ROLE OF THE GULF STREAM ON EXTRATROPICAL CYCLOGENESIS

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1. INTRODUCTION

Rapidly intensifying coastal storms, sometimes called Nor'easters, develop along the Atlantic states during the fall, winter and spring months. These extratropical cyclones produce gale force winds, heavy snow, ice, and coastal storm surges with intense beach erosion, and are responsible for severe property damage along the eastern seaboard. The majority of these storms form within the coastal zone from South Carolina to Virginia. This region is unique in its position adjacent to the warm waters of the Gulf Stream. Extensive research has been done on the pre-storm marine boundary layer and air-sea interactions associated with extratropical cyclones (e.g. Bosart et al. 1972; Sanders and Gyakum 1980; Kuo and Low-Nam 1990).

This paper discusses air-sea heat exchange and it's horizontal gradient, processes that contribute to the formation of surface cyclones along the east coast of the United States. This interaction occurs when cooler continental air moves over warmer water. The sensible heat flux into the atmosphere can aid in the formation of the offshore closed circulation and alter the track of a surface low. The southeasterly facing coastline of the Carolinas yields a favorable angle for the perpendicular offshore flow typical of the winds from a cold-air outbreak (Wayland and Raman 1989). These winds further enhance the already large thermal contrast resulting in large marine boundary layer baroclinicity. Fantini (1991) has shown that this pre-storm destabilization may act to significantly increase the likelihood for subsequent rapid cyclogenesis. This rapid growth can lead to the formation of an intense winter cyclone (e.g. Holt and Raman 1990; Kuo et al. 1990; Vukovich et al. 1991).

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Within a hundred kilometers offshore of the coast lies the Gulf Stream, which in the fall, winter, and spring months, has a sea surface temperature warmer than those of the near coastal waters. This temperature gradient causes a highly baroclinic region in the marine boundary layer (MBL) especially when the Gulf Stream is closer to the coast. The degree of MBL baroclinicity is dependent on the ratio of the offshore-onshore air temperature difference to the distance of the Gulf Stream Front to the coast. These horizontal thermal gradients can result in the rapid destabilization of the MBL within the Gulf Stream region (Cione et al. 1998). Cione et al. (1993) show that the pre-storm baroclinicity, which includes the pre-storm Gulf Stream front (GSF) position, sea surface temperatures, and average coastal air temperatures, is strongly correlated to the intensification of coastal cyclones. Results from this study reveal that both the thermal structure of the continental air mass and the position of the GSF, in relation to land, are linked to the rate of surface cyclonic intensification. The distance of the GSF from the coast can fluctuate significantly. Lateral meandering of the GSF can cause distances from the coast vary from 15 to 120km (Cione et al. 1993).

The interaction of upper-level potential vorticity anomalies with the lower-level thermal advection can induce cyclogenesis. However, in the Gulf Stream region, it is important to know to what degree these factors promote storm development. A phase shift exists between the upper and lower-level anomalies that allows mutual interactions between the levels, in which the circulations associated with the upper-level potential vorticity anomalies enhance the surface disturbance and vice versa (e.g. Hoskins 1991; Huo et al. 1999). The association of various potential vorticity perturbations with the contributions of thermally induced baroclinicity must be studied to better understand individual forcing mechanisms driving their mutual interactions. The objective of this study is to investigate the role of the Gulf Stream in the development of east coast extratropical cyclones using the 24-25 January 2000 East Coast storm, as well as a 20 year climatology of extratropical cyclones that occurred along the southeast coast of the United States.

2. ASCII AND PSBI

The Atlantic Surface Cyclone Intensification Index (ASCII) is a forecast index that quantifies the amount of low-level baroclinicity off the coast of the Carolinas during a cold air outbreak (CAO). ASCII is based on the gradient between the coldest 24 hour average temperature during a CAO and the temperature of the GSF. The resulting pre-storm baroclinic index (PSBI) is used to forecast the probability that a cyclone in the domain will exhibit rapid cyclogenesis. PSBI values less than 1.0°C/10km indicate that a storm in the domain would be unlikely to undergo explosive development while values greater than 1.7°C/10km indicate that there is a strong chance for rapid cyclogenesis.

The initial ASCII data covered years 1980-1990 (Figure 1: Cione). This dataset was recently expanded to cover the years 1991-2002 (Figure 1: Jacobs).

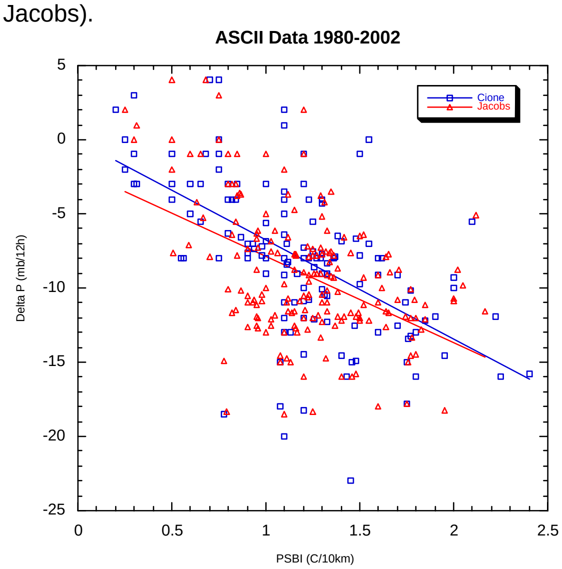

Figure 1. The updated ASCII dataset (1980-2002) of ETC's ΔP/12h vs. PSBI. The linear regression fits are for the separate dates of storms (1980-1990 and 1991-2002).

Storms that formed or tracked within a zone off the coast of the Carolinas were catalogued during the months October through April. The deepening rate and PSBI were recorded for each storm. A second linear regression fit was done for the additional set of storms. This fit verifies the original fit, as well as the PSBI. The 1980-1990 dataset and the 1991-2002 dataset have correlation coefficients of .52 and .56 respectively. This suggests that about 50% of the variance is explained by the PSBI.

The second set of ASCII storms were separated into bins based on the strength of the nearest maximum of 500mb absolute vorticity associated with the surface low (Fig. 2). The maximum absolute vorticity value for each storm was selected during the same 12h period used for ΔP in the ASCII dataset. In order to achieve similar quantity in terms of the amount of storms per bin, the absolute vorticity values covered by each bin are greater at both extremes. For example, bin "16" is by itself because there were several storms with a maximum absolute vorticity greater than $15x10^{-5}s^{-1}$ and less than $17x10^{-5}s^{-1}$. Likewise, bins "21+" and "13-" were grouped because of the lack of storms with maximum absolute vorticities above $20x10^{-5}s^{-1}$, and below $14x10^{-5}s^{-1}$. Tropical to extratropical transition storms were not included.

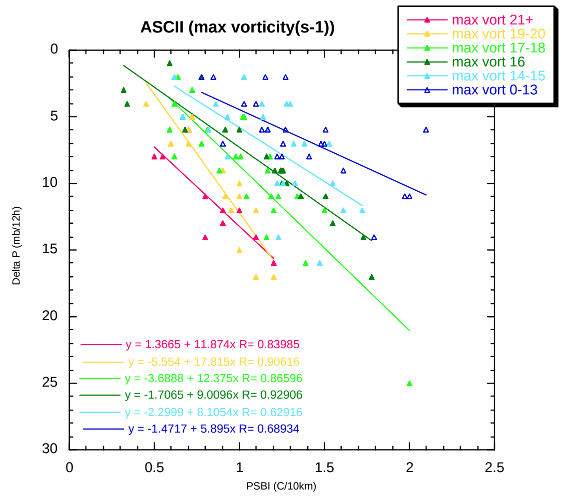

Figure 2. The ASCII dataset (1991-2002) of ETC's (ΔP/12h vs. PSBI) broken down into bins of 500mb vorticity ($10^{-5}s^{-1}$).

Linear regression fits were done for each bin in Figure 2. The decrease in slope with lower 500mb vorticity, as well as the stratified bin positions, suggest that there is a shifting degree of mutual dependency on surface versus upper level forcing. The correlation coefficients range from .63 to .93 suggesting that there is a stronger dependence with weaker upper forcing for bins in the middle. For storms in the "21+" bin, the increasing cyclonic vorticity advection with height outweighs the PSBI to the point where it does not follow the other bin's slope trends as closely.

3. CASE BACKGROUND

Four days prior to the 24 January 2000 storm development, a small coastal low tracked northeast from Cape Hatteras to New England. A Canadian high pressure extended southeastward behind this first coastal low. The strong northwesterly winds advected the cold air mass off the coast of the Carolinas and over the Gulf Stream. The 24h temperature observations beginning 12Z Jan 20 decreased as much as 15°C in the coastal region between Wrightsville Beach and Morehead City, NC. This cold offshore flow remained in place until 20Z Jan 23, and set the stage for an intense winter storm that formed off the southeastern coast of the United States on 24 Jan 2000. The winter storm brought heavy snowfall from the Carolinas through the New England region. Record snow amounts fell across North Carolina with the RDU airport reporting a snowfall accumulation in excess of 20 inches. This event broke the previous snowfall record for a single storm and established a new monthly total accumulation record for RDU.

The surface low began to form northeast of Florida, and downstream from an upper level trough on 24 January 2000. As it moved over the Gulf Stream from Charleston, SC to Cape Hatteras, NC, the pressure dropped at a rate in excess of 1.3mb/hr. A closed circulation formed off South Carolina around 15Z on Jan 24, and moved northeast along the coast following a frontal boundary that set up along the temperature gradient formed by the western boundary of the Gulf Stream. The surface low continued to move north and was located east of New England by 00z Jan 26 in conjunction with an upper level circulation (not shown).

Prior to the explosive development, the Eta, MRF, and AVN model runs failed to predict not only the track, but the deepening rate and the precipitation amount for the event (e.g. Zhang et al. 2002; Buizza and Chessa 2002). Most forecasts placed the storm far to the east and called for less than 4 to 6 inches of accumulated snowfall. Although the models listed above showed the development of a nor'easter type storm, the intensity and precipitation were grossly under predicted. During this event, the GSF was less than 50-60km off the shoreline. As a result, the ASCII index was predicting rapid cyclogenesis. The pre-storm baroclinic index was estimated to be greater than 2°C /10km.

4. MODEL DISCRIPTION

The parameters for the control and experimental simulations were identical with the exception of the SST input dataset. The simulations were initialized at 00Z 24 Jan 2000.

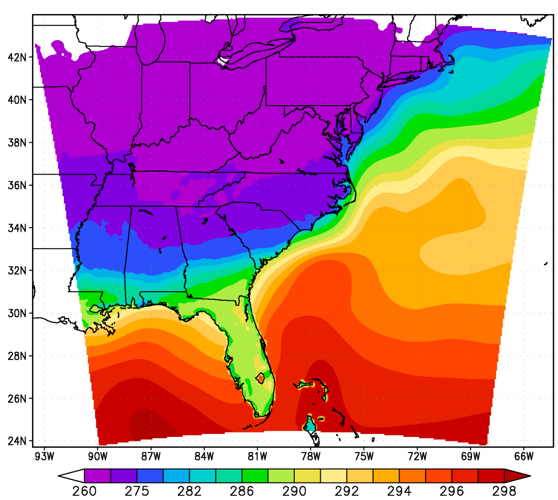

Figure 3. The control simulation's SST input file of NCEP's 2.5° resolution data.

The domain of 10 km grid spacing, shown in Figures 3 and 4, extended well beyond the Gulf Stream region. Both simulations were run for 48h stopping at 00Z 26 Jan 2000. The control simulation's SST input file (Figure 3) used NCEP's 2.5° resolution data. The experimental simulation used 1.4km high resolution data (Figure 4). The experimental SST data were derived from digital images acquired by the Advanced Very High Resolution Radiometer (AVHRR).

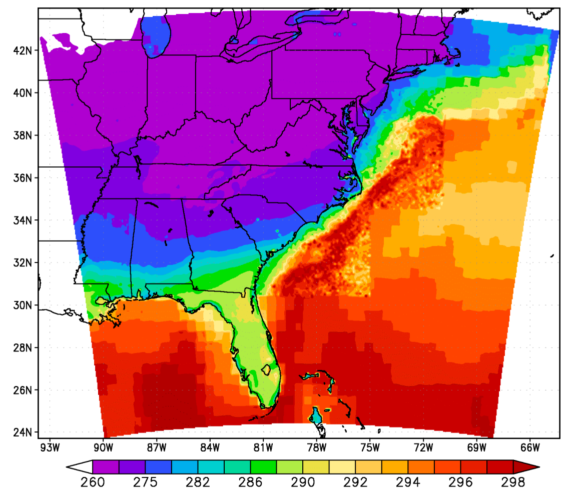

Figure 4. The experimental simulations SST input file of 1.4km grid spacing.

First, a single pass 1.4km resolution dataset was found by looking at imagery prior to the cold air outbreak with as little cloud cover as possible. The chosen image was from 22 January 2000, less than two days prior to the start of the simulation. This was soon enough to reveal the influencing features, but prior to the increase in cloud cover. This image had less than 10% clouds over the Gulf Stream region. To fix this problem of cloud cover, an interpolation routine was used to plot grid points based on the closest non-flagged values without clouds on all four sides. This was done for the cloud fringe data as well. Once the cloud free SST dataset was constructed, it was mapped over the corresponding grid points in the SST dataset of the control run and then used to initialize the experimental simulation. Ship and buoy observations were compared against the imagery in the dataset to validate the abnormally high SST region off the southeast coast of NC (not shown).

5. RESULTS

The preliminary results discussed below include the change in track for both the control and the experimental simulations, as well as the Gulf Stream's influence on vertical storm structure. In both simulations, the sea level pressure showed no significant variation from the observations until 12 hours into the simulation. After 12 hours, the motion, or speed of the center, of the low in the experimental simulation was faster and closer to the observed position. The lateral position of the storm in the experimental simulation was to the west of the control simulation, and closer to the observed track. However, both simulations begin to lag the observed forward speed of the storm beyond this time. When the tracks begin to separate, the pressure drop can be observed. The first major difference between the simulations begin to arise beyond 20Z Jan 24. At this time, the experimental simulation (Figure 5) is tracking the center of the low pressure system closer to the GSF where a sharp 3mb pressure drop occurs.

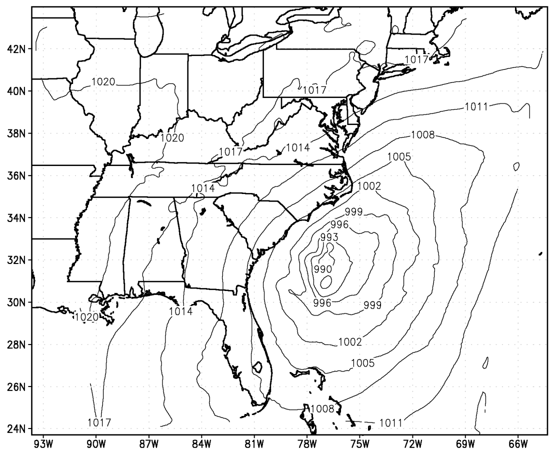

Figure 5. The experimental simulation's sea level pressure valid 04Z 25 January 2000.

However, in the control simulation (Figure 6), the low is not accelerating at the same rate, nor does it experience the abrupt pressure decrease.

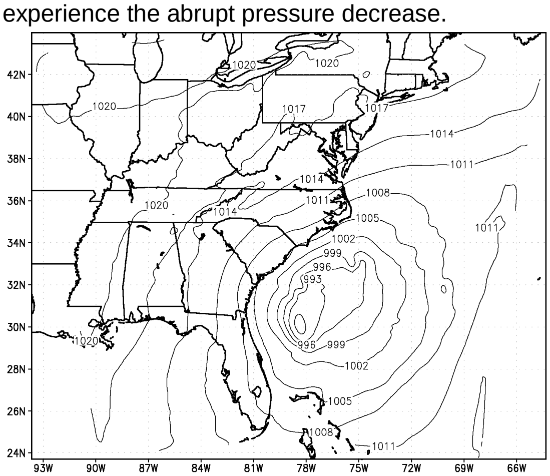

Figure 6. The control simulation's sea level pressure valid 04Z 25 January 2000.

Since the western edge of the GSF is not adequately resolved in the control simulation, there is a lack of surface level convergence in the region above where the GSF would be located (Figure 7). Plots of 10m wind vectors and 2m temperatures show typical convergence extending off the northeast quadrant of the low pressure system in both simulations.

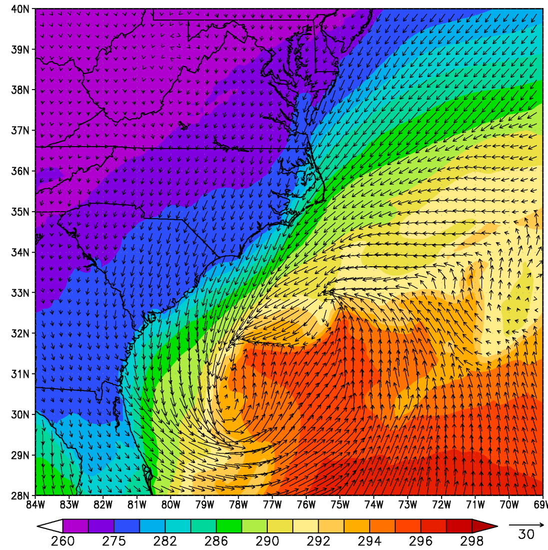

Figure 7. The control simulation's 10m wind vectors (m/s) and 2m temperatures valid 04Z 25 January 2000.

However, the convergence in the experimental simulation is more defined along a line following the GSF (Figure 8).

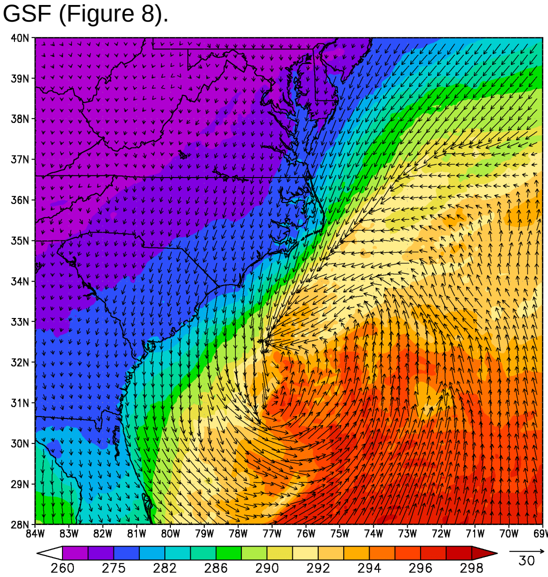

Figure 8. The experimental simulation's 10m wind vectors (m/s) and 2m temperatures valid 04Z 25 January 2000.

This is the major difference in the simulations, as well as a sign that the high resolution SST data did affect the track by changing the strength and location of the frontal boundaries as shown in the 10m wind plots. Stronger near-surface vorticity is seen in the experimental relative to the control in the vicinity of the front extending NE from the low center. As the upper trough approached, enhanced effectiveness of vortex stretching can explain the tendency for a stronger storm in the experimental simulation.

7. CONCLUSIONS

The overall performance of the experimental simulation (with high resolution SST) was better than the control simulation in all aspects of simulating this major event. In the control simulation, the reduced development compared to the experimental simulation may be due to the improper SST representation. The control simulation's poor forecast in track position may be linked in part to the weakly defined GSF. In the experimental simulation, the surface low tracked farther west along the now more accurately represented GSF. This track is a result of the surface low pressure following a zone of preexisting vorticity along the coastal front. It is the vortex stretching associated with the convergence along this frontal boundary that enhances the cyclogenesis. This coastal front, which formed above the tight marine thermal gradient of the GSF, is not seen in the control simulation.

The extension of the ASCII dataset covered additional 11 years. This more than doubled the number of storms. This additional data provides similar position and slope of the linear regression fits verifying the previous threshold values defined in the PSBI. The additional grouping of storms within the dataset based on 500mb vorticity reduced the scatter and further isolated the contributions of surface forcing versus upper level forcing on extratropical cyclogenesis.

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