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BOUNDARY LAYER COUPLING BETWEEN AN EXPLOSIVE CYCLONE AND THE GULF STREAM DURING ERICA

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ABSTRACT

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1.INTRODUCTION

An explosive cyclone is defined as an extra-tropical low pressure depression that deepens at least 10 mb in 6 hours for a period of 6 hours or longer (Hadlock and Kreitzberg, 1988). Explosive cyclones are smaller than the synoptic storms normally associated with weather patterns, having an effective radius of 500-1000 km; moreover, they are relatively short lived, lasting only 3-4 days. The explosive nature of this development has peak wind speeds reaching as high as 40 m/s in just a few days. Explosive cyclones are an autumn/winter phenomena, usually found off the east coast of continents (Sanders and Gyakum, 1980), developing in the vicinity of western boundary extension currents; i.e., the Kuroshio and the Gulf Stream. Both of these western boundary extension currents are associated with well defined sea surface temperature fronts separating warm tropical water on the south from colder subpolar water on the north. In the North Atlantic, explosive cyclones form principally over the north side of the Gulf Stream Front. Shay (1989) and others have correlated the position and strength of the Gulf Stream to the frequency of explosive cyclone development; i.e., when the Gulf Stream was displaced poleward of its normal position east of Cape Hatteras, the frequency of explosive cyclones in a particular autumn-winter season was increased. This suggests that explosive cyclogenesis is related to the intensity of the sea surface temperature front associated with the Gulf Stream. Little is known about the response of the Gulf Stream to the passage of these explosive cyclones.

Of principal interest in the general study of explosive cyclone development is their relationship with and dependence upon the sea surface temperature front of the western boundary extension currents. As a first step in examining this question, we examine the response of the gradient structure of the Gulf Stream to the explosive cyclone development over it. The following basic questions are answered. Does the development of explosive cyclones over the Gulf Stream modify the intensity of the sea surface temperature front of the Gulf Stream? Does explosive cyclone development modify the subsurface baroclinic structure of the Gulf Stream? Does the sea surface temperature front associated with the Gulf Stream modify the planetary boundary layer of the explosive cyclone and, hence, the fluxes of momentum, heat, and moisture across the air-sea interface? Answering these questions allows us to test certain hypotheses about how the Gulf Stream and explosive cyclones might be coupled.

Earlier efforts to determine the oceanic response to explosive cyclone development was conducted during GALE, focusing upon the Gulf Stream south of Cape Hatteras, where it flowed over the continental slope separating the warm tropical Sargasso Sea water from the colder shelf/slope water. Bane and Osgood (1988) examined AXBT sections across the Gulf Stream in this region both before and after a cold air outbreak that led to incipient cyclone development over the Gulf Stream. They found the mixed layer to have deepened on both sides of the Gulf Stream Front, but less on the inshore side where depth was limited by a rather shallow shelf of approximately 50 m depth. The result of this mixed layer deepening, caused principally by evaporative cooling of the surface mixed layer, was an excessive decrease in temperature over the shelf compared to that in the Sargasso Sea, resulting in an intensification of the sea surface temperature gradient across the Gulf Stream.

Historically, explosive cyclones have been found to develop over the Gulf Stream Front well east of Cape Hatteras, between 70°-55°W, where the continental slope/shelf (see **Fig. 1**) lies well north of the Gulf Stream (Shay, 1989). Therefore, we chose to examine the influence of an explosive cyclone upon the sea surface temperature front over this range of longitudes, well east of the continental shelf. To accomplish this, and to answer the questions posed above, oceanographic and atmospheric data were analyzed from the Experiment on Rapidly Intensifying Cyclones over the Atlantic (ERICA). During ERICA, an explosive cyclone event occurred over the period December 13-14 1989, labeled IOP-2. This cyclone was particularly interesting because it developed in the absence of upper air baroclinicity that was favorable for storm development, suggesting that its explosive growth was somehow linked dynamically with the sea surface temperature front of the Gulf Stream below it. This explosive cyclone began as a low pressure disturbance over the Gulf Stream sea surface temperature front adjacent to Cape Hatteras (**Fig. 2**). It grew in place from 0600-1800 December 13, then propagated to the east along the

Gulf Stream Front over the subsequent 18 hour period, as shown in both **Fig. 1** and **Fig. 2**. This cyclone deepened into a very intense low pressure system of pressure 974 mb, reaching maximum intensity on 1200 December 14. Thereafter, it continued eastward and dissipated.

We examined a number of parameter fields during a thirty day period bracketing IOP-2; i.e., ERICA sea level pressure, NESDIS sea surface temperature, GEOSAT altimetric sea level and wind speed, and ECMWF air temperature, specific humidity, and surface wind velocity. These data were analyzed, finding the surface and surface gradient structure of the Gulf Stream to respond significantly to the explosive cyclogenesis during IOP-2. The magnitude of this response was unexpectedly large. This suggested that the Gulf Stream effecting a positive feedback with the explosive cyclone that intensified the exchange of vorticity between the two fluid media. Models were used to test this hypothesis. Using an atmospheric planetary boundary layer model, the sea surface temperature front of the Gulf Stream was found to intensify the pattern of wind stress curl associated with the explosive cyclone. Using a sub-inertial pumping model of the ocean, this intensified pattern of wind stress curl was found to increase the gradient structure of the Gulf Stream. Its influence upon the development of the explosive cyclone remains unknown.

2. EXPLOSIVE CYCLONE DEVELOPMENT DURING IOP-2

The explosive cyclone during IOP-2 began as a low pressure disturbance off the East Coast of the U.S. between 0600 and 1800 GMT 13 December 1988, located near 36°N, 72°W directly over the sea surface temperature front of the Gulf Stream east of Cape Hatteras. This incipient low pressure disturbance is displayed in the upper left panel of **Fig. 2**, seen to be relatively weak in the beginning, dwarfed by a well developed low pressure system almost directly to the south of it near 30°N. In subsequent maps, this latter low pressure system diminished in intensity, while the low pressure disturbance to the north of it grew explosively as it tracked the Gulf Stream Front to the east over the next 18 hours (see **Fig. 1**). The low pressure disturbance in this first panel of **Fig. 2** occurred in a very auspicious location; i.e., over the Gulf Stream with the continental shelf not more than 300 km to the west (see lower panel of **Fig. 1**). This was where Bane and Osgood (1989) discovered that incipient cyclogenesis in the atmosphere tends to increase Gulf Stream frontogenesis. We can speculate that this provides favorable conditions for cyclone growth, making this region of the Gulf Stream east of Cape Hatteras the birth place of explosive cyclones. This remains to be seen. Earlier, Shay (1989) tracked 32 explosive cyclones over the Gulf Stream for a number of different winter seasons, finding explosive growth to occur more frequently farther to the east between 70°W and 60°W, and confined to the north side of the Gulf Stream. Still, we find significance in the fact that this low pressure disturbance, in the first panel of **Fig. 2**, stalled during the incipient stage of cyclone growth in the region over the Gulf Stream adjacent to Cape Hatteras for the 12 hours period from 0600-1800 December 13.

Explosive cyclogenesis of this low pressure disturbance began on 1800 13 December and continued over the next 18 hours until 1200 14 December, diagramed over the next three panels in **Fig. 2**. This explosive growth was associated with propagation of the cyclone to the east along the path of the Gulf Stream Front (see upper panel of **Fig. 1**), with an average speed of 14 m/s. By 1200 December 14, the central pressure reached a minimum of 974 mb, with most of the deepening occurring between 0000 and 0600 14 December when the central pressure dropped from 998 mb to 982 mb. This maximum increase in cyclone intensity occurred when the explosive cyclone followed the Gulf Stream Front from 70°-60°W.

Within each of the panels of **Fig. 2**, a warm front in the lower atmosphere extended zonally and eastward from the center of the explosive cyclone approximately along the path of the Gulf Stream Front. Across this front, the southerly geostrophic wind south of the warm front veered sharply easterly north of the front. As we shall see, this is associated with dramatic changes in planetary boundary layer characteristics in the vicinity of this front.

3. CHANGE IN THE SEA SURFACE TEMPERATURE FRONT OF THE GULF STREAM DURING IOP-2

The National Environmental Data and Information Service (NESDIS) produced sea surface temperature maps of the Gulf Stream region every 4 days from AVHRR satellite observations, buoy observations, and ship observations, forming a composite picture of the sea surface temperature for the region (Kreitzberg, 1990). Because of the cloud cover occurring during explosive cyclogenesis, the sea surface temperature field could not be examined using satellite AVHRR alone; *in situ* measurements made an important contribution to this composite picture. The resulting maps allow the sea surface temperature gradients associated with the Gulf Stream both before and after explosive cyclone development to be examined.

Composite sea surface temperature maps for December 11 and December 18 are shown in the two upper panels of **Fig. 3**. The upper panel represents the sea surface temperature distribution before IOP-2 and the middle panel represents the sea surface temperature distribution after IOP-2; in the lower panel is the difference between the two maps above. With this in mind, examination of the two composite sea surface temperature maps in **Fig. 3** show the gradient structure associated with the Gulf Stream after IOP-2 to be much more intense than before. Prior to explosive cyclone development during IOP-2, the Gulf Stream from 72°-60°W was almost meander free, extending quasi-zonally between 37° and 39°N over these longitudes, associated with a moderately intense meridional gradients. After explosive cyclogenesis, the Gulf Stream was associated with much more intense sea surface temperature gradient structure over this longitude band, on average increasing approximately 25-50%, accompanied by an apparent southward displacement of approximately 0.5° latitude.

The overall change in sea surface temperature between these two time periods is given in the lower panel of **Fig. 3**. It shows a general decrease of less that 1°C over most of the northeast North Atlantic, accompanied with a large decrease of 4°-6°C along the north side of the Gulf Stream Front over most of the longitude range from 75°- 64°W. Not surprisingly, this is the longitude band where the explosive cyclone during IOP-2 had its largest decrease in pressure (see **Fig. 2**) as it propagated eastward along the Gulf Stream Front. Over the region 1-2° latitude north of the Gulf Stream, the change is sea surface temperature was mesoscale in nature, related to westward displacements in the numerous warm core rings located north of the Gulf Stream Front during this time period.

4. CHANGE IN THE SEA LEVEL TOPOGRAPHY OF THE GULF STREAM DURING IOP-2

Altimetric sea level data from the Exact Repeat Mission of Geosat has been obtained from the Jet Propulsion Laboratory (JPL) at the California Institute of Technology (Zlotnicki *et al.*, 1989). The Geosat Exact Repeat Mission (ERM) began in November 1986, with the following sampling characteristics: each track was sampled repeatedly every 17 days; adjacent tracks were separated by 1.475° longitude in the zonal direction; each track was separated from a adjacent track at most three days in time, with the eastern track being sampled later; samples of sea level along each track were separated by 7 km. A complete data description was furnished by Cheney *et al.* (1987).

In the present study, altimetric sea level data for approximately one year (June 1987 to June 1988) were employed. The first order of business was the application of the corrections to the raw data provided by JPL, correcting for tidal fluctuations, sea state, humidity, etc. The second order of business was the removal of the orbit error from each track (of order 3 m for the ERM), which if unremoved would have created an overwhelming artificial temporal variability. Since the orbit error has long wavelengths along track (predominantly one per orbit revolution, i.e., 40,000 km), the usual practice is to represent the orbit error along each track segment by a long wavelength function (e.g., a linear or quadratic trend, or a sine function), with magnitudes adjusted to minimize apparent temporal variability. This orbit error removal procedure has been discussed at length in Tai (1989, 1991).

In the present study, ascending and descending track information in the northern hemisphere has been used (i.e., nominal track length 10,000 km), with the sine-and-bias representation of the orbit error removed from each sample. Individual altimetric sea level readings along each track (7 km residuals made relative to the one year mean) are then interpolated onto a 0.5° latitude by 0.5° longitude at 17-day intervals. One serious drawback for the Geosat is the lack of an on-board radiometer to correct for humidity effects. Thus, a substantial part of the variability could be due to variations in humidity. The along-track high-pass filtering, and the smoothing provided by the spatial and temporal binning, relieves this problem on the small-scale side, while the orbit error removal relieve the problem on the large-scale side.

Maps of sea level residual for the two 17-day periods that approximately bracket IOP-2 are presented in **Fig. 4**, together with the distribution of the tracks used to construct these maps. The latter are important because many of the GEOSAT tracks of the Exact Repeat Mission were missing for this northwest North Atlantic region during this one year period. The map for the period before IOP-2 (i.e., December 3) shows mesoscale residuals in the longitude range from 35°-40°N, 75°-60°W being much smaller in magnitude than those for the period after IOP-2 (i.e., December 20). Prior to IOP-2, eddy activity in the near field of the Gulf Stream (i.e., from 73°-60°W) appeared qualitatively similar to eddy activity in the far field. After IOP-2, eddy activity in the near field of the Gulf Stream over this longitude range was enhanced dramatically compared both to that in the far field and to that prior to IOP-2.

These sea level observations can be made more understandable by referencing the residual maps to a climatology of relative dynamic topography for the region. This allows us to see how these changes influenced the dynamic height structure of the Gulf Stream. For this purpose, we have chosen the GDEM dynamic height climatology, obtained from Teague *et al.* (1990), referenced to 1000 db. The GDEM climatology exhibits a 40-50 cm change in sea level across the Gulf Stream extending over approximately 1.5° of latitude. This can be compared to the synthetic climatology used by Kelly and Gille (1990), which changed 90-100 cm across the Gulf Stream over approximately the same distance. This latter difference in sea level compares well with that determined by Levitus (1986) based upon the dynamic height computed relative to 2000 db. The reason we have not used the Levitus (1986) climatology is that the gradient across the Gulf Stream is much less than for either Kelly and Gille (1990) and Teague (1990), due to extensive smoothing. Therefore, the GDEM climatology chosen for use in this study, while underestimating the intensity of the surface currents of the mean Gulf Stream, represents the correct width of the Gulf Stream.

The referenced surface sea level topographies for December 3 and December 20 are given in **Fig. 5**. Upon inspection of both of these maps, we can observe that during the period after IOP-2 (i.e., December 20, lower panel of **Fig. 5**), the gradient is sea level across the Gulf Stream (proportional to the surface current) was much more intense that prior to IOP-2 (i.e., December 3, upper panel of **Fig. 5**). This was particularly true over the longitude range 70°-60°W, not incoincidentally where explosive cyclogenesis had been most intense during the intervening period. This increase in strength of the eastward surface current, by as much as a factor of two in some locations, was due to the very intense negative changes in sea level that occurred on the north side of the Gulf Stream associated with the passage of this explosive cyclone.

This is seen even more clearly when the differences between the two periods are plotted in the upper panel of **Fig. 6**, together with the 100 cm and 150 cm isopleth that define the approximate boundaries of the Gulf Stream in the upper panel of **Fig 5**. Dominant negative changes in sea level can be seen to have occurred on the north side of the Gulf Stream, associated with a series of intense negative eddies separated by regions of much smaller positive values. The net change between the 100 and 150 cm isopleths ove. the band of longitude from 70°-55°W was negative, seen clearly in the average change displayed in the lower panel of **Fig. 6,** where eddy activity was suppressed by low-pass spatial filters. The peak change on the north side of the Gulf Stream was 5-10 cm between 70°-60°, accompanied by smaller positive changes on the south side. The net effect of these average changes was to increase the cross stream slope in sea level by approximately 20-30%, taking into account the fact that that the GDEM climatology underestimates the mean current magnitude by a factor of 2.

This large scale 5-10 cm negative change in sea level on the north side of the Gulf Stream (lower panel of **Fig. 6**) was obscured in the synoptic sense by mesoscale changes of 40-60 cm change occurring over this 17-day time period (upper panel of **Fig. 6**). These intense negative changes, occurring in eddy-like fashion, were separated by positive changes of much less magnitude, accounting for the average negative change over the entire region north of the Gulf Stream from 60°-70°W. These intense negative changes, however, were associated with local increases in the velocity of the Gulf Stream by nearly a factor of two over what was seen prior to IOP-2. This indicates that any wind driven changes that occurred in sea level over this 17-day time period were complicated by fairly rapid mesoscale eddy readjustments that too may have been influenced by the explosive cyclone development.

Comparing these changes in sea level with those in SST shown in the lower panel of **Fig. 3** finds qualitative agreement, with both fields of change associated with an intensification of both the SST and sea level gradients associated with the Gulf Stream. If we allow the maximum changes in SST of approximately 5°C to represented those in average temperature over a mixed layer depth of 100 m, then the corresponding decrease in sea level would have been approximately 10 cm (i.e., using a coefficient of thermal expansion of 2X10-4 °C-1 at sea level with salinity taken as 35 parts per thousand) simply due to the contraction of the volume of sea water in response to the reduction in temperature. Therefore, at 64°W and at 70°W, where SST differences achieved this magnitude on the north side of the Gulf Stream SST Front, associated sea level changes of order 10 cm would be expected; indeed, at these locations large changes in sea level occurred on this order and higher (see **Fig. 6**). These results taken together suggest that the intensification of the surface current of the Gulf Stream occurred in response to a southward displacement of the baroclinic structure on the north side of the Front. Such an increase in the surface momentum of the Gulf Stream from 70°-60°W suggests a mixed baroclinic-barotropic response to wind and thermo-haline forcing associated with the explosive cyclone that tracked the Gulf Stream during IOP-2.

5. BOUNDARY LAYER COUPLING BETWEEN OCEAN AND ATMOSPHERE DURING IOP-2

The foregoing 25% increase in the average speed of the Gulf Stream, extending over approximately 10° of longitude, over a 17-day period was unprecedented in light of our present understanding of the response of the baroclinic structure to transient wind stress forcing (e.g., Veronis and Stommel, 1956). In fact, such intense changes in the baroclinic structure of the Gulf Stream on a weekly time scale are normally attributed to the propagation and growth of meanders and vortices, not storm forcing. However, in the lower panel of **Fig. 6**, the zonal scale of the observed change was too large to be explained by these mesoscale processes. Therefore, a hypothesis can be formulated where these intense large-scale changes in sea level are considered to have been due to the wind stress forcing by the explosive cyclone occurring during IOP-2, but strongly intensified by the presence of the Gulf Stream Front.

Evidence suggesting this hypothesis exists already. During the Frontal Air-Sea Interaction Experiment (FASINEX), a moderate gradient of 2° C in about 5 km was shown to be sufficient to cause large horizontal changes in momentum and buoyancy fluxes across the air-sea interface. In particular, when warm air flows over cold water across a sharp discontinuity a reversal in the buoyancy flux occurs and a stable internal boundary layer (IBL) forms. This layer occurs between the surface and the top of the planetary boundary layer, and may be several hundred meters deep, effectively restricting the vertical exchange of heat, moisture, and momentum between the surface and the upper part of the boundary layer. Observations and numerical simulations have shown as much as a 50% reduction in the surface stress as the air passes from the warm to the cold water (Rogers 1989, Koracin and Rogers 1990). These results suggest that the interaction between mesoscale storms and the upper layers of the ocean may be substantially affected by strong gradients in sea surface temperature such as occur in association with the Gulf Stream Front.

Therefore, we hypothesize that the sea surface temperature front associated with the Gulf Stream encourages strong gradients in atmospheric stability and effective drag coefficient in the planetary boundary layer during explosive cyclogenesis. This is expected to strongly intensify the wind stress curl over the Gulf Stream during explosive cyclogenesis. This amounts to a form of boundary layer coupling between the ocean and atmosphere, where the Gulf Stream Front is able to influence strongly the wind stress curl that forces it. The boundary layer coupling hypothesis is tested in two phases. First, a recent version of the boundary layer model of the atmosphere (e.g., Smith, 1988) is operated during IOP-2, allowing the atmospheric stability, the variable drag coefficient, and the wind stress curl over the Gulf Stream to be computed. We then determine whether or not the influence of the Gulf Stream upon the planetary boundary layer has intensified the wind stress curl. Then, a near-inertial model of the ocean, driven by this version of intensified wind stress curl (and to a lesser degree by the time rate of change of the wind stress divergence) is operated, establishing whether the intensified wind stress curl can explain the intense large-scale sea level changes observed in the lower panel of **Fig. 6**.

We begin testing the boundary layer coupling hypothesis by examining the surface wind vectors and air-sea temperature differences **(Fig. 7)** on 1200 GMT December 14 1988, when explosive cyclogenesis reached maximum intensity. The wind stress fields and the air temperature field were obtained from the European Center for Mid-Range Weather Forecasting (ECMWF), both measured at 10 m; the sea surface temperature field was obtained from

NESDIS, repeated from the upper panel of **Fig. 3**. The distribution of wind vectors in the upper left panel of **Fig. 7** can be compared to the sea level pressure map in **Fig. 2** for the same time period; it shows low level convergence east of the cyclone center along the warm front which lay directly over the Gulf Stream Front. The distribution of the air-sea temperature differences in the upper right panel of **Fig. 7** has the air temperature everywhere colder than the sea surface temperature, with maximum negative differences occurring directly over the Gulf Stream Front. This negative maximum extends as a tongue from near Cape Hatteras (where the air temperature was greater than 10° C colder than the sea surface temperature) eastward to near 60°W. North of the Gulf Stream Front, these air-sea temperature differences achieved local negative maxima over the warm core rings found there in **Fig. 3** and in **Fig. 5**.

Information of the air-sea temperature differences and the wind speed allows both the Richardson number and the variable drag coefficient to be calculated using the planetary boundary layer model presented in the **Appendix.** This boundary layer model is based upon similarity theory developed by Paulson (1970) and Businger (1971), more recently updated by Smith (1988). It presents a standard procedure for conducting the iterative derivation of both the Richardson number and the variable drag coefficient. The distributions over the northeast North Atlantic of the Richardson number and the variable drag coefficient on 1200 GMT December 14 1988 computed from this model are displayed in the lower panels of **Fig. 7**. The Richardson number can be seen to have achieved a negative maximum over the Gulf Stream Front between 60° and 70°W, associated with maximum negative air-sea temperature differences and minimum wind speeds. This is where the planetary boundary layer is most unstable. At this location, the variable drag coefficient achieved minimum values (i.e., approximately 1.0X10-3), while on either sides of the Gulf Stream Front, near 36°N and 40°N, the variable drag coefficient achieved maximum values (i.e., approximately 2.5X10-3).

The distributions of wind stress curl are displayed for 1200 GMT December 14 1988 in **Fig. 8** for the same time period that distributions of the surface wind vector, air-sea temperature difference, atmospheric stability, and variable drag coefficient are displayed in **Fig. 7**. Distributions of the wind stress divergence are not shown, since their influence in the model was mostly less than that of the wind stress curl distributions. The distributions of wind stress curl are computed for a constant drag coefficient (upper panel of **Fig. 8**) of 1.5*10**-3 and for a variable drag coefficient (lower panel of **Fig. 8**) whose values are given in the lower right panel of **Fig. 7**. Upon comparison of these two estimated fields of wind stress curl, that computed with the variable drag coefficient can be seen to have had significantly larger magnitudes (i.e., 2-3 time larger) than that computed with a constant drag coefficient. These large wind stress curl values were due to gradients that existed in the variable drag coefficient in the vicinity of the Gulf Stream Front.

This is true throughout the entire IOP-2, indicating that over the development of this explosive cyclone the wind stress curl forcing by the storm on the upper ocean was significantly influenced by the presence of the Gulf Stream Front. This provides verification for the boundary layer coupling hypothesis. However, the boundary layer model used in the **Appendix** to derive the variable drag coefficient has itself not been verified during the explosive cyclogenesis observed during IOP-2. Direct observations of the variable drag coefficient necessary to verify the boundary layer model were not available. However, we can provide for a kind of indirect verification by demonstrating that only by invoking this kind of boundary layer model can the observed changes in sea level in the vicinity of the Gulf Stream (i.e., see Fig. 6) be explained. This is done next.

6. RESPONSE OF THE GULF STREAM TO BOUNDARY LAYER COUPLING WITH AN EXPLOSIVE CYCLONE

We continue to test the boundary layer coupling hypothesis by demonstrating that the wind stress curl forcing (i.e., that associated with the variable drag coefficient produced by the boundary layer model) by the explosive cyclone during IOP-2 altered the baroclinic structure of the Gulf Stream in a manner similar to that observed in the lower panel of **Fig. 6**. Moreover, we demonstrate that only by using wind stress values derived from variable drag coefficients, as opposed to using wind stress values derived from a constant drag coefficient, can the magnitude of these observed changes be simulated.

Since the entire action of the explosive cyclone upon the Gulf Stream during IOP-2 occurs over a period of less than four days, the wind driven model chosen for use in this demonstration depicts the near-inertial responses of the upper ocean (i.e., the upper layer of a two-layer system) to wind stress forcing (e.g., White, 1988). The vertical displacement of the main pycnocline (h) is given by the following expression;

$$d3h/dt3 + [(fo)**2]*dh/dt = -d[Div\ Tau]/dt - fo*Curl\ Tau \qquad (6.1)$$

where fo is the Coriolis parameter and Tau is the horizontal wind stress vector. In the integration of this near-inertial model, 12 hour time steps are used, with the wind stress computed using ECMWF 10 m winds and drag coefficients derived from the boundary layer model given in the **Appendix**.

The response of the model sea level to the wind stress forcing, is given in the upper panel of **Fig. 9**, resulting from the integration of (**6.1**) for 4 days from December 12 to December 15. This yields the response of the baroclinic structure of the Gulf Stream to the explosive cyclone that occurred during IOP-2. This response is given in terms of sea level, computed from the thermocline perturbation by multiplying the latter by the density contrast (i.e., .005) between the upper and lower layer of the two-layer model ocean. The maximum change can be seen to be approximately 10 cm, confined principally to the north side of the Gulf Stream, centered near 39.5°N, 61.5°W, similar to that observed in the lower panel of **Fig. 6**.

Yet, this wind driven model is not able to yield the mesoscale changes observed in the upper panel of **Fig. 6**. From an inspection of the evolution in the distribution of eddies observed in **Fig. 5**, we hypothesized that this evolution could be modeled by a linear baroclinic Rossby wave model (e.g., White, 1977). In this model, the baroclinic eddy structure in the upper panel of **Fig. 5** propagates toward the west in non-dispersive fashion at a speed given by -beta*gprime*H/f**2, approximately equal to 7 cm/sec for a gprime of 5 cm/sec**2 and H equal to 700 m, yielding the distribution of eddies seen in the lower panel of **Fig. 5**. Integrating this model from December 3 through December 20 yields the sea level change given in the middle panel of **Fig. 9**. These changes are mesoscale in character with maximum changes of 30-60 cm. Their pattern is similar to those observed in the upper panel of **Fig. 6**, yielding a pattern correlation of 0.47, but their magnitudes are smaller by about a factor of two. Part of this difference in magnitude is due to the influence of wind forcing given in the upper panel of **Fig. 9**.

The total response to the wind driven near-inertial model and the linear baroclinic Rossby wave model is displayed in the lower panel of **Fig. 9**. Clearly, the total response is dominated by the mesoscale response from the Rossby wave model in the middle panel of **Fig. 9**, with the large scale wind driven response tending to reduce the intensity of the positive mesoscale perturbations and increase the intensity of the negative mesoscale perturbations. This is similar to what was observed in the observed change in sea level given in **Fig. 6**, but with less effectiveness in the model.

A comparison between the observed and model sea level changes over the period from December 3 through December 20 1988 is more easily conducted in **Fig. 10**, where both are displayed together. On the left hand side of **Fig. 10** are the unsmoothed observed sea level changes compared to the total model response; on the right hand side of **Fig. 10** are the smoothed observed sea level changes, compared to the wind-driven model response. The percent of observed variance explained by the total model response is calculated to be 15% over the region 36°-42°N, 55°-70°W, while the percent of smoothed observed variance explained by the wind driven model response is calculated to be 30%. In the latter comparison, the magnitude of approximately 10 cm in the observed field is matched by 10 cm in the model response. Therefore, the wind-driven model is able to explain not only the phase of the observed response of the Gulf Stream to explosive cyclones but also the magnitude. Had we displayed the model response to wind stress values based upon a constant drag coefficient, the magnitude of the response would have been approximately 1/4 of that shown in the lower right panel of **Fig. 10**. This would have been inadequate to explain the smoothed observed changes in sea level seen in the upper right panel of **Fig. 10**.

The wind driven model change in sea level calculated over the four day period of IOP-2 seems to have accounted for the entire smoothed observed change over the 17 day period from December 3- December 20. To verify this, in **Fig. 11** is plotted the time sequence of the sea level at 39°N, 61°W for each 12 hour period between December 3 and December 20 1988. This time sequence shows that most of the change that occurred over the 17 day period occurred over a 24-hour period during IOP-2. This period encompassed 1200 GMT on December 14 where the explosive cyclone during IOP-2 was observed to achieve maximum intensity at this location (see **Fig. 7** and **Fig. 8**).

7. CONCLUSIONS

Attendees of the workshop on Mesoscale Coupled Air-Sea Interaction (Goroch, 1990) raised the issue that explosive cyclone development over the Gulf Stream, like that observed during ERICA, may be influenced by coupled air-sea interaction. As a first step in establishing the likelihood of this prospect, we demonstrated here that the baroclinic gradient structure of the Gulf Stream responded significantly to explosive cyclone development during IOP-2 of ERICA. This response appeared to result from the action of the wind stress curl associated with the explosive cyclone. However, the magnitude of this wind stress curl forcing was found inadequate to explain the magnitude of the oceanic response. Therefore, as the second step is establishing the existence of coupled air-sea interaction, we hypothesized the concept of boundary layer coupling between the explosive cyclone and the Gulf Stream Front. To demonstrate this, we employed a boundary layer model of the lower atmosphere (see **Appendix**) wherein the Gulf Stream Front itself was able to influence the atmospheric stability (i.e., Richardson number) in the planetary boundary layer, causing the drag coefficient to vary significantly across the front and dramatically increasing the wind stress curl attributed to the cyclone. This completing a positive feedback between ocean and atmosphere that has the Gulf Stream influencing the wind forcing that drives it. It remains to establish the influence that this coupled air-sea interaction has upon the development of the explosive cyclone itself.

The idea that coupled air-sea interaction was operating in this encounter between the Gulf Stream and the explosive cyclone stemmed initially from the large magnitude of the oceanic response over a 2-3 day period. The sea surface temperature front of the Gulf Stream between 72°-60°W increased by approximately 25% and the latitude location was displaced southward by .025-0.50°latitude (see **Fig. 3**). This was associated with a 4°-6° C decrease in sea surface temperature along discrete sections on the north side of the front. This was accompanied with an even more significant change in the internal baroclinic structure of the Gulf Stream, manifested in sea level (see **Fig. 5** and **Fig. 6**) where an average decrease in sea level of approximately 10 cm occurred between 70° and 60°N on the north side of the Gulf Stream. The overall effect of this was to increase the average eastward surface current of the Gulf Stream by approximately 25% over a two-week (or less) period.

The impact of this work on the theory of the Gulf Stream is of some interest. The Gulf Stream itself seems to have been able to modify the wind stress curl associated with explosive cyclogenesis in such a way as to promote large local responses in its thermo-dynamical structure over very short periods of time (i.e., on the order of a few days to weeks). Moreover, the character of the response of the Gulf Stream to the exploding cyclone during IOP-2 was such that it intensified both the sea surface temperature front and the eastward flow of the surface current by about 25% over this very short period of time. This undoubtedly established favorable initial conditions for additional explosive cyclones to develop. The next step in testing this hypothesis is to determine if boundary layer coupling occurred over other Intensive Observing Periods during ERICA. If boundary layer coupling is found to have been ubiquitous during explosive cyclogenesis, typically an autumn/winter phenomenon, we may have to alter our ideas about the mechanisms that initiate seasonal variability in the Gulf Stream; i.e., allowing local feedback with the atmospheric cyclogenesis to instigate significant seasonal variability.

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Appendix

Calculation of Surface Momentum, Heat and Moisture Fluxes in the Oceanic Boundary Layer

The surface wind stress, , is conventionally related to the surface wind speed by:

(A1)

where is the density of air, U is the mean wind speed at a selected height (e.g., 10 m) and is the drag coefficient dependent on wind speed, atmospheric stability, and various surface conditions. The surface heat flux, H, is conventionally related to the air-sea temperature difference by:

(A2)

where is the heat flux coefficient, is the specific heat of air at constant pressure, and is the potential temperature at a selected height (e.g., 10 m). The potential temperature is given by

(A3)

where

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