Chapter 4. The Extratropical Transition of Hurricane Floyd (1999) and its Effects on MAP IOP-2B.

4.1 Introduction

When tropical systems threaten the U.S. coastline, including the Eastern Seaboard and the Gulf Coast, much attention is focused on the location of landfall and the subsequent inland track of the storm. Once the system leaves the U.S., the attention is then focused on the eastern Canadian coast. After leaving Canada and North American waters, the storm is no longer of interest in the U.S. and Canada, and little public information is released about the storm. However, storms leaving the U. S. and Canada can still produce effects far downstream, i.e. Europe, as they undergo extratropical transition (ET). For example, Hurricane Hazel (1954), which devastated the eastern U. S., was tracked as far east and north as Scandinavia (Barnes, 2001). The deadliest natural disaster in U. S. history, the 1900 Galveston Hurricane, was observed as far east as Siberia (Larson, 1999).

As tropical systems make the transition from tropical to extratropical, the tropical systems remnants can interact with extratropical weather systems (Matano and Sekoika (1971; Harr and Elsberry, 2000a, b). Matano and Sekoika (1971a, b) suggested that tropical storms can interact with extratropical systems to form either a complex or compound system. Complex systems may form when the tropical system approaches a preexisting front and causes a frontal cyclone that develops as the tropical system decays. Compound systems may form when a tropical system approaches a preexisting extratropical cyclone (Matano and Sekioka, 1971). The upper-level dynamics of the tropical systems can also interact with
extratropical upper-level features to lead to reintensification of the tropical system or intensification of the extratropical system (Bosart and Lackmann, 1995; Hoskins and Berrisford, 1988).

The relationship between the ET of Atlantic hurricanes and heavy rainfall events in northwest Italy was studied by Pinto et al. (2001). They found that over a period of years between 1958 and 2000, 40% of 30 extreme rainfall events in northwest Italy were affected by Atlantic tropical systems. Pinto et al. (2001) divided these systems into three classes: A) recurring systems over the central North Atlantic, B) recurring systems over the western North Atlantic and C) systems recurving near the U. S. East Coast. It is this class, Class C, which is most relevant to the study presented here. In Class C systems, the recurving storm commonly connects with an upper-tropospheric mid-latitude trough, leading to extreme temperature differences caused by the tropical and extratropical systems (Pinto et al. 2001). Bertò et al. (2003) also performed a study tracking water vapor associated with eastern Italian events and found that in some cases, the water vapor originated off the eastern coast of the U. S. near the track of the Class C systems.

The focus of the study presented in this chapter fits the Class C category. On 16 September 1999, Hurricane Floyd made landfall in North Carolina and propagated northward up the Eastern Seaboard and then recurved northeastward as it became embedded in the flow of an upper-level trough (Fig. 4.1a). Floyd was classified extratropical near New England 1200 UTC 17 September and merged with an existing extratropical low over the Atlantic on 19 September. The upper-level trough associated with the low led to heavy rainfall amounts on 20 September during the Mesoscale Alpine Program (MAP) Intensive Observation Period (IOP)-2B (This event is hereafter referred as IOP-2B).
Even though Pinto et al, (2001) do not refer to Floyd, the resemblance of Floyd to their Category C does deserve an investigation as to whether Floyd did indeed have some impacts on IOP-2B. Not only does Floyd's track and merger with the extratropical low indicate possible effects on IOP-2B, but satellite imagery as early as 16 September, while Floyd was still over North Carolina and Virginia, shows upper-level outflow from Floyd was stretching toward Europe (Fig. 4.1b).

In order to study the impact of Floyd's ET on IOP-2B, numerical simulations over the North Atlantic including the U. S. Eastern Seaboard and northern Europe using the PSU/NCAR MM-5 (V3) model will be used. The setup of the simulations will be discussed in Section 4.3. Section 4.2 will give an overview of the synoptic situation between 18 and 21 September. Results of the numerical simulations will follow in Section 4.4 with summary and conclusions in Section 4.5.

4.2 Synoptic overview

The 12 hourly NOGAPS surface analyses beginning with 18 September (17 September data was missing) are shown in Fig. 4.2. The remnants of Floyd were located over southern Canada and the Atlantic low was west of Ireland. The central pressure of Floyd's remnants was 996 hPa and the pressure of the Atlantic low was 960 hPa (Figs. 4.2a-c). By 1200 UTC 19, the two systems were near merger as they were both enclosed by the same 1008 hPa isobar (Fig. 4.2d). This was six hours before the Tropical Prediction Center (TPC) reported merger time. By 0000 UTC 20 (Fig. 4.2e), the two systems had merged and were fully merged by 1200 UTC 20. At this time the strongest part of the storm was over the British Isles (Fig. 4.2f).
The NOGAPS 300 hPa analyses are shown in Fig. 4.3. A trough-ridge system was located over the Atlantic, spanning from North America to Europe at 0000 UTC 18. Over North America, was the trough that initially interacted with Floyd as it made landfall and propagated up the Eastern Seaboard. A ridge was located over the western Atlantic, extending from 60°W to 20°W. Southwest of Ireland, was the trough that was instrumental in the MAP IOP-2B event (hereafter, referred to as the MAP trough).

At this time, Floyd's surface circulation was located near a two-pronged entrance region of a jet on the eastern side of the North American trough. The primary entrance region was the polar jet associated with the trough while the second region to the east of the primary entrance was a secondary entrance region most likely associated with the remaining outflow from Floyd's remnants (see Ch. 2 of this paper). The large scale jet associated with the ridge then flowed toward Europe and around the MAP trough. At 0000 UTC 18 (Fig. 4.3a), the Atlantic low was located just west of the MAP trough axis (see Fig. 4.2). Also note, east of the MAP trough, was a smaller scale trough with its axis over France. Twelve hours later at 1200 UTC 18 (Fig. 4.3b), the MAP trough became a closed low with a minimum height of 8,790 m (hereafter, referred to as the MAP low).

Along the east coast of North America, the two-pronged jet entrance region could be clearly seen (Fig. 4.3b). The western entrance region or "prong" was the entrance region of the primary large-scale jet associated with the trough. The easternmost region or second "prong" was collocated with the eastern side of the low-level circulation of Floyd's remnants and an inverted trough stretching eastward (Fig. 4.2b). As noted previously, this entrance region was the remaining outflow from Floyd's remnants. Also, any convection associated with the inverted trough may have reinforced this jet. As can be seen in Fig. 4.3b, this
second entrance region was a southern extension of the large-scale jet and remained
colloqued with the low-level circulation of Floyd's remnants.

By 1200 UTC 19 (Fig. 4.3d), the MAP low had filled to 8,940 m and Floyd's remnants
were now approaching the crest of the ridge. The exit region of the upper-level jet was
located over Spain and southern France with diffluence over northwest Italy and the MAP
target area.

At 0000 UTC 20 (Fig. 4.3e), Floyd's remnants were still located just west of the ridge
axis and the exit region of the upper-level jet was near northwest Italy with upper-level
diffluence over the region. It was at this time that the heavier rainfall associated with IOP-
2B was occurring. Twenty-four hours later, at 0000 UTC 21 (Fig. 4.3g), the flow over the
Atlantic was almost zonal with the exit of the jet over France and Germany. By now, Floyd's
remnants and the Atlantic low had merged.

4.3 Numerical simulations

As previously noted, numerical simulations are performed using the PSU/NCAR MM5
model. This case study is suited for numerical modeling in that the event occurs over a data
void region, i.e. the Atlantic Ocean. Numerical simulations will give insight into the event
that the observations cannot.

Two simulations will be performed. The first will be a control simulation, denoted
CNTRL, in which the moist physics of the model will be "turned on." The second
simulation, denoted DRY, will be a fake dry run in which all moisture and latent heat
processes are "turned off." This sensitivity test will allow the indirect effects of the hurricane
to be seen as latent heating was dominated by processes associated with Floyd. A similar
approach to transition modeling was done by Yoshino et al. (2003). These simulations were performed on a 200x220 grid (45 km horizontal spacing) with 45 $\sigma_p$ vertical levels. The grid covered the North and South Atlantic Ocean, North America, Africa, and western Europe (Fig. 4.4). The numerical simulations were initialized at 1200 UTC 17 September, when Floyd was over New England, and integrated for 84 hours, ending at 0000 UTC 21 September. This time span was chosen to capture Floyd before merger with the extratropical low and the time before IOP-2B. To accurately simulate the vortex associated with Floyd, the MM5 vortex bogussing scheme (Low-Nam and Davis, 2001) was used with the location and wind speed of Floyd at 1200 UTC 17 September (43.3°N; 70.6°W; 22.5 m s$^{-1}$). The model simulations' initial fields and 12-hourly boundary conditions were based on the NCEP Reanalysis data.

4.4 Results

Results follow for both the CNTRL and DRY simulations. Sea level pressure, 300 hPa heights and winds, 300 hPa potential vorticity and other fields will be discussed. Section 4.4.1 will discuss the simulated extratropical transition over the North Atlantic, Section 4.4.2 will discuss the simulations' results over Europe in regards to IOP-2B, and 4.4.3 will be a discussion of back trajectory calculations from Europe back to Floyd's remnants.

4.4.1 Extratropical transition

Results follow for the extratropical transition (ET) of Floyd.
4.4.1.1 Sea level pressure and surface winds

The initial sea level pressure field is shown in Fig. 4.5. Note that both the CNTRL and DRY simulations were the same at model initialization. Floyd can be seen over New England with a minimum pressure of 993 hPa, about 9 hPa higher than observations. The extratropical low over the Atlantic is south of Iceland with a minimum pressure of 981 hPa.

Throughout the simulation, Floyd propagated northeast toward the extratropical low while the low initially deepened and propagated to the east (Fig 4.6a). The inverted trough present at 1200 UTC 18 in the NOGAPS analysis (Fig. 4.2b) was also present in the CNTRL simulation (Fig. 4.6b). During the period, the remnants of Floyd continued to weaken and at 1200 UTC 19, the first stages of merger with the extratropical low began as a 1012 hPa isobar enclosed both storms.

At 0000 UTC 20 (Fig. 4.6e), a trough had developed southeast of the Atlantic low, near Spain and southern France. Surface winds near Italy were from the south from the Mediterranean. This southerly flow brought in warm moist air, which impinged upon the Alps, aiding in the development of the IOP-2B rainfall (Sénési et al., 1996; Massacand et al., 1998; Buzzi and Foschini, 2000; Fehlmann and Quadri, 2000; Tripoli et al., 2000; Lin et al., 2001). This trough persisted through 1200 UTC 20.

With latent heat and moisture effects removed, the two systems were 10-12 hPa weaker in the DRY simulation at 0000 UTC 18 September (Fig. 4.7). Twelve hours later, the Atlantic low was 18 hPa weaker than in the CNTRL simulation. Unlike the CNTRL simulation, the closed isobar representing Floyd's remnants was still evident at the end of the simulation (Fig. 4.7g). This may be due to the initial bogussing of the vortex of Floyd, which could counteract the weakening due to no latent heat release. Without latent heat release, it
appeared that the Floyd's remnants became a secondary circulation of the merged low with the Atlantic low the primary circulation.

### 4.4.1.2 300 hPa heights and winds

The initial 300 hPa heights, winds, and isotachs at 1200 UTC 17 September are shown in Fig. 4.5b. At the model initialization, the North American trough was present as well as the ridge and MAP trough. A closed upper-level low was located west of the British Isles. A broad region of strong winds (> 40 m s\(^{-1}\)) associated with the trough stretched from North America toward Europe, almost reaching the Iberian Peninsula. Winds stronger than 70 m s\(^{-1}\) were located south of Greenland. A southern extension of the broad jet was located over New England in association with the remaining outflow of Floyd and the jet associated with the North American low. This outflow merged with the flow around the upper-level ridge and into the trough. By 1200 UTC 18, Floyd's remnants were located just west of the ridge axis, with the southern extension of the jet still in place (Fig. 4.8b). The heights of the upper-level low west of the British Isles had risen by 120 m since 1200 UTC 17. The core of the broad jet over the Atlantic had weakened from 70 m s\(^{-1}\) to 60 m s\(^{-1}\) and the width of the jet exit region had also decreased in size with multiple maxima evident. Over Europe, the height field indicated a secondary trough with the axis just west of Italy.

By 1200 UTC 19 (Fig. 4.8d), the upper-level jet had become narrow, with the entrance region now located to the southeast of Floyd's remnants. By this time also, the surface Atlantic low was located under the upper-level low and began to weaken. When compared to the NOGAPS 300 hPa analysis (8,880 m), the simulated upper-level low (9,120 m) was
about 240 hPa higher than the NOGAPS low and the simulated remnants of Floyd were west of the NOGAPS Floyd remnants.

At 0000 UTC 20 (Fig. 4.8e), the upper-level jet had broken into three jets. One was on the eastern side of the MAP trough, over France. The second, which appeared to still be the main jet, was along the western side of the MAP trough from the ridge axis and stretched around the base of the trough. The third jet was just west of the upper-level ridge axis and was associated with Floyd's remnants. Over Italy, the wind field showed a diffluent pattern that was conducive to upward motion.

By 1200 UTC 20 (Fig. 4.8f), the upper-level ridge was beginning to weaken and an almost zonal flow was present over the Atlantic. The closed low near the British Isles continued to fill in and now the jet was a narrow jet extending from Spain back to south of Canada. The wind field continued to exhibit diffluence over Italy. By 0000 UTC 21 (Fig. 4.8g), the ridge continued to weaken with a new ridge building east of Canada. What had been the North American trough at the beginning of the simulation was now east of Canada.

The DRY 300 hPa heights and winds are shown in Fig. 4.9. At 0000 UTC 18 (Fig. 4.9a), the overall pattern in the heights and winds are similar to the CNTRL for the same time. However, the heights are lower in the DRY simulation. Note that the 8,880 m height contour is larger in area than the same contour in the CNTRL simulation. This is very likely due to the lack of latent heat release, which can increase the mean air temperature, thus increasing the upper-level air thickness and heights, as according to the hypsometric equation.

By 0000 UTC 19 (Fig. 4.9c), the DRY upper-level jet is much smaller in area than the CNTRL jet at the same time. The height fields also were smoother with heights continuing to be lower in the DRY simulation. Twelve hours later, at 1200 UTC 19 (Fig. 4.9d), the
upper-level jet was located between the upper-level ridge axis and MAP trough axis, with maximum winds just over 50 m s⁻¹. Winds in the CNTRL simulation were 60 m s⁻¹.

At 0000 UTC 20, the upper-level jet had all but disappeared in the DRY simulation (Fig. 4.9e). Over Italy, there was very little if any upper-level diffluence over the MAP target area. Twelve hours later (Fig. 4.9f), there still was very little diffluence over Italy and the upper-level height and wind fields were very smooth. This continued through the end of the simulation at 0000 UTC 21 (Fig. 4.9g). The DRY simulation showed that without the moisture effects and latent heat release, the upper-level jet could not be sustained.

4.4.1.3 300 hPa potential vorticity

The initial 300 hPa potential vorticity (PV) distribution is shown in Fig. 4.5c. The PV maxima associated with Floyd and the extratropical low are evident over the northeast U. S. and the Atlantic respectively. The evolution of the PV field for the CNTRL simulation can be seen in Fig. 4.10. As the simulation progressed, the PV of Floyd and the upper-level trough began to merge and by 0000 UTC 19, the PV of both systems was contained in one large area (Fig. 4.10c). The PV associated with the trough contained two local PV maxima west of England and Spain. It can be seen that the PV from Floyd was advected around the upper-level ridge, joined a second PV streamer from Greenland and the two PV streamers advected into the trough (Figs. 4.10d-f). This continued for the rest of the simulation.

In the DRY simulation, the PV distribution was weaker than the CNTRL simulation (Fig. 4.11). This was expected since there was no latent heat release and generation of PV is an indicator of latent heat release. At 0000 UTC 18, the PV of Floyd's remnants and the upper-level trough were merging, however, earlier than the CNTRL simulation. When compared to
the CNTRL PV, the DRY PV is less concentrated. This can be seen by comparing the North Atlantic trough in both simulations.

4.4.2 IOP-2B

In order to determine the effects of latent heat release and moisture on IOP-2B, this section will focus on the simulations' results over southern Europe, encompassing the MAP study area, the Lago Maggiore Target Area (LMTA). Results will focus on 20 September during the period of observed heavy rainfall.

The CNTRL 6-hour rainfall totals for the period 0900 UTC through 1800 UTC 20 September are shown in Fig. 4.12. In the simulation, the rainfall did not reach the western Italian Alps, until around 0900 UTC 20 September (Fig. 4.12a). The maximum rainfall occurred between 1500 UTC and 1800 UTC, about 15 hours later than the observations (see Ch. 3). This lag in the rainfall development is mostly likely due that this time period was near the end of the simulation period (~72 hours) as well as the relative coarseness of the model horizontal resolution. Note that the distribution of the rainfall followed the east-west slopes of the Alps, as seen in the simulations presented in Chapter 3. With the DRY simulation, there was no rainfall, since moisture effects were removed from the simulation.

The CNTRL and DRY sea level pressure and surface winds are shown in Fig. 4.13 beginning at 0900 UTC and ending at 1500 UTC 20 September. At 0900 UTC, the circulation of the MAP low could be seen along the western coast of France. Over the Mediterranean Sea, a secondary trough was evident just south of France. A surface confluence zone was located in the Piedmont region (the concave region of the Alps in Italy) of Italy as southern winds associated with the trough in the Mediterranean Sea became
juxtaposed with easterly and southeasterly winds from eastern Italy and the Adriatic Sea (Fig. 4.13a). The easterly winds were caused by the southeasterly winds from the Adriatic impinging upon the eastern Alps and turning to the west. Already along the confluence zone, rainfall had begun to form (Fig. 4.12).

The DRY sea level pressure field was much different than the CNTRL. A much weaker Atlantic cyclone was located over England, west of its position in the CNTRL simulation. Over Africa, a surface ridge was present where the Atlantic low had extended in the CNTRL simulation. The ridge east of Italy had also been present in the CNTRL simulation but was much stronger in the DRY simulation. Over Italy, the surface wind field was similar to the CNTRL, with a weaker confluence zone over the Piedmont.

The same general pattern held for both simulations through 1500 UTC. At 1200 UTC, the confluence zone in the CNTRL simulation was better defined with more easterly winds becoming juxtaposed with the southerly winds. In the DRY simulation, winds were more southerly in the Piedmont region and the confluence zone was weak if existent at all. This lack of confluence in the DRY simulation existed through 1500 UTC.

A key ingredient for Alpine rainfall events is the presence of a low-level jet (LLJ) to transport warm moist air into the concave region of the Alps (Sénési et al., 1996; Buzzi et al., 1998; Massacand et al., 1998; Buzzi and Foschini, 2000; Fehlmann and Quadri, 2000; Tripoli et al., 2000; Lin et al., 2001). In the CNTRL simulation the LLJ can be found in the 850 hPa wind field (Fig. 4.14). AT 1200 UTC 20 September, the LLJ was approaching the western Italian coast and three hours later, the LLJ jet was impinging on the slopes of the Alps near the Italian-Swiss border. Also, at 1200 UTC, there appeared to be a well developed
confluence zone in the Piedmont region as southerly winds became juxtaposed with easterly winds. For both 1200 UTC and 1500 UTC, the LLJ was located just east of the heaviest rainfall.

In the DRY simulation at 1200 UTC, the winds were weaker than the CNTRL winds by about 10 m s\(^{-1}\) near the Italian coast (Fig. 4.14c). The DRY simulated 850 hPa heights were also much higher, approximately 120 m, than in the CNTRL simulation, indicative of a weaker trough in the DRY simulation. Also, a confluence zone was not present over the Piedmont region. Weaker winds persisted through 1500 UTC. Without moisture and latent heat effects, the wind speeds were lower as the trough was weaker in the DRY simulation. The elimination of convection in the DRY simulation could have also weakened the inflow into Italy, as the convection would have led to convergence and more inflow into the region. Buzzi et al. (1998) showed that latent heat was important to the flow pattern around the Alps.

Since the low-level jet's role is to transport warm moist air into the region, the 850 hPa equivalent potential temperature (\(\theta_e\)) distribution was analyzed. Collocated with the LLJ was a tongue of high \(\theta_e\) air extending from Italy southward toward the North African coast (Fig. 4.15). This tongue represented warm moist air being transported from the Mediterranean toward the Alps. The high \(\theta_e\) air impinged on the mountains and ascended, leading to the rainfall seen in Fig. 4.12. In the DRY simulation, because of the lack of moisture and the LLJ, there was no tongue of high \(\theta_e\) air. In fact, \(\theta_e\) values never exceeded 312 K (Fig. 4.15).

The upper-air environment is also important to the development of Alpine rainfall events. The presence of an upper-level jet (ULJ) and trough can lead to upward motion and possible rainfall via ageostrophic circulations, i.e., the ULJ's circulation can become coupled with the LLJ's circulation, resulting in enhanced upward motion.
The simulated 300 hPa heights and winds are shown in Fig. 4.16 for 1200 UTC and 1500 UTC. At 1200 UTC, both simulations showed evidence of upper-level diffluence over Italy. The height field in the DRY simulation was also smoother than the CNTRL heights and the DRY simulated heights were also lower than the CNTRL. This was because of the lack of convection and latent heat release leading to column warming and higher heights. Over Italy and just south of Italy winds were about 5 m s\(^{-1}\) stronger in the CNTRL simulation and southerly as the trough was actually more pronounced and had more curvature.

At 1500 UTC, diffluence existed in both simulation but DRY wind speeds were weaker than the CNTRL winds. Over Italy, the strongest winds in the CNTRL simulation were located over the western portion of the Alps with speeds of about 40 m s\(^{-1}\). This appeared to be a local jet streak. These locally strong winds were located northwest of the LLJ. In the DRY simulation, there did not appear to be a jet streak as wind speeds over the Alps were fairly uniform at 30 m s\(^{-1}\) (Fig. 4.16). Also, the height gradient continued to be fairly uniform and smooth over the domain in the DRY simulation.

In order to see if the upper and low-level flows had become coupled and how the lack of latent heating and moisture affected the coupling, a vertical cross section was taken across northern Italy near the LLJ (see Fig. 4.14 for location). The isentropes, ageostrophic circulations, and vertical velocity fields across the cross section are shown in Fig. 4.17. At 0600 UTC (Fig. 4.17a), upward motion was located along the western slopes of the Alps as air impinged on the slopes and was forced upward. A large circulation, covering almost the entire cross section was anchored around the 300 km mark and between 700 and 800 hPa. Judging by the orientation of the isentropes, this circulation appeared to be thermally direct. The low-level flow of the circulation was from the east toward the western Alps and
Piedmont. This is the same easterly flow seen at the surface and 850 hPa levels (Figs. 4.13 and 4.14). The rising branch was over the mountains, as previously noted, and the return flow was from the west at the upper levels.

The area of upward motion, as indicated by the vertical velocities, grew in coverage and propagated east with time. By 1500 UTC (Fig. 4.17e), the maximum upward motion was located over the Piedmont between 700 and 600 hPa. At this time, the LLJ was located almost midway within the cross section (see Fig. 4.14). The location of the LLJ was coincident with the location of the rising branch of the low-level circulation. The location of the maximum upper-level winds in the cross section was determined by another cross section showing the isotachs and horizontal winds (not shown). The maximum upper-level winds were located near the 100 km mark around 300 hPa (denoted by the U). Coincident with the upper-level maximum winds was a possible weak circulation with weak rising motion juxtaposed with the rising branch of the low-level circulation. From this cross section, it would appear that there was some coupling occurring between the upper and low-level flow fields.

In the DRY simulation, at 0600 UTC (Fig. 4.17b), the vertical velocities were much weaker than the CNTRL, only reaching 5 cm s$^{-1}$. The low-level circulation was also much weaker. By 1500 UTC, the DRY simulated vertical velocities were still weaker than the CNTRL and the distribution of vertical velocities appeared to tilt to the west with height. From the isentropes, the lower atmosphere was well mixed. The low-level circulation was weak and only extended to 700 hPa. There was no LLJ (see Fig. 4.14) or locally strong upper-level winds. Based on this and the weakness of the rising branch as well as the absence of an upper-level circulation, there was no coupling of the upper and low-level flow.
fields. Without latent heat and moisture effects, there was no LLJ, which reduced the low-level circulation. Without convection within the simulation, the vertical velocities were much lower. Also without the convection, the upper-level height field was more uniform with no local jet streak over the Alps.

4.4.3 Back trajectories

It has been shown that removing the effects of latent heat and moisture had a tremendous impact on the mesoscale environment of IOP-2B and the extratropical transition of Floyd. However, these findings have not established a link between the two events. To aid in establishing a link between IOP-2B and Floyd's ET, 64 back trajectories from northwest Italy were calculated for the period 0900 UTC 20 back to 1800 UTC 18 September from an initial pressure level of 300 hPa. The start time, 0900 UTC was chosen because this was a time just before the heaviest rainfall totals in northwest Italy. The trajectories were calculated for both the CNTRL and DRY simulations and the resulting trajectories are shown in Fig. 4.18. The upper-level trough and ridge pattern can clearly be seen for both sets of trajectories and appear similar to the results of Bertò et al. (2003). The trajectories for the DRY simulation were flatter than the CNTRL, showing that the DRY trough and ridge were weaker than the CNTRL. The 1800 UTC 18 positions of the CNTRL trajectories were located over the low-level circulation of Floyd and the southern extension of the upper-level jet, which was the remaining outflow of Floyd (Fig. 4.19a). The DRY trajectories were located east of these positions over the DRY simulated Floyd remnants (Fig. 4.19b) but there was no southern extension of the polar jet. Also, all of the DRY start positions were clustered around the center of Floyd's remnants while the CNTRL positions were located over Floyd's center and
downstream within the polar jet. From both sets of trajectories, it was highly likely that there was some type of link or influence between Floyd's ET and IOP-2B in that the parcels over the Alps during the heavy rainfall had their origins from Floyd's remnants.

Even though there was a link between Floyd's ET and IOP-2B for both simulations, the parcel characteristics were different between the two simulations. Several parcel diagnostics were calculated as the back trajectories were calculated (Fig. 4.20). These diagnostics included absolute vorticity, potential vorticity, wind speed, parcel height, and vertical velocity. One detail that was readily apparent was the greater variability of the parcel characteristics among the CNTRL trajectories. The DRY parcels were less variable among each other and had smoother trends with time.

For the most part, the absolute vorticity of the CNTRL parcels was greater than the DRY absolute vorticity. There was greater variability among the CNTRL parcels as well. With the DRY parcels, a sinusoidal pattern could be seen in the vorticity fields as the parcels moved over the ridge and trough over the Atlantic. The sinusoidal pattern was harder to detect in the CNTRL trajectories even though evidence was there. With the higher vorticity values, the CNTRL simulation was more likely to have rising motion under the trajectories.

In agreement with Fig. 4.10, the CNTRL parcels had greater PV values than the DRY parcels (Fig. 4.20b). The DRY parcels had values between one and two PVU with little variability with time. CNTRL parcels also propagated faster than the DRY parcels (Fig. 4.20c). This showed how the DRY winds were weaker than the CNTRL winds, explaining why the DRY parcels were farther east in origin than the CNTRL parcels. At 69 hours, some CNTRL parcels were approaching 40 m s⁻¹. Recalling Figs. 4.16 and 4.17, these parcels were most likely associated with the upper-level winds discussed in the cross section.
The time series of parcel heights showed little variability in the DRY trajectories (Fig. 4.20d). DRY parcel heights varied between 8.5 and 10.5 km while the CNTRL trajectories varied from approximately 10 km to 4.5 km. Consistent with Fig. 4.17, the CNTRL parcels had stronger vertical velocities than the DRY parcels (Fig. 4.20e). It appeared near the time of the rainfall, that most of the DRY parcels had a negative vertical velocity and were actually descending while some CNTRL parcels were ascending.

4.5 Summary and conclusions

Two simulations of Hurricane Floyd's ET were performed to see if there was link between Floyd's ET and the MAP IOP-2B rainfall event in September 1999. The first simulation, CNTRL, was a full physics simulation with latent heating effects included. A second simulation, DRY, was performed with latent heat effects removed. This simulation would weaken Floyd thus measuring its effects on IOP-2B.

Simulations showed that Floyd's remnants merged with an extratropical cyclone west of the British Isles on 19 September. The DRY simulated Floyd and extratropical low were much weaker than the CNTRL systems. At the upper levels, in the CNTRL simulation, a southern extension of the polar jet, associated with Floyd's remaining outflow existed, transporting parcels from Floyd into the upper-level polar jet and toward Europe. This simulated southern extension was consistent with the observed winds from the NOGAPS analysis. In the DRY simulation, this southern extension of the polar jet did not exist and the polar jet was much weaker than the CNTRL jet. Results over Europe showed that without latent heat release, the LLJ and tongue of high $\theta_e$ air did not develop to aid in rainfall formation, as these were dependent on the pre-trough flow that was weaker in the DRY
simulation. At upper levels, winds in the DRY simulation were more uniform spatially and an upper-level jet streak did not develop over northwest Italy as in the CNTRL simulation. Cross sections revealed that the ageostrophic circulations associated with the upper-level jet streak and LLJ became coupled, creating a region of maximum upward motion in the CNTRL simulation. Again, without latent heat release from convection, this coupling did not occur in the DRY simulation.

Back in time trajectories showed that indeed there was a link between Floyd's remnants and IOP-2B in both simulations. The major differences were that parcel characteristics differed between the two simulations, in particular wind speed and vertical velocity, thus affecting the dynamics of IOP-2B.

The results of this study support the work of Pinto et al. (2001) and Bertò et al. (2003) that Atlantic hurricanes can affect Alpine events and that air parcels can travel from the eastern coast of North America to the Alps and affect events in that region. The study also provides evidence that once the storms leave the U. S. and Canada, they should continued to be monitored as they can affect regions in Europe other than the British Isles or the west coast of continental Europe.

Further research on Floyd's ET and IOP-2B will involve removing Floyd's circulation from the initial model state. This can be done by using PV inversion or by using the MM5 bogussing scheme (Low-Nam and Davis, 2001) to remove Floyd's vortex in the initial conditions. Removing Floyd's circulation and only leaving the background flow we can obtain a better measure of the effects of Floyd on the MAP low and IOP-2B.
Figure 4.1  a) NHC 6-hourly track of Floyd 7 - 19 September and b) GOES-8 visible image for 1445 UTC 16 September. Upper-level outflow from Floyd is indicated.
Figure 4.2. NOGAPS sea level pressure (hPa) and surface winds (knots) for a) 0000 UTC 18, b) 1200 UTC 18, c) 0000 UTC 19, d) 1200 UTC 19, e) 0000 UTC 20, f) 1200 UTC 20, and g) 0000 UTC 21 September 1999. Isobars are solid green lines. One half barb equals 5 knots and a full barb equals 10 knots.
Figure 4.2. Continued.
Figure 4.2. Continued.
Figure 4.3. NOGAPS 300 hPa heights (m) winds (knots), and isotachs (knots) for a) 0000 UTC 18, b) 1200 UTC 18, c) 0000 UTC 19, d) 1200 UTC 19, e) 0000 UTC 20, f) 1200 UTC 20, and g) 0000 UTC 21 September 1999. Heights are white lines, and isotachs are shaded with black lines. A half barb equals 5 knots, a full barb equals 10 knots and a pennant equals 50 knots. Heights are contoured every 120 m.
Figure 4.3. Continued.
Figure 4.3. Continued.
Figure 4.4. Model domain of CNTRL and DRY simulations.
Figure 4.5. Initial conditions at 1200 UTC 17 September for both the CNTRL and DRY simulations for a) sea level pressure (hPa) and surface winds (m s$^{-1}$), b) 300 hPa heights (m) (solid lines), winds (m s$^{-1}$) and isotachs (shaded) and c) 300 hPa potential vorticity (PVU). A half barb equals 5 m s$^{-1}$, a full barb equals 10 m s$^{-1}$, and a pennant equals 50 m s$^{-1}$. 
Figure 4.6. CNTRL sea level pressure (hPa) and surface winds (m s$^{-1}$) for a) 0000 UTC 18, b) 1200 UTC 18, c) 0000 UTC 19, d) 1200 UTC 19, e) 0000 UTC 20, f) 1200 UTC 20, and g) 0000 UTC 21 September. Wind barb convention follows that of Fig. 4.5.
Figure 4.6. Continued.
Figure 4.6. Continued.
Figure 4.7. Same as Fig. 4.6 except for DRY simulation.
Figure 4.7. Continued.
Figure 4.7. Continued.
Figure 4.8. CNTRL 300 hPa heights (every 120 m), isotachs (every 10 m s\(^{-1}\)), and winds (m s\(^{-1}\)) for a) 0000 UTC 18, b) 1200 UTC 18, c) 0000 UTC 19, d) 1200 UTC 19, e) 0000 UTC 20, f) 1200 UTC 20, and g) 0000 UTC 21 September. Heights are solid lines and isotachs are shaded with dashed lines. Wind barb convention follows that of Fig. 4.5.
Figure 4.8. Continued.
Figure 4.8. Continued.
Figure 4.9. Same as Fig. 4.8 except for DRY simulation.
Figure 4.9. Continued.
Figure 4.9. Continued.
Figure 4.10. CNTRL 300 hPa potential vorticity (every 0.5 PVU) for a) 0000 UTC 18, b) 1200 UTC 18, c) 0000 UTC 19, d) 1200 UTC 19, e) 0000 UTC 20, f) 1200 UTC 20, and g) 0000 UTC 21 September.
Figure 4.10. Continued.
Figure 4.10. Continued.
Figure 4.11. The same as Fig. 4.10 except for DRY simulation.
Figure 4.11. Continued.
Figure 4.11. Continued.
Figure 4.12. CNTRL simulated 6-hour rainfall (mm) for a) 0900 UTC, b) 1200 UTC, c) 1500 UTC, and d) 1800 UTC 20 September. The heavy solid black line is the 1 km terrain contour. Rainfall contour interval is every 10 mm.
Figure 4.13. CNTRL and DRY sea level pressure (hPa) and surface winds (m s\(^{-1}\)) for 0900 UTC through 1500 UTC 20 September. Isobars are denoted by solid lines and wind barbs follow the convention of Fig. 4.5. Times and simulation names are listed in each panel in lower left and right respectively. Heavy solid line is 1 km terrain contour.
Figure 4.14. CNTRL and DRY 850 hPa heights (m), isotachs (m s$^{-1}$), and winds (m s$^{-1}$) for 1200 UTC and 1500 UTC 20 September. Heights are denoted by solid lines and wind barbs follow the convention of Fig. 4.5. Times and simulation names are listed in each panel at lower left and right respectively. Line AB denotes the location of the cross section in Fig. 4.17. Wind speeds greater than 20 m s$^{-1}$ are shaded. Heavy solid lines are 1 km terrain contour.
Figure 4.15. As for Fig. 4.14 except for 850 hPa equivalent potential temperature (K).
Figure 4.16. As for Fig. 4.14 except for 300 hPa. Heights are contoured every 60 m.
Figure 4.17. CNTRL and DRY cross sections across AB for 0600 UTC, 1200 UTC, and 1500 UTC 20 September. Shaded contours are vertical velocity (every 5 cm s⁻¹), solid lines are isentropes (every 2 K), and arrows represent the ageostrophic circulations. LLJ and U denote the low level jet and maximum upper level winds in the cross section respectively. Times and simulation names are in the lower left and right of each panel.
Figure 4.17. Continued.
Figure 4.17. Continued.
Figure 4.18. 39-hour back trajectories from 0900 UTC 20 (forecast hour 69) to 1800 UTC 18 September (forecast hour 30) from northwest Italy for CNTRL (black) and DRY (red) simulations. Trajectories are plotted as forward trajectories from the 1800 UTC 18 position.
Figure 4.19. 1800 UTC trajectory positions (dots), sea level pressure (hPa), and 300 hPa isotachs (m s\(^{-1}\)) for a) CNTRL and b) DRY simulations. Isobars are every 4 hPa and denoted by solid black lines. Isotachs are shaded (\(> 40 \text{ m s}^{-1}\)) and denoted by dashed lines.
Figure 4.20. CNTRL (black) and DRY (red) back trajectory diagnostics: a) absolute vorticity ($x 10^{-5}$ s$^{-1}$), b) potential vorticity (PVU), c) wind speed (m s$^{-1}$), d) parcel height (km), and e) vertical velocity (cm s$^{-1}$).