P2.11 **Numerical Simulations of the Extratropical Transition of Floyd (1999) Along the U.S. East Coast**

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1. INTRODUCTION

The impact of landfalling tropical cyclones can extend well into the mid-latitudes. Tropical cyclones undergoing an extratropical transition (ET) can develop into powerful, midlatitude cyclones that can cause significant damage from wind and waves in coastal areas. Tropical systems undergoing an ET are also often associated with heavy precipitation to the north and west of the cyclone track, which can result in devastating flooding as well as loss of life and property several hundred kilometers from the storm center. Although there is no strict definition of an ET, typically such transitions are associated with the development of storm asymmetries in the precipitation, temperature, and wind fields as the storm moves toward higher latitudes.

This study focuses on the ET of hurricane Floyd, which made landfall along the southern North Carolina coast at 0900 UTC 16 September 1999 with 50 m s^{-1} winds, which is a category 2 on the Saffir-Simpson scale (Simpson and Riehl 1981). The winds associated with Floyd weakened to tropical storm force as it moved quickly northeastward along the East Coast; however, a swath of heavy precipitation developed ahead of the storm from North Carolina to New England. This paper discusses Floyd's evolution along the East Coast and the mechanisms for the heavy rainfall over southern New England, where 20-40 cm fell in 12-18 h across northern New Jersey, southeastern New York, and central Connecticut. As documented in the NOAA publication *Storm Data*, the heavy rainfall across this region resulted in more than one billion dollars in flood damage and 16 fatalities.

The complexity of the ET process is evidenced by the fact that none of the operational models at the National Centers for Environmental Prediction (NCEP) was able to predict the magnitude and location of Floyd's heavy precipitation even 24 hours leading up to the event (Atallah and Bosart 2001). For example, the 24-h forecast from the 32-km Eta model produced less than half of the observed heavy rainfall over the flooded areas of southern New England (not shown), and the cyclone was too slow by 2-3 hours moving up the coast.

There have been a growing number of climatologies and observational case studies of ET events; however, there have been very few high resolution simulations of particular ET cases. The goal of this paper is to use the Pennsylvania State University - National Center for Atmospheric Research (PSU-NCAR) Mesoscale Model 5 (MM5) to determine the capabilities of this modeling system in simulating the synoptic and mesoscale structures associated with the ET of Floyd (1999). This storm has been investigated using both observations and gridded large-scale analyses (Atallah and Bosart 2003). Using a PV approach, Atallah and Bosart (2003) illustrated the development of the deep baroclinic zone to the north of Floyd and its relation to the deep ascent and heavy precipitation. However, their analysis could not address some of the important mesoscale temperature, wind, and precipitation structures or separate some of the physical processes. Using the MM5 down to 1.3-km grid spacing, this paper addresses the following questions:

- How well can the MM5 simulate the precipitation distribution and amounts associated with Floyd's ET?
- What synoptic and mesoscale processes led to the development of an enhanced low-level baroclinic zone near the coast and the upper-level front aloft over the Northeast?
- Did the topography over the eastern U.S. affect Floyd's evolution and enhance the inland rainfall?
- How important are the surface fluxes and diabatic effects from precipitation in maintaining the circulations associated with Floyd's ET?

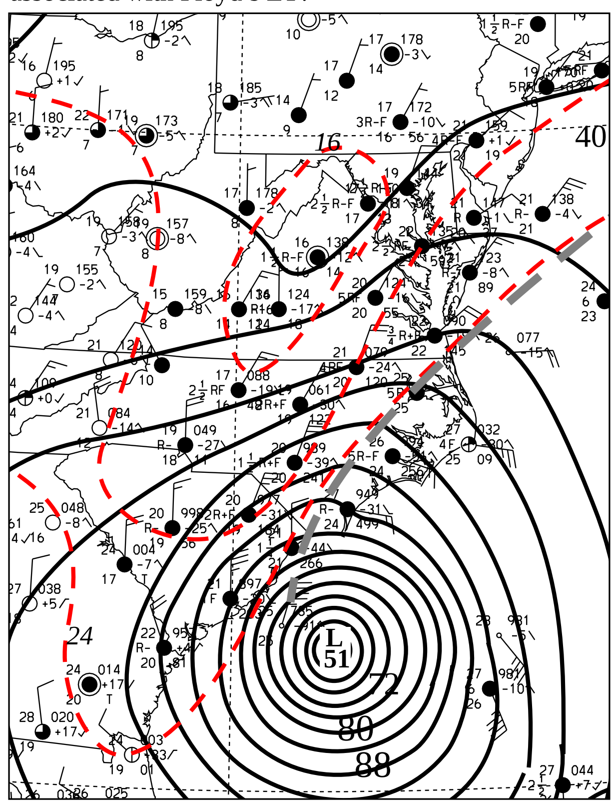

Figure 1. (a) Manual surface analysis for 0000 UTC 16 September 1999. Sea-level pressure and temperature are contoured every 4 mb and 4 ^{o}C, respectively, and the plotted data at each station include temperature (^{o}C), dewpoint (^{o}C), wind (1 full barb = 10 kts and 1 kt = 0.515 m s^{-1}), sea-level pressure, 3-h pressure change, cloud cover, and present weather. The thick dashed line is the surface trough axis.

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2. OBSERVATIONAL ANALYSIS

At 0000 UTC 16 September 1999 (Fig. 1), hurricane Floyd had a central pressure of 951 mb and was located 200 km south of Cape Fear, North Carolina, where it made landfall 9 hours later. As Floyd approached the coast, it began to interact with a broad baroclinic zone that extended northeastward from southeast Georgia to southern New England. Across the baroclinic zone the winds shifted from easterly to northeasterly, while relatively cool air (< 22 ^{o}C) was being advected southward to the west of Floyd. At 500 mb (not shown), the baroclinic zone extended westward from the East Coast to the Great Lakes in association with an approaching mid-latitude trough. There was confluent deformation between the southerly and southwesterly flow over the mid-Atlantic and southern New England, respectively.

As Floyd crossed over the mid-Atlantic region, important mesoscale temperature, wind, and precipitation structures developed near the southern New England coast. For example, at 2100 UTC 16 September 1999 (Fig. 2), tropical storm Floyd was centered near the southern tip of New Jersey. The offshore circulation near the Gulf Stream advected relatively warmer air (> 24 ^{o}C) towards the southern New England coast. Meanwhile, across the warm front from central Connecticut to central New Jersey the surface temperature transitioned from 24 to 16 ^{o}C within 10-20 km, and the winds veered rapidly from southeasterly to northeasterly. In contrast, the temperature gradient was much weaker to the south of Floyd. Overall, Floyd was developing low-level baroclinic characteristics consistent with a marine extratropical cyclone, in which there is well-pronounced warm or bent-back front to the north of the cyclone and a weaker baroclinic zone to the south of the low center.

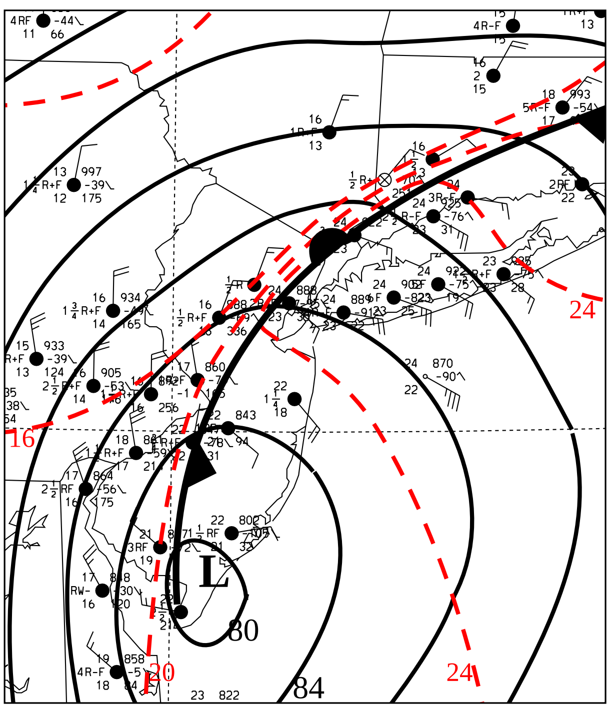

Figure 2. Manual surface analysis for 2100 UTC 16 September 1999 showing sea-level pressure (solid) every 4 mb and temperature (dashed) every 4 ^{o}C.

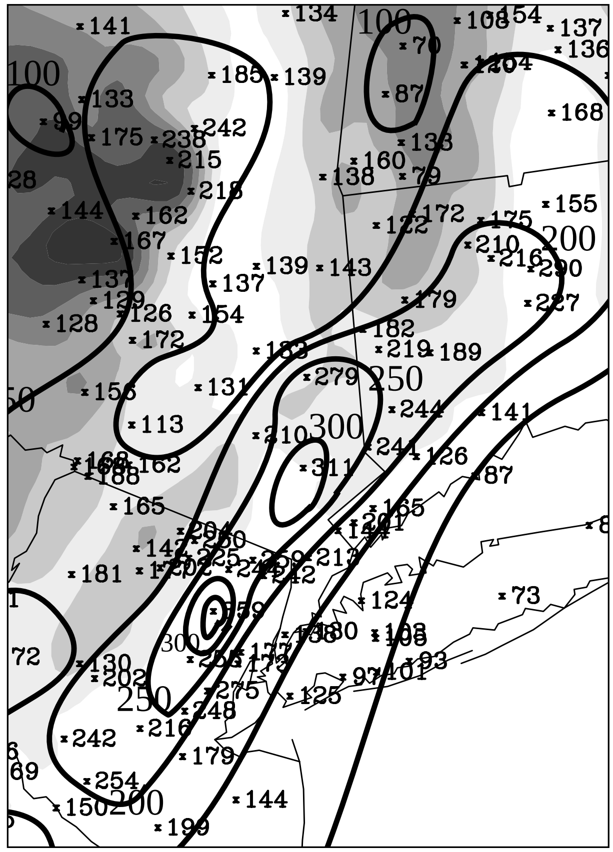

Figure 3. Observed precipitation (solid every 50 mm) from 0600 UTC 16 September to 0600 UTC 17 September 1999. For reference, the terrain from the 4-km MM5 simulation is shaded every 150 m starting at 100 m.

Between 1900 and 2100 UTC 16 September, the heavy precipitation became better organized along the baroclinic zone to the north of Floyd (not shown). Figure 3 shows the storm total precipitation between 0600 UTC 16 September and 0600 UTC 17 September across a portion of the Northeast using nearly 200 National Weather Service (SAO) and National Climatic Data Center (NCDC) cooperative observer (COOP) sites. A swath of heavy precipitation in excess of 20 cm (7.9 inches) fell from northern Connecticut southwestward to extreme eastern Pennsylvania. The heaviest precipitation in excess of 30 cm (11.8 inches) fell around the southeast corner of New York and northeast New Jersey. This region corresponds with the region of enhanced temperature gradient to the north of Floyd at 2100 UTC and the 50+ dBZ reflectivities observed around this time (not shown).

Some of the precipitation across the region was clearly modified by the terrain (Fig. 3). For example, there was windward enhancement (> 17 cm) along the eastern slopes of the Appalachians and Catskills of southeast New York, while rain shadowing (< 10 cm) existed in the lee (west) of the north-south orientated Berkshire Mountains of western Massachusetts. From the observations, it is unclear whether the heavy precipitation swath (> 25 cm) near the coast was enhanced by flow blocking or channeling from the inland hills and mountains. This will be diagnosed using high resolution simulations presented in subsequent sections.

3. MODEL SIMULATION OF THE FLOYD TRANSITION

The MM5 (version v2.12) was used in non-hydrostatic mode to simulate the ET of Floyd in order to provide additional data for diagnosing the structural evolution and associated precipitation structures. For this simulation, stationary 1.33-, 4-, and 12-km domains were nested within a 36 km domain using one-way interfaces. The model top was set at 100 mb. Thirty-three unevenly spaced full-sigma levels were used in the vertical, with the maximum resolution in the boundary layer. Five-minute averaged terrain data were analyzed to the 36- and 12-km model grids using a Cressman analysis scheme and filtered by a two pass smoother/desmoother. For the 4- and 1.33-km domains, a 30-second topography data set was interpolated to the grid in order to better resolve the inland hills and valleys. A 30-second land use dataset from NCAR was used to initialize 13 surface categories for all domains. Initial atmospheric conditions at 0000 UTC 16 September 1999 were generated by interpolating the NCEP Eta model 221 grids (32-km grid spacing) to the MM5 grid. Additional analyses generated in the same manner using the 3-hourly Eta forecasts and were linearly interpolated in time in order to provide the evolving lateral boundary conditions for the 36-km domain. The U.S. Navy Optimum Thermal Interpolation System (OTIS) sea-surface temperature analyses (~30-km grid spacing) were used to initialize the MM5 surface temperatures over water.

The control (CTL) simulation used the explicit moisture scheme of Reisner et al. (1998), which includes prognostic equations for cloud ice and water, snow, rain, and graupel. The Kain-Fritsch convective parameterization (Kain and Fritsch 1990) was applied, except for the 4- and 1.33-km domains, where convective processes could be resolved explicitly. The planetary boundary layer (PBL) was parameterized using NCEP's MRF scheme (Hong and Pan 1996). Klemp and Durran's (1983) upper-radiative boundary condition was applied in order to prevent gravity waves from being reflected off the model top

Floyd was initialized as a 976 mb cyclone at 0000 UTC 16 September (not shown), which is nearly 25 mb weaker than observed at this time (Fig. 1); however, the goal of this study was not to investigate the evolution of the hurricane vortex or eyewall during landfall, but rather to document the larger-scale changes in storm structure and the mesoscale precipitation mechanisms across southern New England. By 2100 UTC (Fig. 4), the position of Floyd was located just east of Delaware, which is about 30 km to the south and within 1-2 mb of the observed (Fig. 2). By this time the coastal temperature gradient had increased across northeast New Jersey and southern Connecticut, where there was significant flow deformation acting on the temperature field. The position of the front in the MM5 was within 20 km of the observed; however, the MM5 temperature gradient was 20-30% weaker than observed since the 24 ^{o}C isotherm did not make it north to the southern New England coast. By 2100 UTC, the MM5 precipitation had increased across northern New Jersey, but the 12-km MM5 did not collapse the precipitation into a narrow band as observed (not shown).

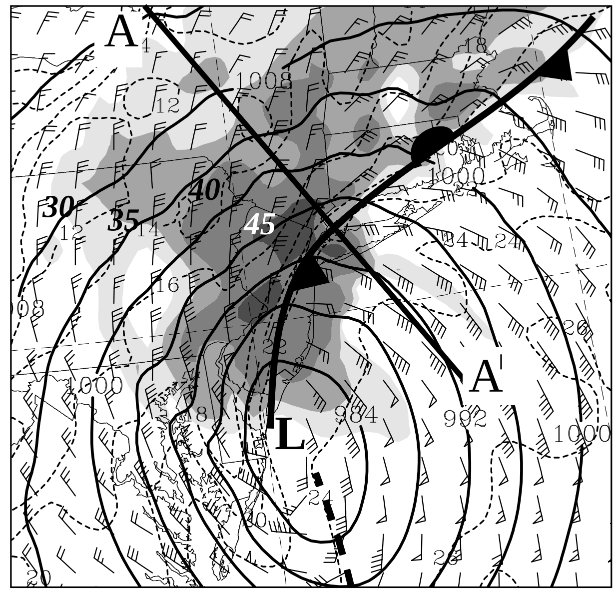

Figure 4. Model surface analysis for the 12-km domain at 2100 (21 h) 16 September 1999 showing sea-level pressure (solid) every 4 mb, surface temperatures (dashed) every 2 ^{o}C, surface wind barbs (full barb=10 kts), and model reflectivities (shaded) every 5 dBZ.

At 36-km grid spacing (not shown), the MM5 realistically simulated the orientation of the heavy rainfall from southwest New Jersey to western Connecticut; however, the simulated precipitation (200-250 mm) is 20-30% less than the observed maximum for northeast New Jersey and southeast New York. Figure 5 shows the MM5 precipitation from 0600 UTC 16 September through 0600 UTC 17 September at 1.33-km grid spacings as well as the observed precipitation at each station. There is large spatial variability in the 1.33 km precipitation since the low-level easterly flow interacts with some of the 100-300 m hills across the region.

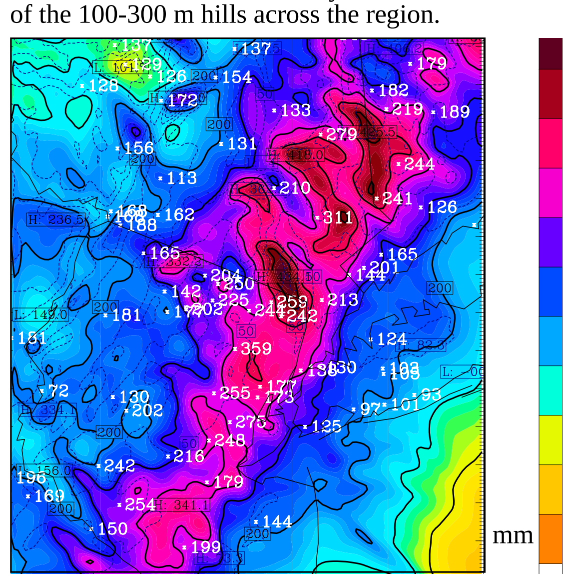

Figure 5. Storm total model precipitation (color shaded and contoured in mm) between 0600 UTC 16 September and 0600 UTC 17 September 1999 for the 1.33 km domain. The observed precipitation totals in mm are also plotted for each station. The 1.33 km terrain is dashed every 100 m starting at 50 m.

For example, there are local minima and maxima in precipitation across the terrain of the southeast corner of New York. The 1.33-km precipitation maximum over northeast New Jersey exceeds 330 mm, which is close to the observed maximum; however, the 1.33-km MM5 overpredicted the precipitation around southeast New York, southwest Connecticut, and the surrounding areas of moderate precipitation (> 250 mm). Overall, from 4- to 1.33-km grid spacing (not shown), the overprediction problem amplifies, and there is little difference at the highest precipitation thresholds for the few available observations. Since the precipitation primarily involved warm rain processes, there was little sensitivity to using a less sophisticated microphysical parameterization, which does not include graupel or supercooled water (not shown).

4. FRONTOGENESIS CALCULATIONS

The observed and model analyses in the previous sections suggest that the heavy precipitation is related to the increasing baroclinicity along the mid-Atlantic and Northeast coasts as Floyd moved northward. To quantify this change in horizontal temperature gradients and the forcing necessary to generate the vertical circulations, the Miller (1948) frontogenesis equation was calculated on pressure levels. Frontogenesis at 36-km grid spacing was also calculated from northwest to southeast across the Northeast U.S (section AA' in Fig. 4) as the low-level temperature gradient increased around 1800 UTC 16 September (Figs. 1,2). Total frontogenesis is maximized within the frontal zone below 850 mb as a result of the strong deformation frontogenesis (Fig. 6a). Deformation frontogenesis greater than 2 K/(100 km h) also extended upwards to 600 mb, which resulted in a thermally-direct circulation and the organized band of heavy precipitation.

Both the radar and model cross sections suggest that convection may have enhanced the precipitation rates along the front. In particular, the mesoscale (30-km wide) and intense nature of the precipitation band orientated parallel to the low-level baroclinic zone is consistent with the release of conditional symmetric instability (CSI) (Schultz and Schumacher 1999). Figure 6b shows the saturated equivalent potential temperature (θ_e*), geostrophic momentum, and shaded regions of negative MPV_g* for cross section AA' at 1800 UTC 16 September from the 36-km domain. To the south of the coastal front, there is conditional instability since $d\theta_e*/dz < 0$, which corresponds to an area of $MPV_g* < 0$. Meanwhile, within the frontal zone, air parcels were nearly neutral to moist gravitational instability below 900 mb, but there was CSI from the surface up to 800 mb as indicated by the MPV_g* < 0. The slope of the θ_e* and geostrophic momentum surfaces are nearly parallel to each other by around 700 mb, thus suggesting moist symmetric neutrality. Even though a deep layer of strong CSI instability did not to mid-levels near within the frontal zone, weak symmetric stability or neutrality can still accelerate the frontogenetical circulation by developing a more concentrated updraft on the equatorward flank of the frontal zone (Emanuel 1985). This enhanced frontogenetical circulation resulted in the mesoscale band of precipitation just inland of the surface front during this event. This frontal circulation becomes more narrow and intense as the grid spacing is decreased in the MM5, thus resulting in increased precipitation rates.

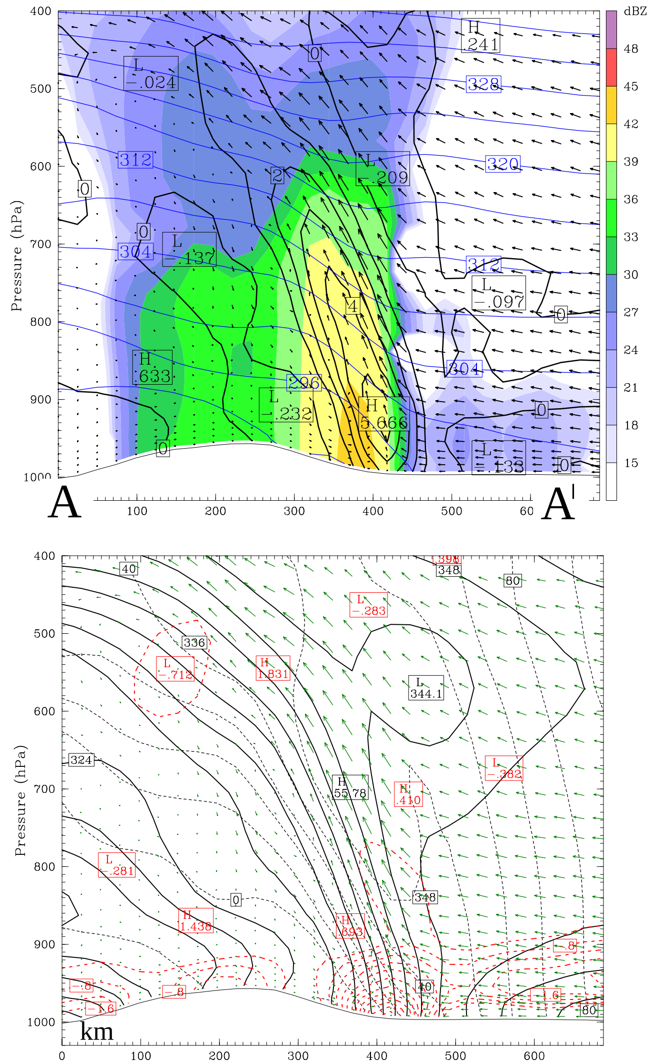

Figure 6. (a) Cross section AA' from the 36-km MM5 showing deformation frontogenesis (solid every 2 ^{o}K (100 km h)$^{-1}$), winds parallel to the section, and potential temperatures for 1800 UTC 16 September 1999. Model derived reflectivities in dBZ are color shaded. (b) Same as (a) except showing geostrophic momentum (dashed every 10 m s^{-1}) and saturated equivalent potential temperature (solid every 4 K). Negative moist geostrophic potential vorticity is dashed every 0.4 PVU,

5. SENSITIVITY RUNS

Interestingly, the axis of heaviest rainfall was orientated along and nearly parallel to some of the 100-300 m hills over southern New England (Fig. 2); therefore, what role did the coastal hills and Appalachians have on the precipitation distribution? Atallah and Bosart (2003) hypothesized that the channelling of cool, northeasterly flow just inland of the coast may have enhanced the frontogenesis at low levels during the Floyd event. In other ET events, such as Agnes (1972), upslope flow over the inland terrain was suggested to be important in the flooding (Bosart and Dean 1991).

In order to remove the influence of terrain during the Floyd event, the Appalachians and coastal hills over the eastern U.S. were replaced by flat land at sea-level at the start of the simulation (not shown), and all MM5 domains were rerun (NOTER experiment). There was little or no terrain impact on the position of Floyd and the intensity of coastal baroclinic zone as compared to the CTL run (not shown). The absence of terrain flow deflection is consistent with the large Froude numbers to the north of the surface front {$Fr = U/(h_m N)$, where U is the ambient flow, h_m is the topography height, and N_m is the moist static stability}, which were around 4 for this event (where U = 20 m s^{-1}, h_m =1000 m, and N_m = 0.005 s^{-1}).Figure 20 shows the 1.33-km precipitation between 0600 UTC 16 September and 0600 UTC 17 September from the NOTER run. As in the CTL run (Fig. 5), there is a mesoscale band of heavy (> 300 mm) precipitation orientated from southwest to northeast just inland of the coast. This suggests that flooding would have occurred even without the coastal hills and Appalachians. Since the large Froude number regime favors ascent over the topography, the precipitation differences between the NOTER and CTL are largest over some of the narrow ridges (Figs. 5, 7), where the precipitation is enhanced over some of the eastward facing slopes by 20-30%. Overall, the terrain impact on the precipitation was of secondary importance to the intense frontogenetical forcing described above in section 4.

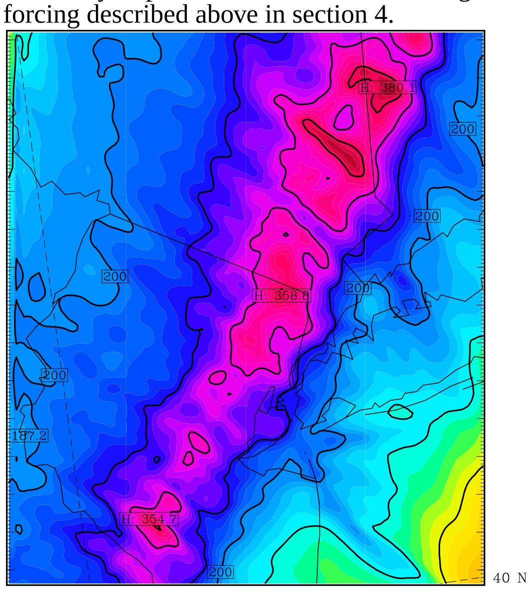

Figure 7. Storm total model precipitation (color shaded in mm) between 0600 UTC 16 September and 0600 UTC 17 September 1999 for the 1.33 km NOTER simulation.

Two sensitivity runs were completed to quantify the role of the diabatic precipitation processes on Floyd's evolution. One experiment designated "NOLH" allowed precipitation to occur, but the latent heating/cooling effects were turned off (i.e., a so-called "fake dry" run), while another "NOEVAP" experiment turned off only the evaporative cooling from cloud and rain water. At 1200 UTC 17 September in the NOLH run (Fig. 8), the surface cyclone is nearly 25 mb weaker than the CTL run (cf. Fig. 3) and shifted over 500 km south of the CTL along the North Carolina coast. The surface temperature gradient in the NOLH to the northeast of Floyd is also only half as large as the CTL. The 500 mb trough was much weaker than the CTL and it became decoupled from the low level circulation given its few hundred kilometer separation downstream (northeast) of the surface low (not shown). The downstream ridge is also much less amplified, with no thermal ridge extending over the Northeast. It is clear that the latent heating from precipitation was critical in maintaining the intensity of Floyd as it went through its ET life cycle.

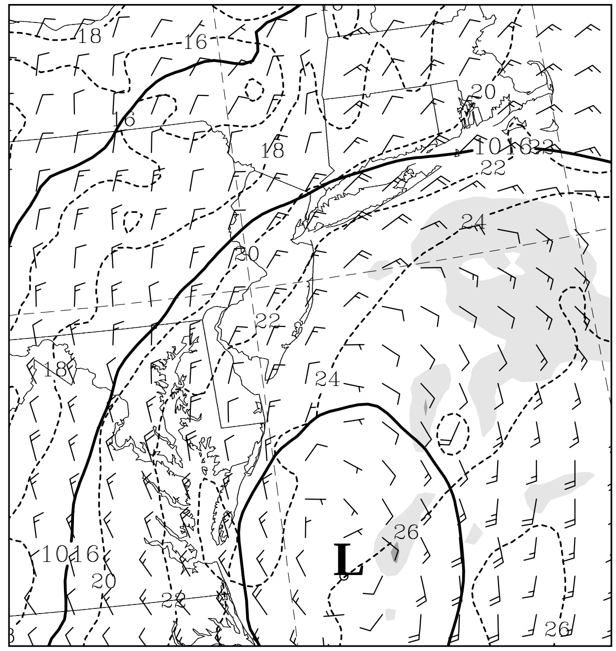

Figure 8. Model surface analysis from the NOLH experiment for the 12-km domain at 1200 (36 h) 17 September 1999 showing sea-level pressure (solid) every 4 mb, surface temperatures (dashed) every 2 ^oC, surface wind barbs (full barb=10 kts), and model reflectivities (shaded).

At 2100 UTC 16 September in the NOEVAP run (not shown), the absence of evaporation from falling precipitation resulted in inland surface temperatures that are around 2 ^oC warmer than in the CTL. This resulted in a somewhat weaker low-level baroclinic zone associated with Floyd, and the central pressure of Floyd is 4-5 mb weaker in the NOEVAP than the CTL. The NOEVAP area of maximum precipitation occurred over approximately the same area as the CTL, but the NOEVAP amounts were 10-20% less than the CTL.

A separate experiment was conducted at 36- and 12-km grid spacing in which the surface heat fluxes were turned off (NOFLX). The frontal temperature gradient to the north of Floyd was about 30% weaker than in the CTL (not shown), and the horizontal extent of the heavy rainfall (> 40 dBZ) was also 30-40% less in the NOFLX experiment. The central pressure of Floyd in the NOFLX run is also 4 mb weaker than the CTL. Overall, surface heat fluxes are not as important as latent heating.

6. CONCLUSIONS

This paper uses high resolution model simulations to diagnose the structural evolution of an extratropical transition (ET) along the East Coast of the U.S. This study focused on Floyd (1999), in which devastating flooding occurred from eastern North Carolina northward to Connecticut.

The MM5 control simulation and subsequent experiments serve to illustrate the complex interrelationships among the thermal, wind, and precipitation structures in the coastal zone. A deep layer of deformation frontogenesis extended from the surface to 400 mb in association with the southeasterly circulation around Floyd at all levels merging with the inland northeasterlies at low levels and southwesterlies aloft. The combination of strong frontogenesis as well as moist symmetric instability below 800 mb and neutrality aloft resulted in a narrow and intense band of precipitation just inland of the coast. A separate simulation without the Appalachians and the coastal terrain resulted in little change in Floyd's pressure and temperature evolution, and only a 10-30% reduction in precipitation over some upslope areas; therefore, terrain played a secondary role in the devastating flooding for this particular event.

The experiments with no latent heating, evaporation, and surface fluxes illustrate the importance of diabatic effects in slowing Floyd's weakening after landfall and enhancing the frontogenetical circulations near the coast. Without latent heating from precipitation, the storm was about 25 mb weaker than the full physics simulation by 24 hours into the simulation, and it only slowly propagated up the coast. Without evaporation from precipitation, the low-level front, with 10-20% weaker and Floyd was about 4 mb weaker. Another simulation without surface heat fluxes also resulted in Floyd being 4-5 mb weaker and 10-20% less precipitation than the control simulation.

In summary, this study illustrates the complex, multi-scale contributions to the ET transition of Floyd; therefore, it is perhaps not surprising that the operational models had difficulty forecasting the evolution and quantitative precipitation for this event. The large difference between the MM5 and operational Eta and AVN models in predicting Floyd's evolution as well as the large diabatic sensitivities within the MM5 suggests that there may be significant model physics sensitivities associated with this event, especially with the convective and boundary layer parameterizations. Future work will investigate the predictability of the location and amount of heavy precipitation for this event as a function of the model parameterizations applied.

7. ACKNOWLEDGEMENTS

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