones or just a general incidental difference between two East Coast storms cannot be concluded from this analysis. What does become apparent is the lack of significance SHF and LHF have on the evolution of the case studies discussed in the 24-hours preceding development. SHF and LHF values were lowest for D-02 in that time period, yet it was the strongest system exhibiting the greatest deepening; while D-03 boasted the highest values for SHF and LHF but was the weakest storm in terms of upper-level forcing and the deepening of the surface low. Surface wind direction may be critical to development during the 24-hours prior to development, as D-03 was appreciably different from D-02 and F-06 HR at T-12 and T-00. Winds at these times were generally northerly and easterly for D-03, while winds were shifting to be more southerly for D-02 and F-06 HR. The presence of a southerly or easterly wind component may help pre-condition the MBL and PBL near the storm track by ushering in warm moist air underneath the colder and drier air aloft, something that will be considered when examining the cross-sections below. The southeasterly component of the wind observed at T-00 during D-03 may have developed too late to contribute significantly to the development the surface low.

5.4.4 Surface Convergence

All three cases begin with a relatively weak 10-m wind convergence-divergence pattern at T-24 (Fig. 5.12a,d,g). By T-12 (Fig. 5.12b,e,h) noticeable convergence has developed in D-02 and F-06 HR. The east-west line of convergence in D-02 is associated with an inverted surface trough ahead of the main ASCII low, while in F-06 HR the
convergence along the Gulf Stream is due in large part to the precursor surface low which
develops near a GS warm core eddy and possibly enhanced by general convergence
associated with the GS. A weaker line of convergence develops in D-03 at T-12. By T-00
(Fig. 5.12c,f,i) the convergence pattern in D-02 and F-06 HR becomes noisy making it
difficult to discern noteworthy areas of convergence. Strong convergence does eventually
develop along the coast for D-03 as the 500-mb absolute vorticity advection increases with
the relatively late approach of the upper-level trough and 500-mb vorticity maximum. The
convergence patterns suggest the strongest vertical motions associated with the development
of the surface low will be observed at T-12 for D-02 and F-06 HR and at T-00 for D-03.
This is verified in the next section.

5.4.5 Cross-sections

Cross-sections were taken at T-36, T-24, T-12, and T-00 along a transect from 34.7ºN
and 78.8º to 32.3ºN and 75.9ºW (Fig. 3.33 and Fig. 5.4). The cross-sections are roughly
parallel to the coast and are plotted from the surface to 700-mb with arrows for total wind
plotted every 10-mb. Other parameters included are potential temperature, mixing ratio, and
omega.

The MBL for F-06 HR at T-36 (Fig. 5.13) is significantly drier than that of D-02 or
D-03 with maximum surface mixing ratios only around 2 g-kg⁻¹ in the west and 4 g-kg⁻¹ in
the east, about half of what is observed in D-02 and D-03. The vertical gradient of potential
temperature is greatest over the GS in all three cases, suggesting low static stability, with the
lowest static stability associated with F-06 HR. Vertical motion is generally weak. By T-24
(Fig. 5.14) mixing ratios begin to exceed 5 g-kg⁻¹ for F-06 HR over the GS region, still lower than the values observed in D-02 and D-03. Higher mixing ratios have also begun to extend deeper into the atmosphere in response to increasing vertical motion. Static stability in the BL has decreased, especially in D-02 and D-03 and in the western portion of the F-06 HR cross-section.

The atmosphere is markedly different for the three cases by T-12 (Fig. 5.15). Mixing ratios for F-06 HR have increased to 7-9 g-kg⁻¹, which is now more comparable to D-03, although still slightly lower than D-02. In all cases, static stability remains strongest over the western portions and lowest over the GS. Onshore flow at the surface is strongest with the D-02 and D-03 with very little observed for F-06 HR. Strong vertical motion has developed in F-06 HR in association with the developing precursor surface low with some subsidence noted over the west where the ASCII low will pass around T-00. General large scale ascent is also noted for D-02. This is in agreement with the surface convergence pattern at the same time, which suggested the strongest vertical motion would be noted at this time for D-02 and F-06 HR.

Fig. 5.16 show the final cross-section for T-00, as the surface low begins to deepen in each case, surface mixing ratios range from around 9 g-kg⁻¹ for F-06 HR to 12 g-kg⁻¹ for D-02. Strong static stability is noted in the bottom left of the D-02 and D-03 cross-sections, just west of the surface lows in these cases. This is not noted in F-06 HR because the surface low is near the western edge of the cross-section or just outside of the analysis domain. Vertical motion has reached a maximum for D-03, in accordance to what was suggested by the surface convergence.
The cross-sections do not suggest there is any significant differences in the evolution of the MBL for Miller Type-A and Type B events. It can be concluded that lack of low-level moisture is not to blame for the stunted development of D-03 with respect to what was expected by I-ASCII, as F-06 HR developed accordingly despite having similar or lower values than observed in D-03. Another interesting observation is that the surface low for each case passed near a horizontal gradient of potential temperature where the BL transitions from high to low static stability. This area denotes the GS induced baroclinic zone where there exists a high gradients of SSTs and heat fluxes. In addition, this area was the location for enhanced vertical motion in each case giving cyclones a favorable boundary to propagate along. In this case of F-06 HR, this instability and ascent may have been responsible in the development of the precursor surface low that developed ahead of the ASCII low.

5.5 SST Sensitivity Test of the February 12, 2006 Blizzard

5.5.1 Model Setup

Most regional operational numerical models operate with grid spacing of no finer than 20-km with coarse SST analyses (generally 0.5-1.0° grid spacing, although the operational WRF uses SST analyses with approximately 0.08° grid spacing). SST analyses with smaller grid spacing resolves mesoscale features in the GS which are smoothed out on coarser analyses. This can result in differences in heat flux patterns and MBL structure which ultimately affect model accuracy. In order to investigate the sensitivity of WRF to the resolution of the input SST analyses, an additional WRF model simulation was performed using a low resolution SST analysis. The model configuration for each run was as described
in section 5.3.1. For F-06 HR, the 3-km CW_SST composite was overlaid on the 0.5º RTG_SST analysis and ingested into the WRF model. For F-06 LR, the 0.5º RTG_SST was ingested directly into the WRF model. Fig. 5.17 shows a close up of the two analyses in and around the ASCII domain. The F-06 HR analysis picks up on GS features, including two warm-core eddies, which are not visible on the F-06 LR analysis. In addition, temperatures in the core of the GS for F-06 HR are about 2ºC warmer than F-06 LR. An analysis of output from each model simulation follows.

5.5.2 Sea-level Pressure

Fig. 5.18 compares F-06 HR (top), F-06 LR (middle), and the NARR analysis (bottom) at T-24, T-12, and T-00. At T-24 (Fig. 5.18a,d,g), the F-06 HR and LR analyses are virtually identical. By T-12 (Fig. 5.18b,e,h), the SLP pattern in areas removed from the GS remain nearly identical. Over the GS itself, there is one slight difference, the development of the precursory low in F-06 HR. Although the inverted trough is visible in both analyses, the precursor low is not analyzed on the F-06 LR simulation. There is some hint of the existence of the precursory low by T-00 (Fig. 5.18c,f,i) for F-06 LR, although it is still not analyzed as a closed low. But, differences associated with the precursor low remain the only considerable difference between the two simulations.

SLP analyses contoured at 1-mb (not shown) are used to compare the deepening rate for each storm. Both storms enter and leave the ASCII domain at approximately the same time (21Z Feb 11, 2006 and 09Z Feb 12, 2006 respectively). The deepening rate turns out to be identical for each run, 12 mb/12hr with no difference in the central pressure between the
two storms. The finer analysis also reveals that the precursor low does develop in F-06 LR, albeit 3-hours later than in F-06 HR. The normalized deepening rate within the ASCII domain for the precursor low does differ slightly for the two simulations. In F-06 HR, the low enters the domain and deepens from 1015-mb to 1004-mb in nine hours yielding a deepening rate of 14.7 mb/12 hrs. The low enters the domain and deepens from 1015-mb to 1007-mb in nine hours during F-06 LR, which yields a deepening rate of 10.7 mb/12 hr, 4 mb/12 hr lower than F-06 HR. However, both lows reach approximately the same minimum central pressure (997-mb for F-06 HR versus 998-mb for F-06 LR) before the low dissipates ahead of the ASCII low around 03Z Feb 12, 2006.

It cannot be determined from the 1-mb NARR SLP analysis whether or not the precursory low was observed. Fig. 5.19 shows NARR SLP contoured at 0.5-mb intervals and 10-m horizontal wind. The parameters are plotted for six analysis times beginning with 06Z Feb 11, 2006. Initially (Fig. 5.19a) a weak 1017-mb surface low is located near the coast of South Carolina. In Fig. 5.19b-5.19c, the low is not analyzed. A prominent inverted surface trough appears by 15Z Feb 11, 2006 (Fig. 5.19d). Although there is not a closed low analyzed at this time, 10-m surface winds do indicate weak cyclonic flow around the trough. A closed low reappears at 18Z Feb 11, 2006 (Fig. 5.19e), but again is analyzed open at 21Z Feb 11, 2006 (Fig. 5.19f). Satellite derived surface winds (from the NASA Quick Scatterometer, QuikSCAT) were plotted on a 0.125º grid for 12Z Feb 11, 2006 (not shown). A clear circulation could not be detected. Infrared satellite imagery for the times corresponding to Fig. 5.19 is displayed in Fig. 5.20. At 09Z Feb 11, 2006 (Fig. 5.20b) an area of clouds in the vicinity of the forming trough (and potential surface low) becomes
clearly visible protruding from a swath of clouds across the southeastern U.S. This protrusion of clouds moves northeast along the inverted trough and eventually becomes indistinguishable by 18Z Feb 11, 2006 (Fig. 5.20e). One last piece of evidence is buoy observations from Station 41025 (Diamond Shoals) and Station 41001 (150 NM East of HSE). Fig. 5.21 shows the locations of the buoys and Table 5.4 shows wind speed and direction from those buoys from 04Z Feb 11, 2006 – 19Z Feb 11, 2006 as obtained from the State Climate Office of North Carolina’s CRONOS database. Note that the wind develops a northerly component between 08Z and 17Z at Station 41025, while winds remain southerly at Station 41001, suggesting a cyclonic flow over the region during this time.

Evidence supporting a precursor low is not overwhelming, but lack of resolution of the NARR data and the reduced observations over the ocean, it is possible that a small surface low was present but not fully captured in the reanalysis. Given that a closed surface low is analyzed in two of the frames in Fig. 5.19, an area of general ascent is present in the same area (as visible in the satellite imagery in Fig. 5.20), and a cyclonic flow is evident from buoy observations, it is likely that a weak precursor low did form. However, given the lack of detail on this low, a full quantitative analysis of WRF performance with respect to the low cannot be given.

5.5.3 Sensible and Latent Heat Fluxes

Some differences in SHF and LHF are observed between F-06 HR and F-06 LR. Plots of SHF for T-24, T-12, and T-00 are displayed in Fig. 5.22. Initially, at T-24, the general SHF pattern matches up well, but there are slight differences in the magnitude of
SHF. Values over the GS in F-06 HR exceed 150-Wm$^{-2}$ in several locations, with a maximum approaching 250-Wm$^{-2}$ observed just east of HSE. Values over the GS are slightly lower in F-06 LR with the maximum values east of HSE just below 225-Wm$^{-2}$, which is to be expected given the slightly cooler SSTs in the GS core for F-06 LR. The same general differences hold for T-12, however, subtle differences can be detected in the wind field primarily in response to the presence of the precursory low (just east of the North Carolina – South Carolina border) which has not yet developed in F-06 LR. These will be discussed in more detail in the next section.

Larger discrepancies in SHF values between the two simulations are observed at T-00. A widespread area of SHF values exceeding 250-Wm$^{-2}$ can be seen over the GS for F-06 HR with a maximum in the south in excess of 300-Wm$^{-2}$. At the same time, values generally remain below 200-Wm$^{-2}$ for F-06 LR. No discernable difference in the wind field is noted at this time. A local maximum of SHF around 175-Wm$^{-2}$ can be seen towards the top right of the analysis domain in F-06 HR. The same area in F-06 LR does not exceed 150-Wm$^{-2}$. There is little difference in SHF away from the GS.

LHF values follow the same general trend, with slightly higher values observed over the GS in F-06 HR (Fig. 5.23). Differences become most pronounced at T-00 where values exceed 1000-Wm$^{-2}$ in F-06 HR but generally remain at 800-Wm$^{-2}$ or less for F-06 LR. Values for LHF are generally in good agreement away from the influence of the GS.

5.5.4 Surface Convergence

Convergence of the 10-m wind for F-06 HR and F-06 LR is displayed in Fig. 5.24.
The convergence fields are nearly identical at T-24 (Fig. 5.24a,d). However, some variations between the two plots can easily be seen at T-12 (Fig. 5.24b,e). As mentioned in the previous section, the precursor low which is already present east of the North Carolina – South Carolina border in F-06 HR has not yet formed in F-06 LR. A line of convergence has developed along the western edge of the GS in both cases. North of HSE, the line of convergence is positioned the same for each case. However, this is not the case south of HSE. For F-06 HR the line runs approximately parallel to the coast until it meets the precursor surface low. From that spot, it juts eastward then southward once more.

Incidentally, Fig. 5.17 reveals the presence of a warm core eddy in this location. Whether the presence of the eddy is directly influencing surface convergence is not entirely clear. But it should be noted that the line of convergence in F-06 HR is generally more ragged than in F-06 LR, which may be a result of the greater detail in the SST analysis. In F-06 LR the line of convergence is smoother, more closely mimicking the edge of the GS in that simulation. So given the location of the GS eddy with respect to the precursor low, the slower development of the low in F-06 LR, and the increased raggedness of the line of convergence in F-06 HR, it is possible the presence of the eddy is enhancing the development of the precursor low.

5.5.5. Cross-sections

To compare the vertical structure of the MBL, a cross section was taken along a transect from 34.7°N and 78.8°W to 32.3°N and 75.9°W (Fig. 5.17), passing over one of the warm core eddies in the high resolution SST analysis. Fig. 5.25 shows the cross sections for F-06 HR and F-06 LR at T-36. Values for surface mixing ratio are comparable between the
two simulations, with only slight differences. As expected, the highest values are observed over the GS with lower values inland. Values over the GS range from 3-4 g·kg⁻¹. Differences in static stability are small, with high static stability observed over land and low static stability over the GS.

At T-24 (Fig. 5.26) differences between the two simulations become more distinct. Mixing ratios begin to exceed 5 g·kg⁻¹ over the GS, with the moist air penetrating slightly deeper into the atmosphere in F-06 HR. Static stability is slightly higher for F-06 LR over the GS. Vertical motion is comparable in both cases, although rising air extends deeper into the atmosphere in F-06 HR, perhaps being responsible for the slightly higher mixing ratios.

The precursor low has developed in F-06 HR by T-12 and is in the process of developing for F-06 LR (Fig. 5.27). The main difference between the cases is the small area of 9 g·kg⁻¹ mixing ratios in F-06 HR near the center of the cross-section. Static stability over the GS remains greater in F-06 LR. The biggest difference in the vertical motion field is the location of the maximum ascent. The pocket of rising air is slightly east in F-06 LR.

Examining Fig. 5.17, this is likely a result of the slightly farther east position of the GS core for F-06 LR.

By T-00 (Fig. 5.28), differences in the mixing ratio field are hard to pick out. The values are generally in good agreement. There are also no huge discrepancies in the static stability field. Once again, the biggest difference is the location of various regions of ascent. But in the west, nearest to the passage of the ASCII low, the vertical motion field is not significantly different.
5.6 Conclusions

Two analyses were presented in this Chapter. The first compared the evolution of two Miller Type-B systems to a Miller Type-A storm. The second was a sensitivity test of a high resolution (3-km) SST analysis compared to a low resolution (0.5º) analysis in WRF. Ultimately, there was no significant difference between D-02, D-03, and F-06 HR that can be attributed strictly to the cyclone type. Because D-02 and D-03 had significant difference on their own, this made it impossible to attribute any differences between those simulations and F-06 HR to the difference in their Miller classification.

However, the comparison of the D-02, D-03, and F-06 HR storms result in one valuable conclusion. While a northerly flow will help enhance SHF and LHF over the GS, it can actually serve to inhibit development (as was the case with D-03). A southerly or southeasterly flow over the future storm track well in advance of ASCII storm helps destabilize the PBL with warm moist air. As a result, large values for SHF and LHF do not necessarily promote rapid development because they are not indicative of surface wind direction which appears to be more important in storm development. A southerly and easterly component to the flow at the surface is necessary well before the storm to help moisten and warm the PBL creating a more unstable environment.

There were some differences between the two sensitivity simulations. Sensible and latent heat fluxes were slightly enhanced over the GS due to the slightly warmer temperatures. In theory, this should have resulted in a more moist and unstable environment for F-06 HR. However, in this case, the differences in heat fluxes were not significant enough to generate drastic differences in the MBL. Cross-sections revealed that differences
in mixing ratio and static stability were minimal. The ragged edge of the GS, due to its meandering, in F-06 HR did result in a more ragged appearance to the line of convergence on the western edge of the GS. In fact, the convergence pattern near on of the GS eddies may have helped promote the development of a precursor surface low which developed three hours earlier in the F-06 HR run than the F-06 LR run. The accuracy of the two simulations in the handling of the precursor low cannot be evaluated due to the lack of observational information about the low. But overall, results suggest that GS features may be important in the simulation of weak synoptic scale lows or meso-lows with weak upper-level support, which supports earlier research (Reddy and Raman 1994; Cione and Raman 1995; Raman and Reddy 1996). Generally, these lows are quick moving or only affect a small area, therefore having little impact on the U.S. mainland. From this analysis, it appears that the inclusion of a high resolution SST analysis in operational numerical models may not have a significant impact on the accuracy of synoptic-scale systems with strong upper-level forcing which do not directly pass over the GS or GSF. Additional sensitivity simulations would be useful in determining whether increased SST resolution significantly improves the accuracy of WRF simulations involving meso-lows and whether the impact of the increased resolution would be greater for storms passing directly over the GS. However, it can be concluded that high resolution SST analyses can produce an environment more favorable for the development of small-scale lows with weak upper-level support, and may impact the accuracy of fine resolution regional model simulations for such events.
Table 5.1. Total snowfall amounts, in centimeters, for selected cities for 11-12 February 2006. Amounts obtained from the NWS and NCDC.

<table>
<thead>
<tr>
<th>Location</th>
<th>State</th>
<th>Amount (cm)</th>
<th>Location</th>
<th>State</th>
<th>Amount (cm)</th>
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<td>MA</td>
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<td>Willow Grove</td>
<td>PA</td>
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<td>MA</td>
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<td>PA</td>
<td>44.5</td>
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<td>Trenton</td>
<td>NJ</td>
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<tr>
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<td>Yardley</td>
<td>PA</td>
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<td>PA</td>
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Table 5.2. Peak wind gusts, in kilometers per hour, for selected cities for 11-12 February 2006. Amounts obtained from the NWS and NCDC.

<table>
<thead>
<tr>
<th>Location</th>
<th>State</th>
<th>Wind Gust (kph)</th>
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</thead>
<tbody>
<tr>
<td>Ocean City</td>
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<td>113</td>
</tr>
<tr>
<td>Barnegat Light</td>
<td>NJ</td>
<td>109</td>
</tr>
<tr>
<td>Breakwater Harbor</td>
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<td>90</td>
</tr>
<tr>
<td>Point Pleasant</td>
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<td>Atlantic City</td>
<td>NJ</td>
<td>80</td>
</tr>
<tr>
<td>Wilmington</td>
<td>DE</td>
<td>71</td>
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Figure 5.1. Sea-level pressure in millibars (solid) and 2-m temperature in °C (dashed) are shown, plotted from NARR data for February 2006. Surface station plots for various locations are also included. Plots are at 12Z Feb 11 (a), 18Z Feb 11 (b), 00Z Feb 12 (c), and 06Z Feb 12 (d).
Figure 5.2. 500-mb heights (solid), sea-level pressure (dotted), and 500-mb absolute vorticity (shaded) plotted from NARR data are shown for February 2006. Units are decameters for heights, millibars for sea-level pressure, and $\times 10^5$ s$^{-1}$ for vorticity. Plots are at 12Z Feb 11 (a), 18Z Feb 11 (b), 00Z Feb 12 (c), and 06Z Feb 12 (d).
Figure 5.3. NARR 24-hour total precipitation analysis (in mm) for 12Z Feb 11 – 12Z Feb 12, 2006.
Figure 5.4. 10-day SST composite (in °C) for February 2006, ending 00Z Feb 10. Composite was generated using CoastWatch SST data reduced to 3-km and overlaid on the 00Z Feb 10 NCEP 0.5° Global SST analysis. The white line denotes the transect for the cross-section used for 5.4.5.
Figure 5.5. WRF simulated sea-level pressure in millibars (solid) and 2-m temperature in °C (dashed) for F-06. Plots are at 12Z Feb 11 (a), 18Z Feb 11 (b), 00Z Feb 12 (c), and 06Z Feb 12 (d).
Figure 5.6. WRF simulated 500-mb heights in decameters (solid) sea-level pressure in millibars (dotted) and 500-mb absolute vorticity $\times 10^5 \text{ s}^{-1}$ (shaded) for F-06. Plots are at 12Z Feb 11 (a), 18Z Feb 11 (b), 00Z Feb 12 (c), and 06Z Feb 12 (d).
Figure 5.7. Total 72-hour precipitation (in mm) for 21Z Feb 9 – 21Z Feb 11 taken from NARR data (a) and WRF simulated output (b). Observed totals are shown for five locations to verify the accuracy of the precipitation totals.

Table 5.3. A summary of the analyzed and simulated deepening rate and central pressure for the D-02, D-03, and F-06 HR case studies.

<table>
<thead>
<tr>
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<th>D-02</th>
<th>D-03</th>
<th>F-06 HR</th>
</tr>
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<tbody>
<tr>
<td>Approximate time in domain</td>
<td>6 hours</td>
<td>6 hours</td>
<td>12 hours</td>
</tr>
<tr>
<td>Deepening Rate (Analyzed)</td>
<td>16 mb/12hr</td>
<td>11 mb/12hr</td>
<td>12 mb/12hr</td>
</tr>
<tr>
<td>Final Central Pressure (Analyzed)</td>
<td>990 mb</td>
<td>1004 mb</td>
<td>994 mb</td>
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<tr>
<td>Deepening Rate (WRF)</td>
<td>24 mb/12hr</td>
<td>8 mb/12hr</td>
<td>12 mb/12hr</td>
</tr>
<tr>
<td>Final Central Pressure (WRF)</td>
<td>987 mb</td>
<td>999 mb</td>
<td>986 mb</td>
</tr>
</tbody>
</table>
Figure 5.8. WRF simulated SLP (in mb) for D-02, D-03, and F-06. Analysis times are 09Z Dec 24, 2002 (a); 21Z Dec 24, 2002 (b); 09Z Dec 25, 2002 (c); 09Z Dec 13, 2003 (d); 21Z Dec 13, 2003 (e); 09Z Dec 14, 2003 (f); 21Z Feb 10, 2006 (g); 09Z Feb 11, 2006 (h); and 21Z Feb 11, 2006 (i).
Figure 5.9. WRF simulated 500-mb heights in decameters (solid) sea-level pressure in millibars (dotted) and 500-mb absolute vorticity $\times 10^5$ s$^{-1}$ (shaded) for D-02. Analysis times are 09Z Dec 24, 2002 (a); 21Z Dec 24, 2002 (b); 09Z Dec 25, 2002 (c); 09Z Dec 13, 2003 (d); 21Z Dec 13, 2003 (e); 09Z Dec 14, 2003 (f); 21Z Feb 10, 2006 (g); 09Z Feb 11, 2006 (h); and 21Z Feb 11, 2006 (i).
Figure 5.10. WRF simulated sea-level pressure (mb), sensible heat flux (Wm$^{-2}$), and 10-m wind vectors (ms$^{-1}$) at T-24, T-12, and T-00. Analysis times are 09Z Dec 24, 2002 (a); 21Z Dec 24, 2002 (b); 09Z Dec 25, 2002 (c); 09Z Dec 13, 2003 (d); 21Z Dec 13, 2003 (e); 09Z Dec 14, 2003 (f); 21Z Feb 10, 2006 (g); 09Z Feb 11, 2006 (h); and 21Z Feb 11, 2006 (i).
Figure 5.11. WRF simulated sea-level pressure (mb), latent heat flux (Wm\(^{-2}\)), and 10-m wind vectors (ms\(^{-1}\)) at T-24, T-12, and T-00. Analysis times are 09Z Dec 24, 2002 (a); 21Z Dec 24, 2002 (b); 09Z Dec 25, 2002 (c); 09Z Dec 13, 2003 (d); 21Z Dec 13, 2003 (e); 09Z Dec 14, 2003 (f); 21Z Feb 10, 2006 (g); 09Z Feb 11, 2006 (h); and 21Z Feb 11, 2006 (i).
Figure 5.12. WRF simulated sea-level pressure (mb), surface convergence, and 10-m wind vectors (ms\(^{-1}\)) at T-24, T-12, and T-00. Analysis times are 09Z Dec 24, 2002 (a); 21Z Dec 24, 2002 (b); 09Z Dec 25, 2002 (c); 09Z Dec 13, 2003 (d); 21Z Dec 13, 2003 (e); 09Z Dec 14, 2003 (f); 21Z Feb 10, 2006 (g); 09Z Feb 11, 2006 (h); and 21Z Feb 11, 2006 (i).
Figure 5.13. WRF model cross sections for at T-36. Analysis times are (a) 21Z Dec 23, 2002 (b) 21Z Dec 12, 2002 (c) 09Z Feb 10, 2006. Cross sections show potential temperature in Kelvin (red), mixing ratio in gkg\(^{-1}\) (blue), omega (ms\(^{-1}\)), and arrows showing the circulation tangential to the cross-section.
Figure 5.14. WRF model cross sections for at T-24. Analysis times are (a) 09Z Dec 24, 2002 (b) 09Z Dec 13, 2002 (c) 21Z Feb 10, 2006. Cross sections show potential temperature in Kelvin (red), mixing ratio in g kg$^{-1}$ (blue), omega (ms$^{-1}$), and arrows showing the circulation tangential to the cross-section.
Figure 5.15. WRF model cross sections for at T-12 for D-02. Analysis times are (a) 21Z Dec 24, 2002 (b) 21Z Dec 13, 2002 (c) 09Z Feb 11, 2006. Cross sections show potential temperature in Kelvin (red), mixing ratio in gkg$^{-1}$ (blue), omega (ms$^{-1}$), and arrows showing the circulation tangential to the cross-section.
Figure 5.16. WRF model cross sections for at T-00. Analysis times are (a) 09Z Dec 25, 2002 (b) 09Z Dec 14, 2002 (c) 21Z Feb 11, 2006. Cross sections show potential temperature in Kelvin (red), mixing ratio in gkg⁻¹ (blue), omega (ms⁻¹), and arrows showing the circulation tangential to the cross-section.
Figure 5.17. A comparison of the input SST analysis used for F-06 HR (a) and F-06 LR (b). The SST analysis for F-06 HR is the CW_SST composite overlaid on the RTG_SST analysis. The SST analysis for F-06 LR is the RTG_SST analysis with no modification. The white line denotes the location of the transects used for the cross-sections in 5.5.5.
Figure 5.18. WRF simulated SLP (in mb) for F-06 HR (a-c) and F-06 LR (d-f) compared with the NARR analysis (g-i). Analysis times are 21Z Feb 10, 2006 (a,d,g); 09Z Feb 11, 2006 (b,e,h); and 21Z Feb 11, 2006 (c,f,i).
Figure 5.19. NARR SLP (in mb) and 10-m wind vectors (ms$^{-1}$) for (a) 06Z Feb 11, 2006; (b) 09Z Feb 11, 2006; (c) 12Z Feb 11, 2006; (d) 15Z Feb 11, 2006; (e) 18Z Feb 11, 2006; (f) 21Z Feb 11, 2006.
Figure 5.20. Infrared satellite imagery for Feb 11, 2006. Analysis times are (a) 06Z Feb 11; (b) 09Z Feb 11; (c) 12Z Feb 11; (d) 15Z Feb 11; (e) 18Z Feb 11; (f) 21Z Feb 11. The red circle highlights the cloud mass associated with the precursor low.
**Figure 5.21.** Map showing the location of two maritime buoys, station 41025 and station 41001.

**Table 5.4.** Observed wind speed (ms$^{-1}$) and direction for station 41025 and station 41001 from 04Z Feb 11, 2006 - 19Z Feb 11, 2006.

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<thead>
<tr>
<th>Date/Time</th>
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<th>Station 41001 Winds</th>
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<td>Direction</td>
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<tr>
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<td>6</td>
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</tr>
<tr>
<td>06Z Feb 11, 2006</td>
<td>6</td>
<td>290º</td>
</tr>
<tr>
<td>07Z Feb 11, 2006</td>
<td>6</td>
<td>290º</td>
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<tr>
<td>08Z Feb 11, 2006</td>
<td>4</td>
<td>310º</td>
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<tr>
<td>09Z Feb 11, 2006</td>
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</tr>
<tr>
<td>10Z Feb 11, 2006</td>
<td>2</td>
<td>360º</td>
</tr>
<tr>
<td>11Z Feb 11, 2006</td>
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</tr>
<tr>
<td>19Z Feb 11, 2006</td>
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<td>140º</td>
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Figure 5.22. WRF simulated sea-level pressure (mb), sensible heat flux (Wm\(^{-2}\)), and 10m wind vectors (ms\(^{-1}\)) at T-24, T-12, and T-00 for F-06 HR (a-c) and F-06 LR (d-f). Analysis times are 21Z Feb 10, 2006 (a,d); 09Z Feb 11, 2006 (b,e); and 21Z Feb 11, 2006 (c,f).
Figure 5.23. WRF simulated sea-level pressure (mb), latent heat flux (Wm$^{-2}$), and 10m wind vectors (ms$^{-1}$) at T-24, T-12, and T-00 for F-06 HR (a-c) and F-06 LR (d-f). Analysis times are 21Z Feb 10, 2006 (a,d); 09Z Feb 11, 2006 (b,e); and 21Z Feb 11, 2006 (c,f).
Figure 5.24. WRF simulated sea-level pressure (mb), surface convergence, and 10-m wind vectors (ms⁻¹) at T-24, T-12, and T-00 for F-06 HR (a-c) and F-06 LR (d-f). Analysis times are 21Z Feb 10 (a,d); 09Z Feb 11 (b,e); and 21Z Feb 11 (c,f).
Figure 5.25. WRF model cross sections for at T-36 (09Z Feb 10, 2006) for F-06 HR (a) and F-06 LR (b). Cross sections show potential temperature in Kelvin (red), mixing ratio in gkg$^{-1}$ (blue), omega (ms$^{-1}$), and arrows showing the circulation tangential to the cross-section.
Figure 5.26. WRF model cross sections for at T-24 (21Z Feb 10, 2006) for F-06 HR (a) and F-06 LR (b). Cross sections show potential temperature in Kelvin (red), mixing ratio in gkg$^{-1}$ (blue), omega (ms$^{-1}$), and arrows showing the circulation tangential to the cross-section.
Figure 5.27. WRF model cross sections for at T-12 (09Z Feb 11, 2006) for F-06 HR (a) and F-06 LR (b). Cross sections show potential temperature in Kelvin (red), mixing ratio in gkg$^{-1}$ (blue), omega (ms$^{-1}$), and arrows showing the circulation tangential to the cross-section.
Figure 5.28. WRF model cross sections for at T-00 (21Z Feb 11, 2006) for F-06 HR (a) and F-06 LR (b). Cross sections show potential temperature in Kelvin (red), mixing ratio in gkg$^{-1}$ (blue), omega (ms$^{-1}$), and arrows showing the circulation tangential to the cross-section.
CHAPTER 6: CONCLUSIONS AND FUTURE WORK

6.1 Summary and Conclusions

Winter cyclones that develop in the Southeast U.S. and nearby coastal waters can have a significant impact on life, transportation, business, and local economies in the Eastern U.S. It is therefore imperative that these systems are understood to the extent that current technologies allow. Accurate predictions are necessary to allow communities to prepare or to prevent false alarms. Advances in meteorology in the past few decades have allowed extensive improvements in the forecasting of winter storms, but severe busts such as the 25 January 2000 “surprise” snowstorm reveal remaining weakness in numerical models and the need for continuing research.

The Gulf Stream (GS) has long been known to be an important source of heat and moisture for developing cyclones in the western Atlantic Ocean. Water flowing north is deflected eastward by the Charleston Rise which results in the meandering of the GS and the frequent development of warm core eddies downstream. This induces a large degree of variability in the distance between the coastline and the western edge of the GS or the GS Front (GSF). When the distance between the coast and the GS decreases, low-level baroclinicity can increase, creating a more favorable environment for rapidly deepening cyclones. ASCII was developed to relate the low-level baroclinicity induced by GS meanders to the deepening rate of cyclones and was later improved (I-ASCII) by classifying storms based on their upper-level forcing. This simple tool can assist operational meteorologists in quickly determining how favorable the environment is for rapid deepening giving them an idea of the magnitude of deepening that can be expected for storms moving
into the ASCII domain. It is also useful as an additional tool to check against numerical models to verify the solutions shown are consistent with the current environment.

The first purpose of this research was to identify the sources of error associated with ASCII/ I-ASCII and to test the performance of I-ASCII in an operational setting. In addition, model simulations are meant to help further the understanding of the development of rapidly deepening cyclones by examining various boundary layer-processes and upper-level processes. Another purpose of this work was meant to identify any fundamental differences between the two Miller type cyclones that would affect ASCII/I-ASCII performance. Lastly, this study was meant to demonstrate the sensitivity of WRF to the resolution of the input SST analysis.

6.1.1 I-ASCII Verification for Winter 2005-2006

During the period December 2005-April 2006 I-ASCII was run twice daily (05Z and 17Z) at the ILM (Wilmington) NWS office. Unfortunately, a near neutral to slightly positive Southern Oscillation Index (SOI) and a slightly negative North Atlantic Oscillation (NAO) probably caused a generally warm and dry winter across the Eastern U.S. Storm development was suppressed with only nine ASCII storms being recorded and only one rapidly deepening storm. This prevented a statistically significant analysis from being performed; however, even with this small sample a major problem with I-ASCII was identified.

Each of the nine ASCII storms was categorized as strongly forced, resulting in an average I-ASCII error of 12.9 mb/12hr. For comparison, the average ASCII error was 5.6
mb/12hr and the average error from the operational Eta model was 1.4 mb/12hr. The problem with the I-ASCII forecast is that not every storm was strongly forced. The grid used to determine the 500-mb absolute vorticity was the 190-km GFS operational model, different from the 2.5º NCEP-NCAR grid used when I-ASCII was originally developed. Absolute vorticity is highly sensitive to grid resolution, with storms that would typically be classified as weakly forced on the NCEP-NCAR grid being classified as strong on the GFS grid. The lack of strongly forced events likely added to some of the I-ASCII error observed, as the relationship between the pre-storm baroclinic index (PSBI) and deepening rate for weakly forced events is relatively weak, potentially skewing the observed error. However, it is still obvious that some significant improvements are necessary before I-ASCII can be implemented and used on a regular basis by operational forecasters.

Three “quick-fix” improvement methods were tested in an attempt to improve the I-ASCII forecasts. The first method was to simply interpolate the GFS grid to the NCEP-NCAR grid using GEMPAK. However, this method still retains much of the details associated with the high resolution grids. This method yielded an improvement of I-ASCII forecast error to 12.0 mb/12 hrs, or only 7.3%. The second method was to use all ASCII cases from the period 2002-2004 to correlate the absolute vorticity values from the GFS grid to the values from the NCEP-NCAR grid. A total of 38 ASCII cases were used to develop a regression equation. This method improved the I-ASCII forecast error to 7.3 mb/12hr, or an improvement of 43.4%. However, neither method improved I-ASCII beyond the original ASCII forecast, with the operational Eta forecast outperforming both.

The third, and most successful method, was to use the *ijskip* parameter in GEMPAK
to reduce the grid spacing of high resolution analyses (in this case, the 32-km NARR dataset). It was concluded that an \textit{ijskip} value of six (or plotting every seventh grid point) effectively reduces the detail on the NARR analysis to a level comparable to that of the NCEP-NCAR analysis. Employing this method for all nine cases during the 2005-2006 winter resulted in an improvement of 54\%, an average error of 5.9 mb/12hr. While this is close to the amount of error exhibited by the original ASCII method, it is still greater than the error associated with the Eta model. Continuing to increase \textit{ijskip} up to nine eventually results in an average error of 3.8 mb/12hra, which is only 2-mb greater than the error associated with the original ASCII. However, at that point, the NARR and NCEP-NCAR analyses do not match up well visually. Given the small storm sample size, a long-term test of this method is needed in order to assess its ability against current operational models.

Although the \textit{ijskip} method is the best method for improving I-ASCII of the three methods tested, the best results may ultimately be obtained by recalculating the I-ASCII regression equations using a finer grid to classify the upper-level forcing. Unfortunately, if the resolution is increased too much, the 500-mb absolute vorticity analysis becomes noisy and it becomes difficult to associate any one vorticity maximum with the developing surface low. Therefore, it may also be necessary to test other methods of improving the forcing classifications, such as employing a spatial average or using another approach for estimating upper-level forcing such as Q-vectors.

One of the biggest results of the 2005-2006 operational test is that finer grid spacing does is not always translate to an improvement in the product. In this instance, part of the problem is simply the fact that I-ASCII was calibrated on a coarse 2.5\degree grid, and this
calibration cannot be applied when using finer grid spacing. But finer grid spacing also results in more detail and noise in the plots of 500-mb absolute vorticity. Instead of being able to easily pick out one single vorticity maximum associated with the surface low, there may be several. Ultimately, it is the net effect of these multiple vorticity features that impact the development of the surface low, while individually these features may not all that meaningful. Decreasing grid spacing tends to draw attention to mesoscale feature that may be of little importance for the broad development of a synoptic-scale system. This explains why I-ASCII was so successful despite being developed from a coarse grid, because rather than focusing on the fine details of multiple vorticity maxima, the coarse grid smoothes out the features which better represents their net effect.

6.1.2 Sources of ASCII Error

A post-analysis of I-ASCII performance during the period 2002-2004 revealed a significant error (18.4 mb/12hr) associated with the 14 December 2003 event. While errors of this magnitude do not appear to be a common occurrence, identifying the source of such a large error may help reduce even small errors associated with I-ASCII. In order to diagnose the source of error, two WRF simulations were performed with a 12-km outer domain and a 4-km nested domain. One simulation was performed for the 14 December 2003 event (D-03) while the second simulation was performed for the 25 December 2002 event (D-02) in which I-ASCII had virtually no error (-0.2 mb/12hr). NARR data was used for initial and boundary conditions. The SST analysis used for each case was a 10-day 12-km CoastWatch composite overlaid upon 0.5° Global NCEP analysis.
It was initially hypothesized that heat fluxes may have been suppressed and heat flux gradients were weaker in the D-03 event resulting in a more stable boundary-layer and a less favorable environment for rapid deepening. Results from the WRF simulation revealed the contrary, higher values of heat flux and heat flux gradients were associated with the D-03 event. The average value for sensible heat flux (SHF) within the entire ASCII domain in the 36-hours preceding storm development was nearly twice for the D-03 case as compared to the D-02 case (118 Wm$^{-2}$ versus 65 Wm$^{-2}$). However, the largest horizontal heat flux gradients in the 36 to 24-hour period prior to storm development were associated with D-02, suggesting the most significant modifications to the boundary-layer occurred during this period, consistent with the findings of Kuo et al. (1991). An analysis of latent heat flux (LHF) showed similar results. Cross-sections of wind vectors and vertical motion revealed that a more favorable environment for rapid cyclogenesis (greater moisture and vertical motion and lesser static stability) developed earlier in D-02 than D-03 as a result of the higher values of surface heat fluxes. But because I-ASCII already considers the pre-storm surface environment up to 48-hours prior to storm development, it was concluded that the surface forcing discussed here were not responsible for the poor performance of I-ASCII, which is reinforced by the observation that ASCII error was less than I-ASCII error for D-03 (4.7 mb/12hr versus 18.4 mb/12hr).

Significant differences in the upper-level forcing were noted between D-02 and D-03. Weaker values for 500 mb absolute vorticity and vorticity advection resulted in less upper-level divergence for D-03. It was also noted that D-03 spent only about one hour over the ocean, while D-02 spent nearly three hours over the ocean. The additional time spent over
land may have stunted the development of the D-03 surface low. But ultimately, the ingredients needed for D-03 to properly take advantage of its highly baroclinic environment came together too late, with the static stability and upper-level forcing not becoming comparable to that of the D-02 case until 12 to 24-hours prior to the surface low moving into the ASCII domain. So it seems likely that the source of I-ASCII error is two-fold. First, I-ASCII does not take into account the amount of time a storm spends over land. And secondly, the measure of upper-level forcing fails to take into account the proximity and timing of the 500-mb absolute vorticity maximum with respect to the surface low, and therefore does not always provide the most accurate measure of the upper-level divergence and forcing.

6.1.3 Comparison of Miller Type-A and Type-B Cyclones

A third case study was simulated and examined to determine if the Miller cyclone type has an impact on I-ASCII performance. The 12 February 2006 blizzard (F-06) developed in the Gulf of Mexico and moved up the East Coast as a single system (Miller Type-A scenario) as opposed to undergoing secondary redevelopment along the coast (Miller Type-B scenario) like D-02 and D-03. In other words, the storm developed in the Gulf of Mexico and moved up along the East Coast without secondary redevelopment or a transfer of energy. A WRF simulation was performed for F-06 identical to that performed for the D-02 and D-03 events. Because of the significant differences between December 25, 2002 (D-02) and December 14, 2003 (D-03) storms, it was impossible to attribute any differences between the February 12, 2006 (F-06) storm and the D-02 and D-03 storms to the difference in their
Miller classification. However, the addition of the third case (F-06) in which I-ASCII performed reasonably well, additional conclusions were drawn concerning I-ASCII error, which will be discussed in section 6.1.5.

6.1.4 Sensitivity of WRF to SST Resolution

Most operational models operate with a grid spacing no finer than 20-km, with a coarser resolution SST analysis often used. Many GS features are lost when a coarser resolution analysis is used which may ultimately impact model performance and accuracy. To determine how sensitive the WRF model is to differences in the resolution of the input SST analysis, two simulations of F-06 were performed. In the first simulation, a 10-day 3-km resolution SST composite was overlaid upon a 0.5º SST analysis was used (F-06 HR) for the input SSTs. In the second simulation, the 0.5º SST analysis was employed over the entire model domain (F-06 LR).

The most noticeable difference between the F-06 HR and F-06 LR runs was the earlier development of a precursor low ahead of the main ASCII low in F-06 HR. This was likely driven by the presence of slightly higher heat fluxes in the F-06 HR simulation and the surface convergence pattern near the GS warm core eddy which was not resolved by the F-06 LR simulation. While the accuracy of the two simulations in the handling of the low cannot be assessed due to lack of observational information on the low, increased SST resolution does lead to a more favorable environment of such a low with increased convergence associated with GS features. Because the two simulations were very similar in the evolution of the main ASCII low, it is possible that increased SST resolution may have the greatest
impact on weak synoptic scale lows or meso-lows with weak upper-level support. The impact of GS features on stronger storms is likely minimal due to the dominance of upper-level support and other forcing mechanisms, although additional sensitivity simulations are needed to confirm this.

6.1.5 Role of Advection

Overall, I-ASCII is most sensitive to the method of classifying the upper-level support, due to the sensitivity to grid resolution and the subjectivity involved in choosing the associated 500-mb absolute vorticity maximum. The method of obtaining PSBI from the coastal surface temperature and SST at the GSF is adequate. However, an additional result obtained by comparing the D-02, D-03, and F-06 runs was the importance of surface wind direction prior to storm development. The increased pre-storm baroclinicity helps modify the atmosphere directly over the baroclinic zone which only impacts storms moving over that zone. In all three simulations, the surface lows passed west of the baroclinic zone. In these cases, surface wind direction is crucial as it advects warm moist air inland towards the storm track. In D-02 and F-06, an easterly or southeasterly component to surface flow was present long in advance of the surface low, allowing for the sufficient destabilization of the atmosphere along the storm path. This onshore flow was later to develop in D-03 and may have contributed to the lack of development associated with the event.

6.2 Future Work

Results from this study leave a number of possibilities open for future research. The
most significant is potential improvements to I-ASCII. First and foremost, a more extensive operational test of I-ASCII is necessary to fully gauge its performance over a broader spectrum of storm strengths. It would be helpful to employ a method to resolve grid spacing issues prior to the operational use to ensure the I-ASCII method being tested is close to the method derived by Jacobs et al. (2005) as possible. Following a full operational test, it can then be assessed whether it would be useful to re-calibrate I-ASCII based on a less subjective and more representative method of classifying upper-level forcing such as a spatial average of 500-mb absolute vorticity advection or Q-vectors. It may also be helpful to examine the accuracy of I-ASCII in relation to the amount of time a particular storms spends over land. If there are significant differences between storms which are primarily land-locked versus those that remain over the ocean, then a degree of uncertainty will have to be emphasized for storms passing over land. Another possibility would be to add an additional caveat to the ASCII storm classification that requires storms to spend the majority of their time in the ASCII domain over water. Another opportunity for future research lies in the impact of pre-storm surface flow near the storm track. An extensive examination of the importance of onshore flow prior to ASCII events may prove significant and help improve I-ASCII performance in the future. Finally, the sensitivity of WRF to SST resolution may vary significantly depending on the strength of upper-level forcing and the proximity of the surface low to the GS. In addition, the impact of SST resolution on simulation accuracy needs to be further investigated. Further research on these issues may have significant implications to operational numerical models.
REFERENCES


