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The Near-Inertial Response to Hurricane Ivan

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UNIVERSITY OF MIAMI

A thesis submitted in partial fulfillment of
the requirements for the degree of
Master of Science

THE NEAR-INERTIAL RESPONSE TO HURRICANE IVAN

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The Near-Inertial Response to Hurricane Ivan (December 2013)

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As part of a US Naval Research Laboratory Experiment (Teague et al., 2007), fourteen Acoustic Doppler Current Profilers (ADCPs) were deployed on the northern rim of the DeSoto Canyon in 2004, as well as four Aanderaa current meters placed at approximately 950 m depth. Preceding placement of this moored array, dozens of conductivity, temperature, and depth probes (CTDs) were ship-cast in the surrounding area. At 0000 UTC on September 16, 2004, hurricane Ivan passed directly over the mooring array, leaving all fourteen moorings intact. Unprecedented measurements of currents in three dimensions and bottom temperature and pressure were recorded during storm passage. Using the HYbrid Coordinate Ocean Model (HYCOM) and the Navy Coupled Ocean Data Assimilation (NCODA) system, background Loop Current features and geostrophically balanced eddies were included in the simulation of Ivan’s response at the fourteen ADCP moorings. With such a large span of data, spatial and temporal resolution of features is possible at the array site.

A comprehensive study is conducted of the near-inertial dynamics left behind in Ivan’s wake. Vertical and horizontal wave characteristics such as phase, wavelength, and frequency are determined for the observed and simulated currents. Near-inertial motions were separated into barotropic and baroclinic modes, and rotary kinetic energy of these components was examined. The importance of individual modes was determined through a least squares fit of modal amplitudes to the observed currents. Modal contribution to
shear at the mixed-layer base (MLB) was examined in light of rotary kinetic energy output at the same depth. Background relative vorticity was calculated to determine if it affects frequencies and length scales of near-inertial motions. Surface heights from altimetry and depth-averaged mass divergence calculated from HYCOM simulations were used to further understand the barotropic mode and its interactions with the depth-independent current and baroclinic modal response to Ivan.

The baroclinic near-inertial amplitude at 50 m is 70 to 80 cm s\(^{-1}\), and the depth-averaged (barotropic) amplitude is 6 to 8 cm s\(^{-1}\), consistent with previous studies. Oscillations similar to the ADCP-measured barotropic response are recorded by the Aanderaa current meters at a depth of 950 m. Altimetry and HYCOM output suggest a free surface depression of approximately 15 cm from Ivan and a sinusoidally oscillating depth-averaged mass divergence at near-inertial frequencies. A wavelength of 229 km was calculated for mode one, and carrier frequencies of 1.08\(f\) to 1.20\(f\) (where \(f\) is the local Coriolis parameter) were observed for surface-intensified flows induced by a hurricane. The presence of background relative vorticity may have shifted near-inertial frequencies. The calculation of clockwise (CW) rotary kinetic energy for HYCOM output and ADCP observations indicated that HYCOM slightly overestimated response amplitudes and kinetic energies. An analysis of vertical shear indicated that mode two is responsible for mixing in the upper thermocline according to ADCP data (which does not include the mixed-layer (OML)). By contrast, an examination of HYCOM simulations (which include the OML) indicated that mode three controls shear at the MLB, and hence lowers the Richardson number below the critical value required for vertical mixing.
Acknowledgments

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<th>Description</th>
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<tr>
<td>ACW</td>
<td>Anticlockwise</td>
</tr>
<tr>
<td>ADCP</td>
<td>Acoustic Doppler Current Profiler</td>
</tr>
<tr>
<td>CCE</td>
<td>Cold Core Eddy</td>
</tr>
<tr>
<td>CTD</td>
<td>Conductivity Temperature and Depth probe</td>
</tr>
<tr>
<td>CW</td>
<td>Clockwise</td>
</tr>
<tr>
<td>GOM</td>
<td>Gulf of Mexico</td>
</tr>
<tr>
<td>HYCOM</td>
<td>HYbrid Coordinate Ocean Model</td>
</tr>
<tr>
<td>IP</td>
<td>Inertial Period</td>
</tr>
<tr>
<td>MLB</td>
<td>Mixed-Layer Base</td>
</tr>
<tr>
<td>MS</td>
<td>Mooring</td>
</tr>
<tr>
<td>NHC</td>
<td>National Hurricane Center</td>
</tr>
<tr>
<td>OML</td>
<td>Oceanic Mixed-Layer</td>
</tr>
<tr>
<td>SHA</td>
<td>Surface Height Anomaly</td>
</tr>
<tr>
<td>SST</td>
<td>Sea Surface Temperatures</td>
</tr>
<tr>
<td>WCE</td>
<td>Warm Core Eddy</td>
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Chapter 1: Introduction and Background

When a strong tropical cyclone passes over the ocean, its anticlockwise (ACW) rotating winds impart their momentum throughout the oceanic mixed-layer (OML) and create diverging currents below the storm center. In the northern hemisphere, winds induce oceanic mass transport to the right of their motion, causing upwelling, displacing isopycnals (surfaces of constant density) throughout the water column, and creating horizontal pressure gradients. As the storm passes, the upwelling is followed by downwelling approximately one-half inertial period (IP; defined here as the inverse Coriolis parameter \(f\)) later. In response to the changing mass field, the momentum field rotates CW to restore balance and redistribute excess wind-driven kinetic energy in the upper ocean. Thus, the response is a three-dimensional geostrophic adjustment problem with near-inertial waves radiating away from the storm (discussed by Rossby (1938)).

Storm translation speed determines the extent to which the oceanic response is geostrophic or near-inertial. Slow moving storms, for example, leave behind fairly intense upwelling that forces large horizontal pressure gradients and generates geostrophically balanced flows. By contrast, fast moving storms leave behind a wake of strong near-inertial response that includes CW-rotating currents having both a strong depth-dependent and weaker depth-independent responses.

Although diverging currents in the OML cause upwelling throughout the water column, the largest isopycnal displacement occurs in the upper thermocline. Pressure-coupling between the two layers leads to excitation of strong, oscillating currents at depth at the near-inertial frequency. OML near-inertial currents lose their energy by spreading laterally and through coupling with the thermocline to inject kinetic energy as part of the
three-dimensional response. A large fraction of near-inertial kinetic energy from the OML propagates vertically into the thermocline before it can be lost, with the horizontal loss occurring on a longer time scale than the vertical exchange (Price, 1983). These near-inertial oscillations possess wave dispersion characteristics such as phase and group speeds and wave characteristics such as amplitudes, phases, and horizontal and vertical wavelengths, that aid in the reestablishment of geostrophic equilibrium after storm passage.

Mixed layer pressure coupling with the thermocline causes downward energy propagation (upward phase propagation); however, near-inertial motions can also reflect off the ocean bottom and lead to upward energy propagation (downward phase propagation). The direction of energy propagation is found by splitting the near-inertial response into CW- and ACW-rotating components. Kinetic energy of those components is compared to determine which processes (upward or downward propagation) dominate at different depths. Downward energy propagation (associated with CW rotation) dominates by a factor of four to one (Shay and Jacob, 2006) most often at the base of the OML and in the upper thermocline, where large shear values cause vertical mixing. Isopycnal displacement and the establishment of horizontal pressure gradients are often associated with net horizontal transports away from the storm center induced by the surface wind field. However, upward energy propagation (ACW rotation) indicates interactions with steep bathymetry and reflection of wave-like features (Leaman, 1976).

As noted, the propagating OML currents must be near-inertial (the period is shifted slightly off of the time scale associated with the local Coriolis parameter, $f$). One IP is approximately one day at the latitude in which this study takes place ($\sim 29.2^\circ$N), although
its latitudinal dependence means that it can span from 12 h at the poles to infinity at the equator. Previous studies partitioned near-inertial oscillations into several components, or vertical modes, which sum together to represent the dominant response. Baroclinic modes make up the spectrum of near-inertial current oscillations, but it is the low-order modes (typically the first three to five) that generally govern the momentum response. Higher order modes have smaller amplitudes but can become dominant over a sufficiently steep bottom slope (Vlasenko et al., 2005) or at the base of the OML where vertical shear is strong (Shay et al., 1989). Each mode has its own wavelength, amplitude, and phase. Mathematically, parameterizing the current flow as a sum of baroclinic modes allows for a linear analysis and typically represents 65% to 80% of the observed variance (Shay and Elsberry, 1987; Shay et al., 1989).

The possibility that higher-order modes are dominant at the MLB has important implications. Strong storm-induced currents in the OML are separated from weak thermocline currents by strong density stratification, causing vertical shear to develop. Shear at the interface between OML and thermocline controls mixing, changes in OML temperature and depth, and oceanic heat content, all of which are important oceanographic parameters. Shay et al. (1989) found that higher order modes (three and four) may control a large fraction of the vertical shear at the MLB through the opposing velocities of their zero crossings, expending near-inertial energy to do so. High-order modes would be expected to control shear because they occur over scales similar to that of the shear layer. Other features, such as background vorticity, affect shear by trapping near-inertial oscillations in, or expelling them from, the OML.
In this context, background conditions must be considered when characterizing a storm-induced ocean response. Propagating eddies in the Gulf of Mexico have a scale of approximately 100 to 200 km, rotate either CW or ACW, and modulate near-inertial responses. ACW-rotating eddies spin off from the Gulf of Mexico Loop Current and CW-rotating eddies form nearby due to instabilities. Eddies change the background vorticity and can increase shear and mixing or lead to a reduced response. Increased shear and mixing is caused by ACW rotating eddies, which increase stratification in the thermocline and create an MLB that is harder to overcome with vertical shear, confining strong wind-driven currents to the OML. Conversely, warm CW-rotating eddies have weaker stratification and facilitate downward energy propagation, creating favorable oceanic conditions for vertical movement of near-inertial motions from the OML into the ocean interior. Rotational effects of eddies determine the amount of near-inertial energy that can exit the OML without being used to drive vertical shear and induce mixing, overcoming the OML-thermocline interface. Therefore, background vorticity from total sea surface heights is used to diagnose Loop Current features that may affect near-inertial motions, mixed layer depth, and shear magnitude (Jaimes and Shay, 2010).

Near-inertial energy can contribute to global ocean internal wave power and global meridional heat fluxes if it propagates downward to lower levels (Emanuel, 2001; Shay and Jacob, 2006; Jaimes and Shay, 2009), so energy expended on vertical shear and turbulent mixing counts as energy lost to the layers below. If background vorticity features such as cold core eddies (CCEs) increase stratification, near-inertial oscillations must expend more energy to overcome it and reach lower layers, leaving less energy available on a global scale. When background vorticity features such as warm core eddies
(WCEs) have less stratification at the MLB, near-inertial oscillations penetrate into the thermocline easily, giving more energy to the layers below. While the extent of energy contribution is uncertain (~$10^{11}$ to ~$10^{15}$ W, depending on the cited studies), background features are important for the very reason that they impact near-inertial oscillations on a local scale but have global-scale consequences. Furthermore, if modes three and four are the main contributors to vertical shear at the MLB, then they are responsible for energy loss in that area. Energy loss from these two modes can be quantified to provide insight into the amount of near-inertial energy that hurricanes contribute to the global oceans.

Shear and background vorticity are features which can affect baroclinic near-inertial motions. Prior to the excitation of a strong baroclinic response, however, the storm induces a barotropic response (mode zero) throughout the entire water depth. Although the barotropic response is constant from the ocean surface to the bottom and weaker than the baroclinic oscillations, its net transports can be significant. Because its wave speed is higher, it is thought that the barotropic response propagates away immediately after storm passage, leaving the baroclinic currents to redistribute thermocline kinetic energy. Previous studies suggest that the barotropic response is a result of free surface changes in the storm wake and that it penetrates to the ocean bottom in the form of a depth averaged current (9 to 11 cm s$^{-1}$; Shay et al., 1990).

As noted above, mixed-layer water is forced to the right of the ACW-rotating wind stress, initiating a process that will lead to the excitation of baroclinic and barotropic near-inertial motions. Storm-induced mass divergence causes an elongated free-surface depression of approximately 15 to 20 cm in the along-track direction known as the barotropic trough where geostrophically balanced currents rotate ACW around a cold
wake (Geisler, 1970). In addition, for a moderate to fast moving storm, surface oscillations of 3 to 7 cm occur in the trough along the track at approximately near-inertial periods. Most of the near-inertial response to these undulations consists of CW-rotating barotropic motions, and in fast-moving storms baroclinic near-inertial currents make up a larger fraction of the total response (Geisler, 1970; Shay and Elsberry, 1987; Shay et al., 1989).

The near-inertial waves discussed above are found all over the globe and are associated with many processes; however, tropical cyclones create momentum responses that are intense (on the order of 80 to 120 cm s\(^{-1}\) in the OML and 40 to 60 cm s\(^{-1}\) in the upper thermocline) and confined with wavelengths that scale with storm translation speed and the local inertial period (Shay and Elsberry, 1987; Shay et al., 1989; Shay and Chang, 1997). Strong pressure gradients in the storm wake are the result of intense upwelling and changing sea surface height and lead to a relatively weak barotropic response at depth, followed by a strong baroclinic response. Understanding how near-inertial waves radiate and redistribute storm energy (on what temporal and spatial scales, using which physical mechanisms) is pivotal to understanding how storms affect the OML depth and heat content and forecasting strong currents and upwelling.

1.1 Hurricane Ivan

An exceptionally strong hurricane, Ivan reached category 5 status three times along its 5600 n mi long track and lasted 22.5 days (Figure 1.1; from Mainelli et al. 2008). On September 2, 2004, Ivan began as a tropical depression off the west coast of Africa and subsequently reached hurricane status on September 5, 2004 1000 n mi east of Tobago. During the next ten days, Ivan reached the Caribbean, caused considerable damage in
Grenada, Jamaica, and Grand Cayman, and entered the Gulf of Mexico. Passing over the mooring site on September 16, 2004, Ivan made landfall near Gulf Shores Alabama, where it caused many flash floods and tornadoes. Subsequently, Ivan lost its tropical status, moved to the Atlantic ocean, and circled back over south Florida to re-enter the Gulf of Mexico (GOM) on September 21, 2004. Over the GOM, Ivan became a tropical storm once more and made landfall over Louisiana on September 24, 2004 (Stewart, 2004).

Figure 1.1 Pre-storm OHC, storm track, and intensity for hurricane Ivan (from Mainelli et al., 2008)

According to Teague et al. (2007):

Ivan had been the most expensive hurricane ever for the oil and gas industry in the gulf, prior to Hurricane Katrina in 2005. The Minerals Management Service
(MMS) reported that Ivan forced evacuation of 75% of the staffed platforms in the gulf (574 platforms) and 59% of the drilling rigs (69 rigs), set adrift 5 rigs, and sunk 7 rigs entirely.... Aside from obvious leaks, some pipelines were reported to have moved about 915 m while others were buried under 9 m of mud and could not be found. The most extensive damage to the pipelines was attributed to undersea mudslides (equivalent to a snow avalanche) caused by extreme surface waves and hurricane wind-driven currents.

1.2 Motivation

Strong storms cause damage to offshore structures and impact ecosystems, including coral reefs, in the GOM. They also change ocean heat content (OHC) by deepening and cooling the mixed layer. Since near-inertial oscillations distribute energy and reestablish equilibrium after a storm, understanding these energetic motions is the first step to mitigating damage and improving predictions of OHC. Effects of topography on near-inertial wave propagation are not well understood, so the ADCP measurements’ continuous resolution in time and space and their location over the DeSoto Canyon is ideal. Understanding this process throughout the water column is central to characterizing wind-forced currents and their possible impact on thermal structure for weather and climate studies. To obtain that knowledge, the mechanisms of energy redistribution and their interactions with topography must be studied, hence the desire to examine details of near-inertial wave excitation throughout the water column.

In addition, oil spills in the GOM have been a major problem in the past and will continue to be a liability for the Gulf ecosystem (e.g., Deepwater Horizon (DwH)). Strong tropical cyclones inflict damage to offshore and onshore coastal structures through
severe weather at landfall where costs to rebuild are in the tens of billions of dollars. Storm surges are one of the main causes of death and property damage along the coast. The devastation has a major impact on the economy of coastal communities that can extend for several years depending on storm strength at landfall and the ensuing severe weather (Marks and Shay, 1998).

Along the northern GOM, there are more than 4000 oil platforms and structures located in coastal areas, on the shelf break, and, in the case of the DwH site, in depths of up to one mile. Given the large number of wells, oil spills will have a major impact on the Gulf ecosystem from which coastal communities drive their economies through fishing, marine transportation and shipping, and recreation. In the case of the DwH, this was a subsurface spill in approximately 5,200 feet of water. A thorough understanding of key oceanographic circulation processes was needed to track the potential pathways of the oil over 87 days. Real time oceanographic data was required by predictive models (Shay et al., 2011).

Oil was found on the surface and in deep water plumes at various depths in the DeSoto Canyon and Sigbees Escarpment located due south of New Orleans. During this environmentally sensitive period, the northern GOM did not have to contend with a major hurricane, which may have amplified effects of the spill. Hurricane-induced rightward deflected currents (e.g. storm surge) would likely have forced more oil shoreward.

The communities along the northern GOM were not only exposed to oil, but also dispersants used to help weather oil over a prolonged period. Hurricane-induced upwelling may bring submerged weathered oil to the surface and drive it into coastal areas months or years after it originally spilled. This situation may be exacerbated if a
storm slows down, causing upwelling from depth as recently observed with hurricane Isaac. In this broader context, accurate prediction of barotropic and baroclinic responses to hurricane forcing is necessary to diagnose possible reappearances of weathered oil, as well as possible impacts of multiple storms on oil rigs and underwater pipelines. The presence of a large canyon complicates the dynamical response that may lead to enhanced upwelling processes along steep canyon walls.

1.3 Goals and Objectives

Wave characteristics of the near-inertial response to Ivan over varying bathymetry in the GOM will be examined and characterized using simulations from a model sensitivity study where several attributes of the model were changed relative to a control numerical experiment (Halliwell et al., 2011). Output from the control experiment is used here. In particular, the evolution of carrier frequency, group and phase speed, amplitude, and wavenumber will be investigated, along with implications for vertical kinetic energy transfer from hurricane winds to the thermocline and below. The presence of standing modes will be assumed as well considering the post-storm response is likely a mixture of standing modes and propagating waves (Blumenthal and Briscoe, 1995). Baroclinic oscillations seek to redistribute kinetic energy, although they can be further complicated by background geostrophic flow (not to be confused with the geostrophic response to hurricane upwelling). Eddies, or mesoscale geostrophic currents, are part of Loop Current (LC) dynamics in the GOM may either be CCEs (ACW flow around a low pressure, raised isopycnal surfaces) or WCEs (CW flow around a high pressure, lowered isopycnal surfaces). Baroclinic near-inertial motions rotate CW (in the Northern Hemisphere) around an upwelled center, so CCEs can modulate their frequencies and shift them
further above $f$. WCEs act in the opposite sense and may dampen near-inertial oscillations, shifting their frequencies below $f$ (so the new frequency is $f + \zeta$; Jaimes and Shay, 2010). Here, the ADCP data fields will be compared to the HYCOM simulations to evaluate the model’s representation of baroclinic and barotropic responses to the storm and the success with which it parameterizes background flow and its effects on near-inertial frequencies.

Past investigations, discussed below, documented the ocean response to strong wind forcing in the GOM, identifying the importance of baroclinic near-inertial waves in momentum redistribution and reestablishment of a post-storm equilibrium state. However, those studies utilized moorings with limited vertical sampling over several IP, such as Shay and Elsberry (1987) in which the array was made up of three to four instruments. Another issue was limited temporal sampling; Shay et al. (1989) used expendable profilers during hurricane Norbert and obtained comprehensive measurements vertically, but only during a single snapshot in time. The data set used for this investigation is fairly comprehensive in that it covers flat and sloped bathymetry, is measured continuously throughout the water column, and spans a time frame from before, during, and after the passage of Ivan on both sides of the storm track (Teague et al., 2007). Thus, these data present a unique opportunity to study wave parameters throughout the water column (up to 950 m) and their changes as time progresses. Thus, an improved characterization of wavelike and modal behaviors and the relationships between modes and energy redistribution is now possible.

The first research objective is to **isolate the possible effects of background flow on the near-inertial response by analyzing background relative vorticity.**
• HYCOM was initialized to include a CCE that was in the vicinity of the ADCP moorings during Ivan’s passage (Halliwell et al., 2011). Therefore, its ability to parameterize background conditions can be evaluated using the ADCP data, as well as its ability to model the impacts of geostrophically balanced eddies on forced near-inertial motions.

The second objective is to compare wave characteristics derived from the model simulations and ADCP data, as well as the energy balances associated with the barotropic and baroclinic modal responses in both products.

• Comparing the data sets would allow for a diagnosis of the data and model fields. The DeSoto Canyon is a particularly difficult area to model due to its changing slope, and freshwater outflow from the nearby Mississippi River complicates the density profile. Such a layout allows for an analysis of how well HYCOM represents variable topographic and stratification effects. More importantly, comparing model barotropic and baroclinic responses to those found in the data would provide a baseline for representing the vertical, horizontal, and temporal extent of those features’ wave dispersion characteristics, their amplitudes, and their kinetic energy.

The third objective involves using a rotary kinetic energy analysis to understand the importance of individual modes for specific processes, including shear at the MLB.

• Modes work to overcome strong stratification at the MLB and throughout the thermocline, making vertical energy propagation possible. Analyzing the magnitude, timing, and direction of energy propagation (which depends on the
CW or ACW rotation of near-inertial waves) may give clues as to the effects of background features and the role of individual modes. Hurricane energy must be redistributed by the near-inertial response via dissipation (shear), conversion to potential energy (structuring of stratification), or dissemination into the ocean depths. Any feature which affects energy redistribution is of interest due to the contribution of storm energy to the global near-inertial energy budget.

The fourth objective is to **characterize the barotropic response by documenting its spatial and temporal scales, as well as the impact of the trough left behind by the storm.**

- Strong, ACW-rotating hurricane winds create a trough in the free surface and affect pressure and density gradients at all depths. Depth-averaged mass divergence causes an oscillating barotropic response of ~7 cm s\(^{-1}\) (Shay and Chang, 1997) which reaches the ocean bottom. Sea surface heights from satellite observations and model fields, as well as temperature, salinity, and current measurements will give a more complete picture of the barotropic response at all depths.

### 1.4 Available Resources

The following instrumentation and model resources will be used to address the previously stated research objectives and resolve the near-inertial response, which is the goal of this study.
1.4.1 ADCP Data

The Naval Research Laboratory deployed fourteen Acoustic Doppler Current Profiler (ADCP) moorings (MS01 to MS14) on the northern rim of the DeSoto Canyon in the Gulf of Mexico (GOM) in May 2004 (Teague et al., 2007; Figure 1.2). Part of the Slope to Shelf Energetics and Exchange Dynamics (SEED) project, these moorings were to assist in measuring heat, water, mass, and momentum exchanges over steep shelf topography. Hurricane Ivan passed directly over the moorings on September 16, 2004 while they were still taking measurements. Each mooring survived the strong storm and provided unprecedented measurements of Ivan’s oceanic response impacts on the shelf break, over the shelf, and on the northern rim of the DeSoto Canyon.

Figure 1.2 Map of 14 ADCP moorings (black dots) located over the DeSoto Canyon in the Gulf of Mexico and Ivan’s track (black dashed line). Infrared satellite image of Ivan inset in the bottom right-hand corner. Taken from Teague et al. (2007).
1.4.2 HYCOM

Numerical experiments using the HYbrid Coordinate Ocean Model (HYCOM) were conducted for hurricane Ivan using the Navy’s global HYCOM analysis which is initialized by the Navy Coupled Ocean Data Assimilation (NCODA) system to create five-day hindcasts. Output data were taken at the 14 ADCP mooring sites before, during, and after Ivan’s passage for direct comparison. This particular experiment, called GOM1, used a K-Profile Parameterization (KPP) vertical mixing scheme, Donelan et al. (2004) drag coefficient for surface wind stress, and bulk formulae from the Coupled Ocean Atmosphere Research Experiment (COARE) 2.6 algorithm for surface turbulent fluxes. Model output utilized here includes current velocities in three dimensions, temperature, salinity, and surface height fields. Not only is the model data comparable to the ADCP data, but it provides an extension of measurements to the OML (above 50 m) and allows for estimations of mass divergence using the temperature and salinity fields (which were not observed during Ivan). The mooring instrumentation and HYCOM simulations will be discussed in more detail in Section 2.1.2.

1.5 Comprehensive Background of Past Studies

The problem of ocean readjustment to wind stress at the surface was examined by Rossby (1938). Three simple models: one homogeneous (constant stratification), one with two separate and constantly stratified layers, and one linearly stratified were compared to each other under the assumption that vorticity must be conserved in dynamic equilibrium after the application of wind stress to the surface. For all systems, a new mass balance was established in which currents shifted to the right of the forcing (in the northern hemisphere) to balance the momentum input, and near-inertial counter-currents
were responsible for propagating excess energy. Near-inertial oscillations in the homogeneous system were weak (approximately 3 cm s\(^{-1}\) for a 50 cm s\(^{-1}\) current) and broad, although in the two-layer system oscillations were strong (20 cm s\(^{-1}\)), narrow, and confined to the upper layer. In the two-layer system, up to 89\% of wind energy was transferred into near-inertial oscillations, as opposed to 7\% for the homogeneous system.

Effects of a moving positive wind stress curl and negative pressure anomaly were examined in a linear, two-layer ocean by Geisler (1970). In the wake of a moving storm, a free-surface deformation was associated with the barotropic mode and deteriorated into planetary waves within a day. If the storm translation speed was less than the long wave speed of baroclinic modes, no baroclinic near-inertial waves existed. However, if the storm translation speed was fast enough (compared to the first internal mode phase speed; Froude number greater than one) then for several weeks a geostrophically balanced baroclinic ridge existed at the interface between the two layers. Superimposed on this ridge were energetic near-inertial waves, the amplitude of which depended on the translation speed of the storm. For faster moving storms, the baroclinic response amplitude was large under the storm and along its track, indicating that non-linear theory is required to fully describe the response. Important to note is that the magnitude of the geostrophically balanced ridge depended on input of vorticity by wind stress curl, and not necessarily on storm translation speed.

The general theoretical context that frames the study of storm-induced near-inertial motions was developed by Price (1983) and Gill (1984), among others. Price (1983) modeled wind stress driven currents in the OML that diverged to initiate upwelling in the layers below. The energy budget was decomposed into OML and thermocline kinetic
energy (currents rotating around displaced isopycnals), potential energy (strength of stratification), and loss of kinetic energy (horizontally spreading wave wake). Observable in the model output, as well as a set of observed currents during hurricane Eloise (buoy EB-10 in the study), was that kinetic energy was injected into the thermocline through pressure coupling (“inertial pumping”) between the thermocline and OML. Energy was eventually dissipated as near-inertial waves propagated laterally out of the water column. Diminishing amplitude and changing phase in the model OML matched observations well, and EB-10 measured CW-rotating currents in the thermocline, suggesting that coupling did occur with the OML.

Gill (1984) decomposed a modeled near-inertial wave response into individual baroclinic modes. Analysis of the amplitudes and phases of each mode indicated that they oscillated in and out of phase with each other to redistribute energy. When modes were in phase, energy was high; when they were out of phase, they acted deconstructively to lower baroclinic energy levels. Time scales for modal separation from the solution were calculated and have been validated by observations. Typical time limits of separation based on phase speeds for individual baroclinic modes are ~2.5 d for mode one, ~5 d for mode two, and ~7 to 14 d for mode three (Shay and Elsberry, 1987; Shay et al., 1989). Gill noted the upwelling power of surface divergence (inertial pumping) in displacing isopycnals at great depths and initiating a thermocline and, several IP later, a bottom response. Energy density (the sum of kinetic and potential energy in an area) was also calculated and indicated downward energy propagation as the storm’s input was transported from the OML to the thermocline. The largest currents were observed directly under the storm track from one to three times the radius of maximum winds (R_{max}) and a
deeper OML allowed for a larger projection of the baroclinic modes onto the current response.

Observational studies have sought to evaluate the analytic and numerical theories of Gill and Price. Examination of hurricane Frederic oceanic response by Shay and Elsberry (1987) forms much of the methodological basis for this analysis of Ivan. The authors found the horizontal and vertical wave characteristics (carrier frequency, wavelength, wavenumber, and group speed) of the near-inertial response. Amplitudes and phases of the response elucidated its vertical, horizontal, and temporal extent. Next, the authors solved for the normal baroclinic modes and deduced their individual contributions to the response amplitude. As the theory predicted, observations showed a response dominated by the first three modes, strongest in the OML and weaker in the thermocline. These observations verified the temporal and spatial scales that numerical and theoretical studies predicted in the past. Effects of a sloping bottom, when included in the normal mode analysis, explained more of the variance associated with the near-inertial response; bottom topography was felt throughout the column as modes separated from each other. In addition to the baroclinic response, Shay and Elsberry (1987) found barotropic oscillations that were later confirmed through numerical and theoretical studies by Shay et al. (1990) and Shay and Chang (1997).

Shay et al. (1990) compared results from a linear model and a 17-layer non-linear model to verify the existence and characteristics of the barotropic mode. They found that the linear model was sufficient to describe the barotropic response, which had a maximum amplitude of 11 cm s\(^{-1}\) at 2 \(R_{\text{max}}\) and was forced primarily by wind stress curl. Free surface effects (depressions of 18 to 20 cm) created a depth averaged pressure
perturbation and excited a homogeneous response of near-inertial frequency throughout the water column. Superimposed on the barotropic mode were baroclinic modes in the OML and thermocline. Currents in these two layers were 180° out of phase, verifying the Price (1983) study, which found that pressure coupling transferred energy between the wind stress-forced OML and the thermocline. The models’ results agreed well with observations from hurricane Frederic, and they expressed the response decay rate in the OML and lateral spreading over the first few IPs with accuracy. In conclusion, the depth-averaged response measured during Frederic was not a result of limited vertical sampling, but a near-inertial barotropic current induced by the free surface and interaction with the baroclinic modes through the mean mass divergence field (Shay and Chang, 1997).

Shay et al., (1989) developed a solution for forced near-inertial modes in a continuously stratified fluid by extending the Geisler (1970) approach. Current velocity profile data acquired during hurricane Norbert were compared to simulated wind-forced currents in terms of normal modes based on a convolution between the forcing structure and a Green’s function (see Geisler, 1970 for further reference). For an observed Brunt–Väisälä profile (as opposed to a constant profile used in the Frederic study), they found that 70% of the variance was explained by the first four baroclinic modes. However, a key aspect of that study was that the velocity profiles represented a single snapshot in time as opposed to the Ivan data, which extends several IP after storm passage. A contour plot over depth and time of the v-component of the first ten baroclinic modes simulated in the model qualitatively resembles measured currents from the Ivan data set. Such a resemblance indicates that wind stress curl is one of the controlling factors of the current response to Ivan, just as it was in the Shay et al. study.
In a non-hurricane-induced case, Leaman and Sanford (1975) studied the direction of vertical energy transfer in near-inertial waves. Using cross-spectral analysis, they separated observed oscillations into CW and ACW rotary spectra and calculated the horizontal kinetic energy associated with each. Most energy was associated with the CW spectra, indicating downward energy propagation. Leaman (1976) extended the study to include a larger number of observations over smooth and rough topography. CW-rotating spectra still dominated the kinetic energy transfer over smooth topography, although some evidence of ACW-rotating waves was found over rough topography. A frequency spectrum of the data indicated that near-inertial waves were responsible for the energy transfer and allowed for a calculation of vertical energy flux. Unfortunately, not enough observations existed at the time to verify the ACW features. Leaman noted that the upper-thermocline shear zone appeared to be a source of near-inertial waves, although sparse measurements made separating the OML from the shear zone difficult. The discovery by Price (1983) via the Burger numbers for the OML and thermocline showed that the OML pressure couples with the thermocline to transfer energy to lower depths. It may explain the energy source in Leaman’s study, which noted a slight frequency shift off the local Coriolis parameter, enabling vertical energy propagation.

Jaimes and Shay (2010) used airborne expendable profilers, moorings, and altimetry measurements to examine the effects of background geostrophic flow on hurricanes Katrina and Rita. Results indicated that when near-inertial wave wakes encounter CCEs they become confined to the OML, which increases the amount of kinetic energy available for shear and results in more cooling of the OML. In contrast, near-inertial wave wakes that encounter the LC or WCEs are able to propagate downward quickly and
kinetic energy is lost from the OML, resulting in decreased cooling. CCEs (positive geostrophic vorticity) also shift the effective frequency of near-inertial waves above \( f \) while WCEs (negative geostrophic vorticity) act in the opposite manner, shifting \( f_e \) down \( (f_e = \left( \frac{\zeta}{2} + f \right) \), where \( \zeta \) is geostrophic vorticity). Jaimes and Shay also suggested that CCEs allow storm-induced upwelling to penetrate farther into the ocean, creating near-inertial waves deep down that propagate energy upward. In the case of hurricane Katrina, interaction with a CCE caused upward-propagating energy to be larger than downward-propagating energy. The CCE, and not the background Brunt–Väisälä profile, was the main cause of increased kinetic energy in shallow depths and was strong enough to shift the main cooling region from right to left of the storm track.

The results of Jaimes and Shay (2010) supported earlier theoretical development and an analytical ray tracing model by Kunze (1985). In that study the effects of background geostrophic flows on \( f_e \), Doppler shift, and propagation direction were examined. When near-inertial waves were steered toward a jet of geostrophic flow, waves normal to the jet were trapped by the warm side of the jet (negative vorticity; CW rotation) and repelled by the cool side of the jet (positive vorticity; ACW rotation). The trapped waves propagated downward into greater depths and their effective frequency was changed. Waves that approached from a non-normal angle and with the flow were advected until the resulting Doppler shift deflected them away from the jet in the form of a beam. Non-normal waves that approached against the flow were trapped. This logic of trapping and downward propagation due to negative vorticity and vice versa can be extended to WCEs and CCEs, as was seen in the Jaimes and Shay (2010) study, where WCEs reduced mixed
layer cooling by allowing near-inertial waves to move out of the OML quickly (i.e. by trapping them in the vertical).

Downward energy propagation due to WCRs and confinement to the OML due to CCEs can have effects on a global scale. Emanuel (2001) postulated that tropical cyclones may be a dominant driver of the Meridional Overturning Circulation (MOC), contributing energy on the order of $10^{15}$ W. The proposed mechanisms for this large energy transfer were near-inertial oscillations moving to lower layers and redistributing heat (e.g. stratification). Changing potential energy caused a heat imbalance which could only be amended by lateral advection away from the tropics. Jaimes and Shay (2009) made a similar calculation, but found that only $10^{11}$ W reached the ocean depths, a value which was modulated by the presence of background vorticity features. Shay and Jacob (2006) found a similar result from hurricane Gilbert current profiles. CCEs kept energy trapped in the OML and warm Loop Current features allowed vertical energy propagation to occur. The main energy sink for trapped oscillations was shear at the MLB, which, according to Shay et al. (1989), is controlled by modes three and four. Therefore, vertical shear is an important indicator of how much near-inertial energy reaches the global oceans, and the near-inertial contribution to it can be measured.

Modal contributions to processes like vertical shear and OML deepening can be studied further using accurate ocean models. To evaluate the use of ocean nowcasts in coupled models for tropical cyclone prediction, Halliwell et al. (2011) completed a case study of Ivan. The NCODA system used several types of data to initialize the ocean features prior to Ivan, including satellite altimetry and SSTs, moored buoys, and Argo floats. Features such as CCEs and WCEs were initialized well as compared to satellite
observations; both features were directly in Ivan’s path and affected its intensity just before landfall. Ivan was simulated accurately with the model forced by a blend of the US Navy’s NOGAPS and NOAA’s H*WIND products (Powell and Houston, 1996). Model experiments captured the post-storm cooling well, although they slightly overestimated the magnitude compared to satellite observations. If models can simulate storms and their effects accurately, they can be used to study features like near-inertial waves with higher sampling rates and greater flexibility than instrumentation arrays. Accurate ocean nowcasts for initialization are key to improving the models to such a point.

These observational and numerical studies show that current measurements are important for evaluating theoretical and numerical predictions of currents and current shears that lead to vertical mixing and upwelling processes induced by net transport away from the storm center (Halliwell et al., 2011). With the Ivan ADCP velocity data acquired over a steep slope, wave characteristics and an analysis of energy propagation can be used to discover new features. That is, these data will provide a revised perspective on the near-inertial response to storm forcing which includes both barotropic and baroclinic oscillations that can be compared to HYCOM simulations.
Chapter 2: Data Resources and Methods

2.1 Main Data Sources

2.1.1 Instrumentation

This study utilizes horizontal and vertical (U, V, and W) velocity measurements from an array of fourteen ADCP moorings in the GOM along the western edge of the DeSoto Canyon. Of the fourteen moorings, eight were selected to focus on: four on the 500 m isobath, and another four on the 1000 m isobath. All moorings measured at ten meter intervals within the range of 42 to 492 m, although each mooring started measuring at a slightly different level. Therefore, the current records were interpolated to a range of 52 to 492 m with 10 m intervals between each step. The canyon slope in this area is 0.06, which may be high enough to reflect near-inertial motions away from the shore if their characteristic slopes are smaller. ADCP measurements used here do not exceed 500 m

![Figure 2.1](image)

**Figure 2.1** Characteristic slopes of near-inertial motions for MS06, MS10, and MS14. Vertical red line is average slope from 100 to 500 m isobath (shaded in red on map inset). Forward reflected (gray arrow) NIWs shaded gray, backward reflected (red arrow) NIWs shaded pink.
depth, and so the characteristic slope of features possibly affected by the canyon wall cannot be calculated. However, the slopes of features from 50 to 500 m indicate backward reflection above 250 m. Backward reflection is due to the upper-canyon slope which, despite being shallow compared to the lower-canyon slope, was still higher than the characteristic slopes of near-inertial motions above 250 m (Figure 2.1).

**Table 2-1** Instrumentation statistics for the ADCP moorings and Aanderaa deep water probes. (Adapted from Teague et al. 2007)

<table>
<thead>
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<th>Mooring Number</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Start Date 2004</th>
<th>End Date 2004</th>
<th>Δt (hr)</th>
<th>Depth Range (m)</th>
<th>Δz (m)</th>
<th>Bottom Depth (m)</th>
<th>Instrument Type</th>
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<td>1</td>
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<td>88.19°W</td>
<td>05/01</td>
<td>10/31</td>
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<td>6-52</td>
<td>2</td>
<td>60</td>
<td>W</td>
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<td>2</td>
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<td>88.01°W</td>
<td>05/01</td>
<td>10/31</td>
<td>0.25</td>
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<td>2</td>
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<tr>
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<td>05/01</td>
<td>10/30</td>
<td>0.25</td>
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<td>2</td>
<td>60</td>
<td>W</td>
</tr>
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<td>10/30</td>
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<td>W</td>
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<td>0.25</td>
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<td>2</td>
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<td>W</td>
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<td>1.0</td>
<td>50-500</td>
<td>10</td>
<td>1025</td>
<td>LR</td>
</tr>
<tr>
<td>16</td>
<td>29.16°N</td>
<td>87.83°W</td>
<td>05/05</td>
<td>11/07</td>
<td>1.0</td>
<td>921</td>
<td></td>
<td>1025</td>
<td>AA RCM9</td>
</tr>
<tr>
<td>17</td>
<td>29.20°N</td>
<td>87.65°W</td>
<td>05/05</td>
<td>11/07</td>
<td>1.0</td>
<td>42-502</td>
<td>10</td>
<td>1029</td>
<td>LR</td>
</tr>
<tr>
<td>18</td>
<td>29.20°N</td>
<td>87.65°W</td>
<td>05/05</td>
<td>11/07</td>
<td>1.0</td>
<td>925</td>
<td></td>
<td>1029</td>
<td>AA RCM9</td>
</tr>
</tbody>
</table>

The ADCP measurements spanned from May 2004 to November 2004 along with four deep water Aanderaa current meters, which measured temperature, pressure, and velocity
at ~950 m depth in one hour intervals (see Table 2-1 for instrumentation statistics). Hurricane Ivan passed directly over the array at 0000 UTC on September 16, 2004. The ADCP and Aanderaa data were sub-sampled to study the periods directly before, during, and after storm passage, September 7 to October 7, 2004. According to manufacturer data, the accuracy of ADCP and Aanderaa current measurements is +/- 1% of the total velocity (0.5 to 1.5 cm s⁻¹ in this case). RD Instruments Long Ranger (LR) ADCPs were used in deep waters along the continental shelf edge, and Workhorse (W) ADCPs were used in shallower waters on the shelf (see Teague et al. 2007 for more details).

Ship-based Conductivity Temperature and Depth (CTD) profiles were cast in the mooring area during May 2004 to measure background stratification. Using the equation of state for seawater from Gill (1982), the buoyancy (a.k.a. Brunt-Väisälä) frequency was calculated using CTD data from east of 88.1°W and south of 29.2°N to avoid fresh water outflow from the Mississippi River. Overall, fifty CTD profiles were used to find background stratification and an average temperature-salinity profile (Figures 2.2 and 2.3). Note that Figure 2.3 also shows a temperature-salinity profile derived from HYCOM fields. Differences between the two profiles are discussed in Section 3.1. Background stratification is an important parameter used in several analyses in this study. However this particular profile, taken in May 2004, depicts a shallower OML than would be present during Ivan’s passage in September 2004. That caveat likely had small impacts on calculations of modal structure in the OML. Pre-storm stratification and its implications (for OML depth, mixing, and stability) will also be discussed in more detail in Chapter 3.
Figure 2.2 Average Brunt–Väisälä frequency (N; cph) from 50 CTD casts. Error bars are two standard deviations.

Figure 2.3 Average temperature-salinity profile from 50 CTD casts (red) and HYCOM temperature and salinity fields from MS07 to MS14 at -5 IP (blue).
2.1.2 HYCOM

HYCOM combines three coordinate systems to describe different parts of the ocean (Bleck, 2002). The vertical z-coordinate is useful for high resolution in the mixed-layer and other unstratified water; the sigma coordinate parameterizes topography well and is utilized near the coast; isopycnal coordinates allow for better resolution of highly stratified areas. Together, these coordinates are used to model smooth transitions between differing areas of the ocean. For the rough topography of the DeSoto Canyon, turbulent post-storm OML, and stratified water surrounding the ADCP moorings, HYCOM is a reasonable choice to explore this storm-forced oceanic response.

A comprehensive sensitivity study was conducted for hurricane Ivan by Halliwell et al. (2011) to evaluate the accuracy of two types of nowcasts in initializing features such as the LC, CCEs, and WCEs. The two nowcasts used were based in HYCOM but utilized different input data, resolutions, and interpolation techniques. Each was compared to observations for several different initialization situations. In the case of hurricane Ivan, the nowcasts parameterized sea surface temperatures (SSTs) and ocean heat content well, capturing intensification of a cold core eddy near the ADCP mooring site. Also notable in the Ivan SST maps is the presence of warm and (likely) fresh water on the shelf above the DeSoto Canyon. By including pre-Ivan discontinuities in oceanic heat content, the H-NCODA nowcast (Cummings, 2005) gave an accurate description of the ocean’s response during and after the storm.

For this study, HYCOM was forced by the United States Navy’s Coupled resolution) blended with surface wind velocity fields from the Hurricane Research Division (HRD) HWIND analysis for the storm region (Powell and Houston, 1996). It was initialized by
### Table 2-2 Summary of model simulations. Left column is model attributes and central column is specific attributes of control experiment GOM1. The right column lists alternate experiments along with the new attribute (from Halliwell et al., 2011).

<table>
<thead>
<tr>
<th>Model Attribute</th>
<th>Control Experiment (GOM1)</th>
<th>Alternate Experiments</th>
</tr>
</thead>
<tbody>
<tr>
<td>Horizontal resolution</td>
<td>0.04° Mercator</td>
<td>GOM2: 0.08° Mercator</td>
</tr>
<tr>
<td>Vertical resolution</td>
<td>26 layers, 4-8m in OML</td>
<td>GOM3: 21 layers, 7.5-15m in OML GOM4: 31 layers, 3-5m in OML</td>
</tr>
<tr>
<td>Vertical mixing</td>
<td>KPP</td>
<td>GOM5: MY GOM6: GISS</td>
</tr>
<tr>
<td>$C_D$</td>
<td>Donelan</td>
<td>GOM7: Powell GOM8: Large and Pond GOM9: Large and Pond (capped) GOM10: Shay and Jacob GOM11: Jarosz et al.</td>
</tr>
<tr>
<td>$C_{EL}, C_{ES}$</td>
<td>COARE3.0 algorithm</td>
<td>GOM12: Kara et al.</td>
</tr>
<tr>
<td>Atmospheric forcing</td>
<td>27-km COAMPS+H*WIND</td>
<td>GOM13: 27-km COAMPS only</td>
</tr>
<tr>
<td>Outer model</td>
<td>NCODA GOM hindcast</td>
<td>GOM14: Free GOM simulation</td>
</tr>
<tr>
<td>Ocean dynamics</td>
<td>Three-dimensional</td>
<td>GOM15: One-dimensional</td>
</tr>
</tbody>
</table>

an Ocean/Atmosphere Mesoscale Prediction System (COAMPS) model (with 27 km GOM hindcast using the NCODA system, which utilizes 26 vertical layers and used the KPP vertical mixing scheme (Large et al., 1994). The NCODA system assimilates the sea surface height (SSH) field from satellite altimeter missions, satellite and in-situ SST, Argo float, moored buoy, and expendable bathythermograph (XBT) data and vertically projects it into the fluid to create a snapshot analysis. To calculate wind stress from the HWIND data, bulk equations were used with a Donelan et al. (2004) drag coefficient. Bulk formulae and the COARE 2.6 algorithm were used to calculate surface turbulent fluxes. The COARE 2.6 algorithm was developed from the TOGA-COARE field campaign and is tailored to calculate fluxes between a convective atmosphere and
tropical ocean (Fairall et al., 1996). A 0.04° horizontal resolution was used for this HYCOM experiment (GOM1; see Table 2-2 for experiment parameters).

### 2.2 Air-Sea Parameters and Price Scaling

Powell et al. (2003) used observations from a NOAA Hurricane Hunter WP-3D scatterometer and dropsondes to derive an empirical curve for the drag coefficient ($C_D$). The curve shows $C_D$ increasing from $1.5 \times 10^{-3}$ to $2.05 \times 10^{-3}$ before leveling off at wind speeds of $30 \text{ m s}^{-1}$. $C_D$ then decreases to $1.5 \times 10^{-3}$ between wind speeds of $40$ to $50 \text{ m s}^{-1}$. Donelan et al. (2004) used a wind-wave tank at University of Miami to derive a similar profile, the main difference being that $C_D$ remains at a constant value of $2.6 \times 10^{-3}$ in $40$ to

![Wind stress (τ; Pa) derived from 10 m dropsonde data taken on 0000 UTC September 16, 2004. H*WIND data from NOAA HRD (Powell and Houston, 1996).](image)

**Figure 2.4** Wind stress ($\tau$; Pa) derived from 10 m dropsonde data taken on 0000 UTC September 16, 2004. H*WIND data from NOAA HRD (Powell and Houston, 1996).
50 m s\(^{-1}\) wind speeds. Donelan et al.’s measurements represent several laboratory methods which all produce the same result, and Powell’s measurements are the first set of observations at high wind speeds. Donelan concluded that the leveling of \(C_D\) at 30 m s\(^{-1}\) winds is likely due to breaking waves, which create an interruption in surface stress as wind skips from the top of one wave to the next. Sea spray may also have a damping effect on \(C_D\) as small particles interact with the wind. Therefore, once hurricane winds reach a certain speed they no longer act to increase stress at the air-sea interface, but instead reach a saturation point.

Wind stress fields from HYCOM (one using the Donelan et al. drag coefficient, and another using Powell et al.) were compared to examine the difference between the two drag curves. Wind stress derived using Powell et al.’s \(C_D\) curve was half the strength of the Donelan wind stress due to the difference between the two curves above 40 m s\(^{-1}\) (where the Powell et al. decreases with wind speed and the Donelan curve stays nearly constant). Resolving this large discrepancy by choosing the most accurate drag coefficient scheme is important for coupled air-sea models to run correctly.

National Hurricane Center (NHC) best track data were used to estimate air-sea parameters for Ivan. Maximum wind stress (\(\tau\)), the main driver of wind-forced currents, was calculated at several time stamps using Powell et al.’s values for drag coefficient and \(H*\text{Wind}\) 10 m wind speed (Figure 2.4; Powell and Houston, 1996). The radius of maximum winds (\(R_{\text{max}}\)) and storm translation speed (\(U_h\)) were also calculated from the wind data and used to derive the horizontal and vertical OML current magnitudes (\(U\) and \(W\)) and vertical isopycnal displacement (\(\eta\); Table 2-3). The expressions for \(U\), \(W\), and \(\eta\) based on air-sea parameters are as follows: 
\[ U = \frac{\tau R_{\text{max}}}{\rho_0 h U_h} \]  

(2.1)

\[ W = \frac{\tau}{\rho_0 U_h} \]  

(2.2)

\[ \eta = \frac{\tau}{\rho_0 f U_h} \]  

(2.3)

where \( h \) is OML depth and \( \rho_0 \) is background density. To determine OML kinetic energy per unit mass, \( U \) is simply squared:

\[ \text{Wind Driven Energy} = U^2 \]

Table 2-3 Key hurricane Ivan air-sea parameters.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>( R_{\text{max}} ) (km)</td>
<td>40</td>
</tr>
<tr>
<td>( T_{\text{max}} ) (N m(^{-2}))</td>
<td>6.7</td>
</tr>
<tr>
<td>( U_h ) (m s(^{-1}))</td>
<td>5.8</td>
</tr>
<tr>
<td>IP (d)</td>
<td>1.02</td>
</tr>
<tr>
<td>( h ) (m)</td>
<td>35</td>
</tr>
<tr>
<td>( g' ) (m s(^{-2}))</td>
<td>0.03</td>
</tr>
<tr>
<td>( c_1 ) (m s(^{-1}))</td>
<td>2.5</td>
</tr>
<tr>
<td>( \alpha_1 ) (km)</td>
<td>43</td>
</tr>
<tr>
<td>( F_r ) (( U_h/c_1 ))</td>
<td>2.2</td>
</tr>
<tr>
<td>( S ) (( U_h/2R_{\text{max}}f ))</td>
<td>1.5</td>
</tr>
</tbody>
</table>

Price (1984) also derived an expression for OML Burger number, half of which is the predicted response carrier frequency:

\[ \text{Carrier Frequency} = \frac{1}{2} M = \frac{1}{2} \left(\frac{1}{\tau} + \frac{1}{2} - \frac{g' h}{2 R_{\text{max}} f^2}\right) \]  

(2.4)

where \( M \) is the OML Burger number, \( S = \frac{U_h}{2 R_{\text{max}} f} \), and \( g' \) is reduced gravity. The
predicted frequency is 1.03f in this case and will be compared to that deduced from ADCP time series measurements. Variations of $U$, $W$, $\eta$, and other Ivan air-sea parameters with time are presented in Figure 2.5. Important to note is that the Price predictions can be corrected to their thermocline values if multiplied by the thermocline Burger number:

$$M_{Th} = M \frac{b}{h}$$

(2.5)

where $b$ is the thermocline depth scale (taken to be 220 m in this case). For instance, the predicted horizontal current value in the thermocline would be $UM_{Th}$ and thermocline kinetic energy is simply the square of that value:

Thermocline kinetic energy per unit mass = $\frac{1}{2} \overline{U_{Th}^2}$

$$\overline{U_{Th}} = \overline{UM} \frac{b}{h}$$

(2.6)

where $b$ is the depth scale of the thermocline, taken to be 220 m based off the CTD casts discussed above. This measure can be used to examine the near-inertial response below.

Taking a close look at Figure 2.5, as Ivan approaches the mooring array (on September 16, 2004), its $R_{max}$ and $U_h$ increase and its $\tau$ decreases leading to higher horizontal current velocities but lower vertical velocities and isopycnal displacement. Although Ivan did decrease in intensity as it entered the GOM (Figure 1.1), its horizontal scale broadened, increasing horizontal momentum input into the OML. Some horizontal current predictions are suspect, for instance values of over 300 cm s$^{-1}$ on September 15, 2004. Most likely an inaccurate estimation of $\tau$ is the culprit, as implied by differences between the Donelan and Powell drag coefficients above at very high wind speeds. The exact values of these air-sea parameters will be an important baseline to compare near-
inertial carrier frequencies and current velocities to a predicted response, so they will be discussed again in Chapter 4.

Figure 2.5 Ivan air-sea parameters from H*wind (Powell and Houston, 1996) and Price (1983) scaling: Rmax, Uh, τ, U, W, and η from September 10, 2004 to September 18, 2004. Ivan passed over the moorings on September 16, 2004. Predicted wind-driven horizontal (U) and vertical (W) current values are 112.4 cm s⁻¹ and 0.15 cm s⁻¹, respectively. Predicted wind-driven kinetic energy is 1.2x10⁴ cm² s⁻² per unit mass.

2.3 Tides

One problem with analyzing near-inertial motions is that the ambient diurnal tides tend to be the same frequency at 30°N and must be removed using harmonic analysis software. Frequencies determined from the positions of the Earth, moon, and sun are used to identify sinusoidal equations that represent known tidal constituents. Reconstructed tides are then least squares fit to the data, determining their respective amplitudes and phases. In addition to the barotropic tides being stationary relative to near-inertial
motions, another way of distinguishing between the two is that tides exhibit smaller amplitudes and consistent phases (~6 cm s\(^{-1}\)) compared to wind-forced near-inertial waves (100 to 150 cm s\(^{-1}\) in the mixed layer; 30 to 60 cm s\(^{-1}\) in the thermocline). Harmonic analysis software known as Ttide (Pawlowicz et al., 2002) was used to remove eight diurnal and semi-diurnal tidal constituents (M2, S2, N2, L2, Q1, K1, O1, J1; Table 2-4) over a six month period (that of the entire data set). To verify the software, tides were analyzed during June 2004 in which there were no storms. The software found that tidal amplitudes were 4 to 6 cm s\(^{-1}\), a reasonable estimate for the area (Seim et al., 1987).

The six month time period was chosen to maximize efficiency of the harmonic analysis software. Time periods that are approximately one year are long enough to resolve most tidal frequencies and short enough to assume that longer term frequencies are constant over the time span. Time periods that are too short are less suited for fitting longer term tidal frequencies and may lead to a less desirable result. According to Pawlowicz et al. (2002):

There are several drawbacks to classical harmonic analysis. The first is that…an ~18.6 year time series is required to resolve all of the listed frequencies….In practice, record lengths are often 1 year or shorter….The appearance of the total signal will be a sinusoid whose phase and amplitude varies slowly with time. These changes are slow enough to be considered effectively constant for record lengths of up to 1 year. At much shorter record lengths another problem arises. The frequency resolution further degrades until even dissimilar constituents are unresolvable.

Several time periods (one to six months) were used with the de-tiding software to find the most desirable output, and Table 2-5 details the resulting tidal amplitudes. The six
Table 2-4. Total current amplitudes (cm s⁻¹) of the top four most energetic tidal constituents removed by $T_{tide}$.

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>O1</th>
<th>J1</th>
<th>K1</th>
<th>Q1</th>
</tr>
</thead>
<tbody>
<tr>
<td>100</td>
<td>1.10</td>
<td>0.90</td>
<td>2.00</td>
<td>0.30</td>
</tr>
<tr>
<td>200</td>
<td>1.30</td>
<td>1.85</td>
<td>0.80</td>
<td>0.25</td>
</tr>
<tr>
<td>300</td>
<td>0.80</td>
<td>0.90</td>
<td>0.60</td>
<td>0.50</td>
</tr>
<tr>
<td>400</td>
<td>1.00</td>
<td>0.55</td>
<td>0.45</td>
<td>0.50</td>
</tr>
<tr>
<td>500</td>
<td>0.30</td>
<td>0.50</td>
<td>0.50</td>
<td>0.40</td>
</tr>
</tbody>
</table>

Month tidal analysis returned the most favorable results, especially compared to an earlier analysis completed using a 30-day time span. Tidal amplitudes were within the expected satisfactorily.

Table 2-5 Tidal amplitudes derived using differing time spans of data (one month of data, two months of data, etc.).

<table>
<thead>
<tr>
<th>Months Used</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tidal Amplitude (cm s⁻¹)</td>
<td>42.7</td>
<td>19.2</td>
<td>15</td>
<td>10.1</td>
<td>8</td>
<td>6</td>
</tr>
</tbody>
</table>

4 to 6 cm s⁻¹ range (according to Seim et al., 1987) leaving the predominant storm-induced ocean response intact after harmonic analysis (Figure 2.6). Due to the large near-inertial response from Ivan, six months is required to resolve all tidal constituents.

After tidal currents were removed from the velocity data, the detided data were Fourier transformed into separate frequency bins. Periods of 15 to 35 hours (defined here as the near-inertial band) were selected for the corresponding frequency bins and inverse Fourier transformed to construct the band-passed signal and isolate oscillatory motions observed in the time series (Walters and Heston, 1982). The main reason for choosing the 15 to 35 h period range is that near-inertial motions have a period of 24.5 h at 29.2°N.
Figure 2.6 MS14 detided zonal current velocity at 50 m depth: raw data (blue), tides (red), and observed current velocity data (green). 0IP marks the passage of hurricane Ivan.

More specifically, however, the 15 to 35 h band recovers the largest amount of mean kinetic energy from the detided data (68%; Table 2-6) when used for a time period of 0 to 10 IP. Band pass filtering introduces no variance to the data. After the detided data were filtered, the depth average (barotropic component)

Table 2-6 Percent of mean kinetic energy accounted for by filtered signal at MS14, 50m.

<table>
<thead>
<tr>
<th>Filter Interval (h)</th>
<th>Time Interval (IP)</th>
<th>0-3</th>
<th>0-5</th>
<th>0-7</th>
<th>0-10</th>
<th>0-15</th>
<th>0-20</th>
<th>0-25</th>
</tr>
</thead>
<tbody>
<tr>
<td>20-30</td>
<td></td>
<td>33</td>
<td>31</td>
<td>39</td>
<td>51</td>
<td>44</td>
<td>48</td>
<td>42</td>
</tr>
<tr>
<td>20-35</td>
<td></td>
<td>33</td>
<td>50</td>
<td>52</td>
<td>60</td>
<td>56</td>
<td>55</td>
<td>52</td>
</tr>
<tr>
<td>18-30</td>
<td></td>
<td>43</td>
<td>31</td>
<td>46</td>
<td>54</td>
<td>48</td>
<td>53</td>
<td>47</td>
</tr>
<tr>
<td>18-35</td>
<td></td>
<td>43</td>
<td>50</td>
<td>59</td>
<td>64</td>
<td>60</td>
<td>61</td>
<td>56</td>
</tr>
<tr>
<td>15-30</td>
<td></td>
<td>43</td>
<td>43</td>
<td>52</td>
<td>59</td>
<td>52</td>
<td>57</td>
<td>51</td>
</tr>
<tr>
<td>15-35</td>
<td></td>
<td>43</td>
<td>62</td>
<td>65</td>
<td>68</td>
<td>64</td>
<td>65</td>
<td>61</td>
</tr>
</tbody>
</table>

latitude (the latitude of the moorings).
was removed from the band-passed signals and stored separately. Note that moorings 11 to 14 only measure a partial water column, so their depth averages may not reflect the total depth-averaged near-inertial response. Because the eight deep-water moorings start measuring at ~50 m depth, they do not include the pre-storm OML, where wind-driven velocities can exceed 100 cm s\(^{-1}\) during hurricane passage. HYCOM output did not include tides, so raw data were band-pass filtered using the above method.

### 2.4 Wave Characteristics

With this preliminary processing completed, the Ivan data are ready for an examination of the modes' wave characteristics, such as wavelength and frequency. First, the velocity vectors were rotated relative to the storm track (15° CW from due north), where kinetic energy was conserved in the coordinate rotation. To determine the carrier frequency, defined as the frequency that minimizes residual variance, a least squares fit was created between the measured currents and a model using a set of trial frequencies (0.75\(f\) to 1.35\(f\) with a 0.01\(f\) increment). The following expressions were used:

\[
U(t) = A_1 \sin(\sigma t) + B_1 \cos(\sigma t) \quad (2.7)
\]

\[
V(t) = A_2 \sin(\sigma t) + B_2 \cos(\sigma t) \quad (2.8)
\]

After varying the frequency of the near-inertial solution for each fit, the carrier frequency was chosen from the solution with the highest correlation to the data. These fits were performed over an interval of 2 to 5 IP for two reasons: to eliminate any contamination from the initial response and to capture the presence of a CCE during that period (see Section 3.3; Jaimes and Shay (2010) use similar methods and reasoning). Contour plots of the correlation coefficient against depth and carrier frequency allowed
for an analysis of frequency evolution over time and space in the ADCP data and HYCOM output.

To determine wavelength, the coefficients from the carrier frequency-containing solution to the least squares fit were used to generate CW- and ACW-rotating velocity components (Gonella, 1972). Using CW velocity components and modeling the near-inertial oscillations as a circular rotating motion, phases were found by taking the inverse tangent of the along-track over the cross-track velocity component (V/U), and corrected for each quadrant. The phases were unwrapped (made continuous) to avoid phase jumps which would interfere with analysis. Next, distances between moorings were calculated using the Haversine formula (Table 2-7; Gellert et al., 1989), a geometric method employed to determine distance over a curving surface:

\[
a_x = \sin\left(\frac{d\text{lat}}{2}\right)^2 + \cos(lat_1) \times \cos(lat_2)
\]

\[
\times \sin\left(\frac{d\text{lon}}{2}\right)^2
\]

\[
c_x = 2 \times \arcsin(\sqrt{a_x})
\]

\[
d_x = R \times c_x
\]

where dlat is the difference in latitude between lat$_1$ and lat$_2$ of the two points, dlon is the difference in longitude, d$_x$ is the distance between the two points, and R is the radius of the Earth. With the phase shifts and distances, an attempt was made to find the wavelength of the response and how it changed over time. To eliminate the possibility of background flow shifting the phase from one mooring to the next, the carrier frequencies of each mooring were averaged at one depth level. The average frequency was then used
with the least squares model equations to derive coefficients for the complex
demodulation process. In this way, an average phase was found for a given set of
moorings, as opposed to a phase which was shifted by background geostrophic flows. An
averaging period of 7 IP was used, as it gave the best result (similar to Shay and Elsberry,
1987).

### 2.5 Baroclinic Mode Analysis

A Runge-Kutta (RK) fourth order numerical solver was used with the -52 to -492 m
depth interval and background Brunt–Väisälä profile to solve the wave equation and find
wavelengths and phase speeds of individual modes. The following equations are used:

\[
\psi_n(x, z) = Z_n(z) \exp(jkx) \tag{2.12}
\]

\[
\frac{\partial^2 \psi}{\partial z^2} - \gamma^2 \frac{\partial^2 \psi}{\partial x^2} = 0 \tag{2.13}
\]

\[
\frac{\partial^2 Z_n}{\partial z} + \left(\frac{k}{\gamma}\right)^2 Z_n = 0 \tag{2.14}
\]

\[
\gamma^2 = \frac{(\sigma^2 - f^2)}{(N^2 - \sigma^2)} \tag{2.15}
\]

where \( \psi \) is the perturbation streamfunction, \( k \) is the horizontal wavenumber, \( \gamma^2 \) is the
slope of the internal wave characteristic, and \( Z_n \) is the vertical structure of the \( n \)th mode.

Equation (2.12) is the plane wave solution (assuming a changing \( N^2 \) profile and a flat

<table>
<thead>
<tr>
<th>Moorings</th>
<th>MS07 to MS08</th>
<th>MS08 to MS09</th>
<th>MS09 to MS10</th>
</tr>
</thead>
<tbody>
<tr>
<td>Distance (km)</td>
<td>13.8</td>
<td>13.1</td>
<td>13</td>
</tr>
<tr>
<td>Moorings</td>
<td>MS11 to MS12</td>
<td>MS12 to MS13</td>
<td>MS13 to MS14</td>
</tr>
<tr>
<td>Distance (km)</td>
<td>15.9</td>
<td>14.7</td>
<td>13.1</td>
</tr>
</tbody>
</table>
bottom), (2.13) is the governing wave equation assuming harmonic time dependence and 2D flow, and (2.14) is the structure equation for baroclinic modes.

The solver starts with an initial set of modal amplitudes based on the first depth and Brunt–Väisälä frequency values. It then calculates the next set of modal amplitudes by taking a weighted average of possible values over the depth interval (step interval) and adding that average to the previous value. To find the first ten baroclinic modes, the solver used rigid surface and bottom boundary conditions. The depth average was already subtracted from the data, so there was no need to include a free surface boundary condition as it has little or no effect on the baroclinic modes. According to (Shay and Chang, 1997) the equation governing the barotropic response effect on baroclinic structure is:

\[
m_n = \frac{n\pi}{D} \left[ 1 + \left( \frac{N^2}{(nn)^2} \frac{g}{D} \right) \right] \quad (2.16)
\]

where \( n \) is the mode number, \( D \) is the depth, and \( N \) is the Brunt–Väisälä frequency (Figure 2.7). The second term represents free surface effects and is \( O(10^{-8}) \), making it too small to affect the baroclinic structure. The numerical solver generated wavelengths using a changing stratification profile instead of a single value, a calculation which reflects realistic background stratification.

Vertical group speed was calculated using the dispersion relationship for near-inertial baroclinic modes from Brooks (1983):

\[
c_v = -\left( \frac{\sigma}{m} \right) \left( \frac{N^2}{\sigma^2} \right) \left( \frac{k^2}{m^2} \right) \quad (2.17)
\]
Figure 2.7 Baroclinic mode structure of dominant modes one through three calculated numerically using the background stratification profile and mode structure equations.

where $\sigma$ is carrier frequency (from 2 to 5 IP), $m$ and $k$ are vertical and horizontal wavenumbers of mode 1, respectively, and $N(z)$ is the background Brunt-Väisälä frequency profile. These predicted values of group speed will give an idea of how quickly near-inertial energy should propagate vertically into the ocean interior and allow for a comparison to observations.

Downward vertical group speeds (Figure 2.8) were up to -0.27 cm s$^{-1}$ in the upper thermocline and, judging from the stratification profile, mixed-layer depth was 20 to 30 m in the mooring area. Again, the CTD observations averaged to obtain the background stratification profile were taken in May 2004, several months before Ivan’s passage through the area. However, this profile reflects relatively quiescent ocean conditions and
Figure 2.8 Vertical group speed calculated using the baroclinic near-inertial dispersion relation from Brooks (1983) (Figure 2.7), the average background stratification profile (Figure 2.3), and carrier frequencies at all depths during 2 to 5 IP with error bars.

is taken from many CTD casts, so derived values of buoyancy frequency, vertical group speed, and OML depth represent an average background state.

After resolving wave characteristics, an examination was made of baroclinic modes and their contributions to the total response. Using the modal amplitudes, a least squares fit was performed between each mode and the detided, filtered, and normalized current data. Once a mode was fit to the data, the result was removed from the data to eliminate any redundant fitting. The individual modes were then combined to form a modeled current response from the product of the least squares fit amplitudes and the vertical
modal structures. Five modes were required to reconstruct the observed currents so that the difference between the model and the observations was minimized. Using fewer modes did not resolve all of the velocity features and using more modes made little difference in the result.

Other predictions are of interest; for example, according to Geisler’s (1970) linear theory, the expected response wavelength (translation speed (U_h) times 1 IP) is 483 km. This gives an idea of the predicted near-inertial horizontal scale. Also, mode separation times were calculated using phase speeds from the RK computation and the following expression from Gill (1984):

\[ t_n = \frac{\pi f}{k^2 c_n^2} \]  

(2.18)

where \( t_n \) is the time of separation, \( k \) is the inverse scale of the wind stress curl, and \( c_n \) is the phase speed of mode \( n \). Clearly, it is important to know when modes will separate from the OML and propagate downward into the thermocline. The separation times for modes one through three were as follows: \( t_1 = 2.5 \text{ d} \), \( t_2 = 5.1 \text{ d} \), and \( t_3 = 9.6 \text{ d} \).

### 2.6 Background Flow

To analyze the effects of background flow, potential vorticity was calculated from the ADCP velocity data. Detided data were low-pass filtered to 48 h were used for the analysis. Following the method of Emery and Thompson (2001; Ch. 3), three moorings (MS10, MS13, and MS14), positioned in a triangular array, were chosen to perform a linear fit between their velocity components and distances relative to each other:
\[ \zeta(x_c, z) = \frac{\partial v}{\partial x}(x_c, z) - \frac{\partial u}{\partial y}(x_c, z) \]  

(2.19)

\[ \begin{bmatrix} 1 & x_1, y_1 \\ 1 & x_2, y_2 \\ 1 & x_3, y_3 \end{bmatrix} \begin{bmatrix} \frac{\partial u}{\partial x}(x_c, z) \\ \frac{\partial u}{\partial y}(x_c, z) \end{bmatrix} = \begin{bmatrix} u_1(x, z) \\ u_2(x, z) \\ u_3(x, z) \end{bmatrix} \]  

(2.20)

where \( x_c \) is the central point at which vorticity is calculated, \( u \) represents the two horizontal velocity components, \( z \) represents depth, and subscripts 1, 2, and 3 represent the three moorings. Vorticity was calculated at point \( x_c \) from -52 to -492 m and -5 to 20 IP using a 5 IP moving window. The same process was completed for 48 h low pass filtered HYCOM output data at the same location. Vorticity values were recast as an effective Coriolis parameter, \( f_e \). The equation \( f_e = \zeta g / 2 + f \) was used and the result divided by \( f \) to normalize in terms of the background planetary vorticity.

2.7 Shear Analysis

Ivan’s winds imparted momentum at the air-sea interface, forcing large vertical shear between the MLB and relatively quiescent thermocline below. Shear causes turbulent mixing at the MLB, which allows colder water to mix with the warm water above. Such a process deepens the OML, affects ocean heat content, and must be predicted accurately; therefore vertical shear will be quantified using the current data and compared to the HYCOM model.

Due to its importance, the vertical shear’s spatial and temporal extent was documented as well as contributions from individual velocity components and high order baroclinic modes. Along-track, cross-track, and total baroclinic velocities were used to calculate
shear over the shelf and on the slope for comparison. A time interval of -5 to 20 IP and depth range of 50 to 490 m, with an interval of 10 m, were used. Motivation for this analysis was to find how long and intense the shear was and eventually determine if particular modes were responsible for shear at particular depth levels. Gradient Richardson numbers were calculated using the background Brunt–Väisälä frequency and locations of possible vertical instability were noted.

Baroclinic modes one through ten (discussed above) were also least-squares fit to HYCOM and ADCP current data from 50 to 490 m, so RK output from 10 to 50 m was discarded before the fit. HYCOM data (available for the entire ocean depth) were least squares fit a second time, but in a depth range of 10 to 50 m at 10 m intervals to capture the OML. Sums of modal contributions to vertical shear from measured and modeled currents were used to determine the MLB and upper thermocline depths. Percent of shear variance due to individual modes was calculated at each depth, and the dominant modes were noted.

2.8 Objectively Analyzed Currents

2.8.1 Objective Analysis of Horizontal Currents

Using the OAX5 objective analysis package (developed by the Bedford Institute of Oceanography based on the method of Bretherton, et al. (1976)), detided HYCOM and ADCP currents were mapped over the mooring area. OAX5 uses a nearest-neighbor technique to determine values at pre-selected grid points using the input data. Near-inertial energy disperses throughout the ocean and spreads vertically and horizontally, so the vertical and horizontal coordinates (in meters and degrees, respectively) were scaled according to Shay and Uhlhorn (2008; Table 2-8). Although the original vertical scaling
Table 2-8 Horizontal and vertical correlation scales used in the objective analyses of Shay and Uhlhorn (2008).

<table>
<thead>
<tr>
<th>Depth Range (m)</th>
<th>Horizontal Scale L (°)</th>
<th>Vertical Scale (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-50</td>
<td>0.5</td>
<td>20</td>
</tr>
<tr>
<td>50-100</td>
<td>1.5</td>
<td>40</td>
</tr>
<tr>
<td>100-150</td>
<td>2.0</td>
<td>60</td>
</tr>
<tr>
<td>150-200</td>
<td>2.5</td>
<td>80</td>
</tr>
<tr>
<td>200-400</td>
<td>3.0</td>
<td>100</td>
</tr>
<tr>
<td>400-600</td>
<td>4.0</td>
<td>150</td>
</tr>
<tr>
<td>600-700</td>
<td>5.0</td>
<td>200</td>
</tr>
</tbody>
</table>

was maintained, the horizontal scaling from 0 to 50 m was changed to 0.2° as opposed to 0.5° and achieved a better result for this case. The mapping took place at ½ IP intervals for the first 2.5 IP at 50 m. One observed current vector was plotted on top of an objectively analyzed vector at the same location to verify the accuracy of OAX5 in this situation (not shown). Both vectors were approximately the same magnitude and direction at all time stamps.

2.8.2 Kinetic Energy

Objectively analyzed currents were then used to calculate total kinetic energy at 50 m which was normalized by the predicted value of Price (1983) from Section 2.2 and contoured beneath the current vectors. Normalizing by the Price prediction (1.2\times10^4 \text{ cm}^2 \text{ s}^{-2} \text{ per unit mass}, in this case) allows for a comparison between the expected and actual energy values.

Total kinetic energy from rotating baroclinic near-inertial waves can be split into CW- and ACW-rotating components. CW-rotating (with depth, as viewed from above) currents tend to propagate energy downward, and ACW-rotating currents propagate energy upward (Leaman, 1976), so currents’ CW and ACW kinetic energy values show
the direction of energy transfer at different depths. Shay and Jacob (2006) found from hurricane Gilbert profiles that CW energy is up to four times stronger than ACW energy, so only baroclinic CW energy was calculated for MS07 to MS10 at all depths. Detided and filtered ADCP and HYCOM horizontal currents were utilized for the calculation. Using the method detailed in Section 2.5, 5 IP time intervals of currents were taken, centered at the desired time stamp, and used to generate the required coefficients for the CW kinetic energy calculation. CW velocity components ($U_{CW}$ and $V_{CW}$) were then squared, added, and divided by two to obtain the CW kinetic energy in cm$^2$ s$^{-2}$ per unit mass. CW kinetic energy was calculated at all depths below 50 m, so values were normalized by the Price (1983) prediction for thermocline kinetic energy (from Section 2.2). The prediction for horizontal wind-driven currents in the OML is multiplied by the thermocline-adjusted Burger number to determine wind-driven thermocline currents, which are then squared to give the value for thermocline kinetic energy.

### 2.9 Depth-Averaged Mass Divergence

Shay and Chang (1997) found a link between the barotropic near-inertial response associated with the free surface and depth-averaged mass divergence. Density, temperature, and salinity fields were not available at the ADCP site during Ivan’s passage, so mass divergence cannot be calculated there. However, temperature and salinity were available from the HYCOM output and were used with Gill’s (1982) equation of state to determine density at MS12 for a set of interpolated depths (10 to 510 m with a 10 m interval) and over a time span of -5 to 10 IP. OAX5 was used to generate a nine-point square grid (with horizontal distances of 0.05°) at each depth with MS12 at the center, and the grid was used to calculate the density gradient across MS12 in all three
directions \((x, y, \text{ and } z)\). For each time point, the three density gradients were multiplied by their corresponding non-filtered current velocity components \((u, v, \text{ and } w)\) at MS12. The three terms were depth-averaged separately, added together, and plotted against time using the following equation:

\[
[u\rho_x] + [v\rho_y] + [w\rho_z] = DAMD
\]

where \(u, v, \text{ and } w\) are the three velocity components, \(\rho_x, \rho_y, \text{ and } \rho_z\) are density gradients in three directions. The brackets indicate that each value is depth averaged.
Chapter 3: Pre-Storm Conditions

3.1 Remotely Sensed Fields

The Systematically Merged Atlantic Regional Temperature and Salinity (SMARTS) Climatology, which utilizes temperature and salinity profiles from the World Ocean Atlas 2001 and satellite-derived SHAs, was used to determine 20°C isotherm depths (h20) and sea surface temperatures (SSTs) in the moorings’ vicinity (Meyers et al., 2013). SMARTS represents a 37% reduction (compared to a previous climatology used by Mainelli (2000)) in the h20 root mean squared deviation (RMSD) from over 60,000 in situ profiles. It also has a much smaller bias (<1 m) compared to that of the Mainelli climatology (35 m).

Sea surface height (SSH) measurements from three satellites, Envisat, Jason-1, and Geosat Follow-On (GFO), were used to examine the height field left in Ivan’s wake. Satellite measurements were taken during ten day intervals (September 6 – 16 and September 17 – 27, 2004 in this case) and were available directly before and after Ivan’s passage. They give a general view of the larger scale variability in the GOM. Data from all three satellites were broken down into individual days, objectively analyzed using the Mariano and Brown (1992) method, and averaged using a moving 10-day window. SHA data from September 13 and 16, 2004 were differenced to determine the change in sea surface height from before to after Ivan. Change in relative vorticity was calculated from the SHA difference to show Ivan’s effects on that parameter. Surface height fields were available at two hour intervals for the entire HYCOM simulation, so fields at 0000UTC on September 15, 2004 (before Ivan) were subtracted from those at 1200UTC on
September 16, 2004 (after Ivan) to determine height change. The pre-storm state found using these remote measurements and its possible effects on Ivan and the near-inertial response will be discussed in the next section.

### 3.2 Pre-Storm Conditions

Sea surface height anomalies (SHAs) derived from satellite observations in the GOM show a CCE centered at 25°N/87°W, a WCE at 27°N/88°W, and a much smaller CCE at 27.5°N/87°W (Figure 3.1). SMARTS h20 depths and satellite-derived relative vorticity

![Figure 3.1 Altimetry-derived surface height anomaly (SHA; in cm) before (top) and after (bottom) Ivan’s passage.](image-url)
Figure 3.2 Depth of the 20°C isotherm ($h_{20}$; from SMARTS) contoured under satellite-derived geostrophic current vectors on September 16, 2004.

Figure 3.3 Relative vorticity ($s^{-1}$) derived from satellite altimetry data before (September 13, 2004) and after (September 16, 2004) Ivan. Geostrophic velocity vectors are plotted on top of the relative vorticity contours.
support the altimeter-observed presence of all three eddies in the GOM before Ivan (Figures 3.2 and 3.3). This particular layout of eddies along Ivan’s track (CCE, followed by a WCE, then another CCE) exposed Ivan to cooler SSTs (causing it to weaken) and warmer SSTs (causing it to intensify) all within a period of 24 hours before it reached the moorings (Figure 3.4). Halliwell et al. (2011) used microwave satellite measurements from the Tropical Rainfall Measurement Mission’s (TRMM) Microwave Imager (TMI) and Advanced Microwave Scanning Radiometer for the Earth Observing System (AMSR-E) instruments to determine SSTs before (September 10, 2004) and after (September 17, 2004) Ivan’s passage. They found good agreement between observed and modeled features, and post-storm cooling was captured in the correct locations and at the correct magnitude (see Halliwell et al., 2011: Figure 8). SSTs from the SMARTS

**Figure 3.4** Pressure at Ivan’s center (mb) from 1400 UTC September 14, 2004 to 1618 UTC September 16, 2004. Interactions with a large CCE (blue line) and WCE (red line) are marked. 0000 UTC on September 16, 2004 is 0 IP.
climatology show initial temperatures up to 29°C in the northeast GOM, followed by cooling by up to 5°C over the large and small CCEs, and cooling of 2°C over the WCE under Ivan’s track (Figure 3.5). SSTs in the mooring area decrease by approximately 2°C after Ivan’s passage. Such a placement of changing SSTs under Ivan’s track supports the intensity changes seen as it moved north across the GOM (Walker et al., 2009).

Figure 3.5 Sea surface temperatures (°C) derived from SMARTS climatology before (September 13, 2004) and after (September 16, 2004) Ivan. Black stars off the Louisiana coast are the NRL SEED ADCP moorings.
Chapter 4: Characteristics of the Response

4.1 Current Amplitudes

Unless otherwise specified, velocities from detided and band pass filtered records will be used throughout this section. In addition, velocities will be normalized by background stratification and by Price (1983) scaling arguments. The latter scaling allows one to directly compare the measured and simulated current responses to those predicted from scaling arguments based on air-sea interaction parameters. Scaling arguments essentially reflect the near-inertial response since the Froude number exceeds 2 (see Table 2-3); a strong baroclinic response and weak barotropic response are expected (Geisler, 1970).

The moorings’ maximum near-inertial amplitudes are 70 to 80 cm s⁻¹ in the upper thermocline 0 to 5 IP after Ivan’s passage (Figure 4.1). Velocity profiles at MS09 and MS12, under the storm center, show currents rotating clockwise with depth (Figure 4.2) consistent with wind-forced near-inertial wave dynamics excited by Ivan’s surface wind field. Compared to predicted OML currents ($U$), the near-inertial current response at 50 m was 15 to 20 cm s⁻¹ lower. However, velocities at 50 m depth are likely lower than

![Figure 4.1 Detided and filtered zonal currents at MS14 at the 50 m depth level.](image)
those in the OML and higher than those in the thermocline, so that result is expected.
Both the HYCOM simulations and ADCP data indicate near-inertial oscillations of 20 to 30 cm s\(^{-1}\) persisting for up to two weeks following the initial storm passage, similar to those observed by Shay and Elsberry (1987) after hurricane Frederic. Thermocline horizontal currents from both the model and observations were two to three times higher than the Price thermocline prediction of \(U_{Th} = 45.3 \text{ cm s}^{-1}\) at MS07 and MS08, but very close to the prediction at MS10, MS11, and MS13 (Figures 4.3 and 4.4). The ADCP near-inertial vertical velocity component, has maximum amplitudes of 0.10 to 0.15 cm s\(^{-1}\) at the base of the OML and bottom of the column, but reaches two to three times the Price
Figure 4.3 ADCP near-inertial zonal current velocities left of the storm at -0.38 Rmax (MS07), at storm center (MS08, MS11), and right of the storm at ~1 Rmax (MS10, MS13). Normalized by Price (1983) thermocline current velocity.

Figure 4.4 HYCOM zonal near-inertial current velocities left of the storm at -0.38 Rmax (MS07), at center of storm (MS08, MS11), and right of the storm at ~1 Rmax (MS10, MS13). Normalized by Price (1983) thermocline current velocity.
Figure 4.5 ADCP vertical near-inertial current velocities left of the storm at ~0.375 Rmax (MS07), at center of storm (MS08, MS11), and right of the storm at ~1 Rmax (MS10, MS13). Normalized by Price (1983) thermocline current velocity.

prediction of $W = 0.15$ cm s$^{-1}$ in some areas (Figure 4.5). At MS10, the vertical response is particularly strong and exceeds 0.30 cm s$^{-1}$, indicating possible calibration error in the mooring’s vertical velocity measurement. Typically, the first three baroclinic modes represent the highest near-inertial current amplitudes. Least squares analysis allowed individual modes to be separated from the total response. Mode one has the highest response amplitude, 50 to 60 cm s$^{-1}$, and dominates the near-inertial currents (Figure 4.6). Its along-track amplitude decreases slightly from MS07 to MS10, however it increases from MS07 to MS10 in the cross-track direction. Most likely, the increase reflects the movement of measurement location from left to right of the storm track where wind speeds are higher and lead to an amplified response from one to three R$_{max}$, as suggested
in previous studies. That is, resonant excitation when the CCW-rotating wind stress field and CW-rotating near-inertial currents are essentially in phase, which causes an amplified response on the right-hand side of the storm (Price, 1983; Greatbatch, 1983).

Figure 4.6 Least squares fit of the first baroclinic mode to cross-track near-inertial current data (cm s$^{-1}$). A) MS07; B) MS08; C) MS09; D) MS10.

4.2 Rotary Kinetic Energy

Objectively analyzed current velocities and kinetic energies are larger on the right side of the storm in deeper water in agreement with the rightward bias discussed in several previous studies (e.g. Chang and Anthes, 1978). Additionally, Gill (1984), found that OML energy loss rate scales with OML depth, which is usually deeper on the storm’s
right side, due in part to the resonant excitation discussed above. Therefore, at the base of a deep OML, currents will intensify more due to rapid energy loss. HYCOM overestimates the wind-driven kinetic energy by up to 1.7 times Price’s (1983) value and predicts more energy over the shelf during storm passage, followed by a switch over the canyon 0.5 IP later (Figure 4.7). ADCP data show almost perfect agreement with the predicted value and place the highest kinetic energy values just right of the storm track (Figure 4.8). Kinetic energy values decrease to between 0.2 and 0.4 scaled units after 1.5 IP for both HYCOM output and ADCP data.

**Figure 4.7** Objective analysis of ADCP currents at 50 m (vectors; cm s\(^{-1}\)) and their associated kinetic energy (contours; cm\(^2\) s\(^{-2}\)) at 0, 0.5, 1, 1.5, 2, and 2.5 IP. Kinetic energy is normalized by the Price (1983) predicted OML value for comparison.
Figure 4.8 Objective analysis of HYCOM currents at 50 m depth (vectors; cm s$^{-1}$) and their associated kinetic energy (contours; cm$^2$ s$^{-2}$) at 0, 0.5, 1, 1.5, 2, and 2.5 IP. Kinetic energy is normalized by the Price (1983) predicted OML value for comparison.

Figure 4.9 Total CW kinetic energy (cm$^2$ s$^{-2}$) derived from ADCP measurements (top panels) and HYCOM output (bottom panels) at MS07 to MS10. Values are normalized by the Price (1983) thermocline kinetic energy.
Rotary kinetic energy analysis revealed several differences between HYCOM output and ADCP data. Maximum CW-rotating (with depth; as viewed from above) energy values were similar for both data sets (Figure 4.9). However, both data sources show differing amounts of energy escaping into the thermocline, particularly at MS07 and MS08, possibly due to HYCOM’s inability to account for some unknown process. ADCP data at MS07 and MS08 indicate a strong downward energy burst into the thermocline at 2 IP, whereas HYCOM shows less than half as much energy escaping the OML. Differences between CW kinetic energy from ADCP and HYCOM will be discussed in Section 4.6. Vertical group speeds (up to 0.25 cm s\(^{-1}\) in the thermocline) calculated from the baroclinic near-inertial dispersion relation are in Figure 2.10.

### 4.3 Wave Characteristics

The carrier frequency calculation showed several distinct frequencies in the time range of 2 to 5 IP. Although carrier frequencies evolve over space and time, an ACW-rotating feature is present (see Section 3.3) from 0 to 7 IP and in its most intense state from 2 to 5 IP, making this a time range of interest. Jaimes and Shay (2010) completed a similar analysis over 3 IP time ranges that overlapped with background vorticity features after hurricanes Katrina and Rita. In a range of 50 to 100 m at MS14, the carrier frequency was 1.05f (where f is the local Coriolis parameter), from 120 to 200 m it was 1.01f, from 200 to 400 m it was shifted up to 1.12f, and from 400 to 490 m it maintained the same value (1.12f) (Figure 4.10). Due to the one-hour sampling interval, possible error on these estimates is +/-0.08f. Thus, some inaccuracy may exist in carrier frequency measurements at these various levels. Regardless, the carrier frequencies are blue shifted above f as shown in previous studies.
Figure 4.10 Contour of correlation between the solution of equations (2.7) and (2.8) at various carrier frequencies and ADCP current response at MS14, 2 to 5 IP. Black line is background inertial frequency ($f$) and blue line is Price (1983) predicted carrier frequency.

Figure 4.11 Contour of correlation between the solution of equations (2.1) and (2.2) at various carrier frequencies and HYCOM current response at MS14, 2 to 5 IP. Black line is background inertial frequency ($f$) and blue line is Price (1983) predicted carrier frequency.

Based on Price scaling, the predicted carrier frequency of the response in a quiescent
ocean is \( 1.03f \), so frequency changes in the vertical (shifted above the predicted frequency from 50 to 100 m but below it from 120 to 200 m; similar to Figure 6a in Jaimes and Shay (2010)) suggest the influence of a CCE feature in the 50 to 200 m depth range from 2 to 5 IP. HYCOM carrier frequencies essentially match those from ADCP data remarkably well and perhaps reflect the accurate initialization of HYCOM fields to include a CCE in the mooring area (Figure 4.11).

Post-storm background vorticity features may be responsible for the carrier frequency shifts. ACW-rotating features act to increase near-inertial carrier frequencies in shallower depths and decrease frequencies below, and CW-rotating features act in the opposite sense. Near-inertial energy can be confined to the OML by ACW-rotating features, which

![Effective Coriolis frequency, \( fe \), for ADCP observations (top) and HYCOM output (bottom). The red line represents passage of hurricane Ivan at 0 IP. Moorings MS10, MS13, and MS14 were utilized for this calculation.](image)

cause upwelling of cold water to create a more stratified MLB that is harder for vertical
shear to overcome (Kunze, 1985; Jaimes and Shay, 2010). ADCP observations indicate ACW rotation (possibly a CCE) in a depth range of 120 to 200 m from 0 to 7 IP, followed by CW rotation (possibly vertically propagating near-inertial waves) from 7 to 20 IP. Vorticity values derived from both ADCP and HYCOM data slightly exceeded the range of shifting frequencies from the carrier frequency analysis (0.75f for slower CW-rotating features to 1.25f for faster ACW-rotating features; Figure 4.12). Coinciding depths of ACW rotation and shifting carrier-frequency values imply that the background vorticity may be modulating the near-inertial carrier frequency. Similarly, HYCOM shows an ACW-rotating feature in the thermocline from 0 to 7 IP but fails to capture the change in rotation after 7 IP. However, the fact that HYCOM maintains the correct background vorticity from 0 to 7 IP (similar rotation direction and frequency range) is encouraging.

Wavelengths and phase speeds of individual modes are shown in Table 4-1. Deriving wavelengths from ADCP current data was difficult as no deep-water moorings had measurements above 50 m depth. Therefore, the OML was not included in the measurements and high-amplitude (>100 cm s⁻¹) motions were lost, impacting near-

<table>
<thead>
<tr>
<th>Mode #</th>
<th>Wavelength (km)</th>
<th>Phase Speed (m s⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>229</td>
<td>2.59</td>
</tr>
<tr>
<td>2</td>
<td>159</td>
<td>1.8</td>
</tr>
<tr>
<td>3</td>
<td>116</td>
<td>1.31</td>
</tr>
<tr>
<td>4</td>
<td>94</td>
<td>1.06</td>
</tr>
<tr>
<td>5</td>
<td>77</td>
<td>0.87</td>
</tr>
</tbody>
</table>

inertial wave resolution. Currents from MS14 were used to calculate wavelength because
it measured the highest amplitude currents near $R_{\text{max}}$. Wavelengths at 50 m, found using the phase shift between MS11 and MS14 averaged over 7 IP, are shown in Figure 4.13, as well as the phase shift in degrees.

From 0 to 7 IP, the wavelength is slightly higher than the predicted first mode wavelength, possibly due to background flow interference or lack of measurements in the OML. This initial response wavelength is closer to the wavelength of 483 km predicted by Geisler (1970) – a product of translation speed and local inertial period. At approximately 2.5 IP and 5.1 IP (baroclinic mode one and two separation times, respectively), the wavelength increases slightly for a short amount of time before decreasing below its initial value. Separation times predicted using Gill (1984) are also supported by ADCP

![Figure 4.13](image.png)

**Figure 4.13** ADCP wavelength; -1 to 9 IP at 50m (top) and phase shift in degrees (bottom). Top pane normalized by Geisler predicted wavelength (483 km). Black lines are mode 1 (left) and mode 2 (right) separation times and red line is mode 1 wavelength (229 km).
and HYCOM current data in Figures 4.3 and 4.4, where wave packets appear to separate into the thermocline at \( t_1 \) and \( t_2 \) based on the separation times calculated in Section 2.6. Mode three is not visible as it separates, probably due to its weaker contribution to the total response. Eventually, by 7 IP, the wavelength approaches 229 km, the first baroclinic mode wavelength. Other studies (Shay and Elsberry, 1987; Brooks, 1983) have shown that near-inertial wavelengths can be smaller than those predicted by linear theory; however the changing wavelength here, seemingly modulated by modal separation scales, is an interesting result in that the wavelength based on phase differences is approaching that of the first baroclinic mode.

4.4 Vertical Shear and Richardson Numbers

HYCOM-derived vertical shear (Figure 4.14) is strongest at 50 to 60 m (0.025 s\(^{-1}\)) and from 0 to 5 IP, supporting the location of the MLB in the background stratification profile. For both ADCP observations and HYCOM simulations (Figure 4.15; in the 50 to 490 m depth range), the spatial and temporal extent of shear is similar and the cross-track velocity shear component dominates (0.005 to 0.010 s\(^{-1}\) at 80 m). Shear over the shelf is larger than the slope shear (0.030 s\(^{-1}\)). Corresponding gradient Richardson numbers, found using the following expression:

\[
Ri \# = \frac{N(z)^2}{\left(\frac{d\text{Vel}}{dz}\right)^2} \tag{4.1}
\]

where \( d\text{Vel}/dz \) is the total velocity shear squared, are less than their critical value of 0.25 at 50 to 60 m and 80 to 100 m from 0 to 2.5 IP, indicating that vigorous vertical mixing occurs in that depth range (Figure 4.16). ADCP observations are all taken below 50 m, so
Figure 4.14 Total vertical velocity shear from HYCOM simulations (s$^{-1}$) in the OML and upper thermocline (20 to 90 m) at MS10. Zero crossings of modes one through three in the same depth interval are overlaid to show how they correspond with shear layers.

the OML was not captured in the ADCP measurements.

Individual modes were least-squares fit to ADCP and HYCOM current data (50 to 490 m depth range) to determine each mode’s role in upper thermocline (80 m) with respect to vertical shear. HYCOM simulations were used to study the OML (10 to 30 m depth range) and strong vertical shear at the deeper MLB (50 to 60 m). Both data sets show that modes two and five control the highly intermittent velocity shear at 80 m in the upper thermocline (Table 4-2) and HYCOM output agrees well with the ADCP measurements. HYCOM-only shear values in the OML (30 m) indicate that modes three and four may have been in control there. At the MLB, mode three is dominant in HYCOM-derived
Figure 4.15 Magnitude of the near-inertial shear from 50 to 490 m (s\(^{-1}\)) at MS10. ADCP data (middle) and HYCOM output (right). Zero crossings of modes one through three in the same depth interval are overlaid to show how they correspond with shear layers.

Figure 4.16 Richardson numbers using all mooring records for the shear profiles from 0 to 2.5 IP. Values are on a log scale and the vertical black line marks 0.25, the requirement for vertical mixing to occur.
Table 4-2 Percentage of total vertical shear variance for individual modes (ADCP and HYCOM).

<table>
<thead>
<tr>
<th>Depth</th>
<th>50 to 490 m</th>
<th>80 m</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>ADCP</td>
<td>HYCOM</td>
</tr>
<tr>
<td>Mode</td>
<td>Mode 2</td>
<td>Mode 3</td>
</tr>
<tr>
<td>Mode 1</td>
<td>1.8</td>
<td>20</td>
</tr>
<tr>
<td>Mode 2</td>
<td>6.5</td>
<td>36.6</td>
</tr>
<tr>
<td>Mode 3</td>
<td>1.3</td>
<td>18.6</td>
</tr>
</tbody>
</table>

shear values (50% of the shear variance). Layers of alternating shear values (positive, negative, positive, etc.) were vertically stacked in both the ADCP and HYCOM analyses. Individual layers of shear appeared to correspond with zero crossings of differing vertical modes that decreased in amplitude with depth. Therefore, zero crossings of modes one through three were overlaid on the shear values to show their possible roles in controlling individual shear layers and mixing (Figures 4.14 and 4.15). Mode three’s zero crossing in Figure 4.14 corresponds with high MLB shear values and mode two’s zero crossing in Figure 4.15 corresponds with upper thermocline shear values (supporting the shear variance analysis). Determining which modes are associated with high shear values and at what depths is important for understanding how hurricane energy is distributed in the ocean interior and, more importantly, how modes evolve in time at various levels according to Gill (1984). Background stratification (which can be modulated by
background vorticity features) must be overcome via vertical shear-induced mixing after storm passage for energetic near-inertial oscillations in the OML to propagate downward.

4.5 Barotropic Response

Changes in SHA fields prior and subsequent to Ivan show interaction with the CCE centered at 25°N/87°W and the WCE at 27°N/88°W, and a general decrease in total SSH fields following Ivan’s track north (Figure 3.1). Differencing the before and after SHA measurements shows a height change of approximately 15 cm in the ADCP moorings’ vicinity and intensification of the second CCE (centered on 27.5°N/87°W ) by up to 45 cm s⁻¹ (Figure 4.17). A height change of 15 cm agrees well with the results of Shay and Chang (1997) found as part of the barotropic trough in the sea surface height. Relative vorticity calculated from the SHA differencing showed an ACW-rotating trough in Ivan’s wake (Figure 4.18). HYCOM surface height fields (Figure 4.19) decrease 15 to 17 cm in the mooring area after Ivan’s passage and show intensification of the CCE (Figure 4.20). Height changes were not so cleanly structured as the modeled heights in Shay and Chang (1997), which showed undulations of 3 to 7 cm at 1 IP following the hurricane track. Background features such as CCEs, WCEs, and bottom topography, which were ignored in those highly idealized simulations, will complicate the picture.

Depth-averaged mass divergence calculated from HYCOM output oscillates sinusoidally at 1 IP within the first 6 IP of Ivan’s passage as manifested in the free surface oscillations. This motion agrees well with the results of Shay and Chang (1997), who used a 17-layer model to generate a mass divergence-driven barotropic response from hurricane Frederic. One disagreement is the oscillation amplitude (1.2x10⁻⁵ kg m⁻³ s⁻¹), which is two orders of magnitude higher than Shay and Chang’s result. One likely
Figure 4.17 Change in altimetry-derived SSH (cm) due to Ivan’s passage i.e. difference between satellite data from September 13 and 16, 2004. Geostrophic current vectors (derived from the change in surface heights) are in the range of -0.08 to 1.20 m s\(^{-1}\).

Figure 4.18 Relative vorticity (s\(^{-1}\)) calculated from SHA difference in Figure 4.17. Vectors are geostrophic velocity.
Figure 4.19 HYCOM-derived SSH (m) several hours before (top) and after (bottom) Ivan’s passage.

Figure 4.20 Surface height change in HYCOM SSH field (m) between 0000 UTC September 15, 2004 and 1200 UTC September 16, 2004 (before and after Ivan). Geostrophic current vectors are in the range of -0.35 to 0.35 m s$^{-1}$. 
reason for the difference is that Shay and Chang’s model is initialized in a quiescent ocean and of much simpler construction compared to the more sophisticated HYCOM, which includes bottom topography. Depth-averaged mass divergence is calculated in an area of CCW-rotating surface geostrophic currents left behind in Ivan’s wake (Figure 4.21), further supporting the effects of the free surface on the barotropic response.

Aanderaa current measurements also show a distinctive response at approximately 950
Figure 4.22 Depth-averaged zonal near-inertial current (red) at MS14 and Aanderaa zonal near-inertial current (blue) at the same site at 950 m.

Figure 4.23 Comparison of depth-averaged currents from HYCOM and ADCP over a relatively flat bottom (MS07; top) and a sloped bottom (MS10; bottom). m depth which is similar in amplitude to the depth averaged ADCP near-inertial response (Figure 4.22). Bottom currents oscillate at a similar frequency to the depth-averaged currents, implying that the barotropic response is felt at depth, during the storm. The response also persists for several IP following storm passage in both measurements, due
principally to the interaction of the more energetic baroclinic modes and the depth independent response, as opposed to immediately propagating away. A comparison between the HYCOM and ADCP depth-averaged currents shows good agreement at MS07, but decreasing similarity eastward to MS10 (Figure 4.23). The ADCP barotropic response is 8 cm s\(^{-1}\) and the HYCOM barotropic response is 6 cm s\(^{-1}\). Thus, there is agreement both observationally and numerically that Ivan excited a barotropic response.

### 4.6 Comparison of HYCOM to Observations

ADCP data and HYCOM output describe the same general features at similar depths and time periods, however, striking differences appear upon closer inspection (for reference, all differences involve subtracting HYCOM output from ADCP data).

**Figure 4.24** Difference between ADCP and HYCOM velocity profiles at a) MS09 and b) MS12. Differences in speed and direction are shown.
Differencing ADCP and HYCOM stick plots at MS09 and MS12 shows that ADCP currents are much smaller than and out of phase with HYCOM currents at the MLB. At lower depths, ADCP currents are larger by up to 20 cm s\(^{-1}\) (Figure 4.24). A CCE is present from 120 to 200 m depths and 2 to 5 IP in both sets of carrier frequencies; however it is stronger in the HYCOM output. ADCP carrier frequencies are about 0.05\(f\) lower than HYCOM frequencies from 200 to 300 m depths and 0.05\(f\) higher from 300 to 500 m depths (Figure 4.25). However the uncertainty with a one-hour sampling period is +/- 0.08\(f\) for the ADCP instruments.

Compared to the Price (1983) scaling parameters, ADCP thermocline (> 100 m) kinetic energy at MS07 and MS08 is five times the predicted value, and instead approaches the OML kinetic energy value. ADCP data at MS09 and MS10 is more

**Figure 4.25** ADCP and HYCOM carrier frequencies from 2 to 5 IP, background inertial frequency (black) and Price predicted near-inertial frequency (blue).
Figure 4.26 Difference between ADCP and HYCOM CW kinetic energy at MS07 to MS10.

Figure 4.27 A closer look at MS07 and MS08 near-inertial current velocities. Thermocline velocities reach 98 cm s$^{-1}$ at MS07 and 93 cm s$^{-1}$ at MS08.
Table 4-3 Linear regression $R^2$ and $R$ values and lag between ADCP and HYCOM total shear at 80 m, 0 to 3 IP

<table>
<thead>
<tr>
<th></th>
<th>MS07</th>
<th>MS08</th>
<th>MS09</th>
<th>MS10</th>
</tr>
</thead>
<tbody>
<tr>
<td>$R^2$</td>
<td>0.6773</td>
<td>0.3783</td>
<td>0.4737</td>
<td>0.4089</td>
</tr>
<tr>
<td>$R$</td>
<td>0.8230</td>
<td>0.6151</td>
<td>0.6883</td>
<td>0.6395</td>
</tr>
<tr>
<td>Lag (h)</td>
<td>0</td>
<td>0</td>
<td>1</td>
<td>1</td>
</tr>
</tbody>
</table>

similar to the Price value, and HYCOM thermocline kinetic energy is only twice as much as the predicted value at all moorings. ADCP and HYCOM energies at the OML base (50 to 80 m) are much larger than the Price thermocline prediction, likely due to OML deepening, which allows stronger currents to penetrate the seasonal thermocline. Differencing HYCOM and ADCP CW kinetic energies reveals that the ADCP moorings have higher energy in the thermocline at MS07 and MS08 and lower energy at the MLB at MS09 and MS10 (Figure 4.26).

Large ADCP kinetic energy at MS07 and MS08 can be explained by taking a closer look at the ADCP-measured near-inertial velocity for these moorings. At 172 m and 2 IP, the total near-inertial current velocity at MS07 is 98 cm s$^{-1}$, enough to equal twice the Price thermocline prediction (45 cm s$^{-1}$). Similarly, at MS08 near-inertial current velocities reach up to 93 cm s$^{-1}$ at 192 m and 1 IP (Figure 4.27). At MS07 and MS08, thermocline energy levels approach those of the OML 1 to 2 IP after storm passage, which HYCOM fails to show.

The main difference between ADCP and HYCOM total vertical shear from 50 to 490 m is more intense values in the ADCP measurements from 70 to 100 m, 6 to 11 IP (Figure 4.28). Stronger shear values from the initial storm impulse observed in the ADCP
data from 0 to 3 IP, albeit not as different from HYCOM output as shear values from 6 to 11 IP. Cross-correlation was performed using ADCP and HYCOM maximum shear at 80 m from 0 to 3 IP. Lag values from the cross correlation were then used to perform a linear regression between the two data sets. $R^2$, $R$, and lag values are noted in Table 4-3. At MS07, no lag exists and the regression analysis shows an $R$-value of 0.82; however at MS10, the lag is one hour and regression analysis shows a value of 0.64, possibly due to the ADCP sampling interval.

**Figure 4.28** Difference in the vertical current shear between ADCP and HYCOM ($s^{-1}$) at MS10.
Chapter 5: Summary and Future Work

Continuous ADCP observations throughout the water column made possible a unique opportunity to resolve the baroclinic response in the near-inertial band during and subsequent to a major hurricane in the GOM. Within the context of individual contributions to shear and energy propagation, each mode’s spatial and temporal development was visible and its wave characteristics determinable, as well as possible impacts from background vorticity field associated with warm and cold core ocean features. The extent of coverage meant that observations could be compared to HYCOM output data for model evaluation. Several scientific issues were identified and isolated, especially in these difficult-to-resolve coastal waters with highly variable stratification over steep bottom terrain. Consistent with previous studies, a deep barotropic response was observed at depth while the ocean surface was depressed up to 17 cm after hurricane passage. In this framework, HYCOM simulations were an excellent tool for analyzing this barotropic response in more detail, particularly the depth-averaged mass divergence, which was not possible to derive from ADCP moorings due to lack of concurrent background stratification measurements (e.g. temperature and salinity).

5.1 Baroclinic Response

Three viewpoints exist in regards to the nature of the near-inertial response: that it should be modeled as an infinite number of standing modes; it should be modeled as propagating waves; it is a mixture of the two phenomena. Analyses are completed in this study under the assumption that the response is a mixture by evaluating individual modes and determining wave characteristics at the same time. Evolving modal amplitudes are shown via least-squares fits. As the quantitative results have been discussed above, an
interpretation of several figures and synthesis of the results’ meaning will take place here.

Velocity profiles show a predominant zero crossing in the thermocline, indicating the presence of an energetic first baroclinic mode. Profiles also show movement of possible wave packets downward from the OML and rotary kinetic energy analysis suggests that downward-propagating energy erupts suddenly into the thermocline (Figure 4.9). This energy burst may be due to a tipping point in which the buildup of vertical shear is finally large enough to overcome stratification at the OML base (e.g. shear-induced instability) and allow energy to escape downward. Zero crossings of standing modes may cause the high vertical shear necessary for these intermittent bursts of energy and seem to coincide with the levels of highest shear. Shear variance percentages imply that mode three is responsible for shear at the OML base and mode two is responsible for shear in the upper thermocline.

Total rotary kinetic energy was calculated for MS07 to MS10 in both the ADCP and HYCOM data sets (Figure 4.9). A large amount of CW-rotating kinetic energy in the thermocline at MS07 and MS08 (almost as high as OML kinetic energy) was captured by the ADCP data and indicates immediate energy transfer from the OML at those two moorings. HYCOM failed to capture the high thermocline kinetic energy values and instead maintained similar values at all moorings. Observations show that a much smaller amount of kinetic energy remaining in the OML at MS09 and MS10, possibly indicating stronger stratification in that area, or some other feature that prevented energy from escaping downward. However, it should be noted typical e-folding scales of 3 to 5 IP have previously been observed in the OMLs forced by hurricanes of similar strength (Brooks, 1983; Shay and Elsberry, 1987)
Strong background vorticity features were present near MS09 and MS10 during and after storm passage, possibly leading to changes in stratification and modulation of near-inertial carrier frequencies and wavelengths (Figure 5.1). On the moorings’ west side (near MS07), however, strong vorticity features were not present and effective Coriolis frequencies only ranged from 0.93 to 1.08$f$. Downward energy propagation was larger for moorings on the west side (MS07 and MS08) than on the east side (MS09 and MS10). Based on previous studies, this result is counterintuitive, although changing vorticity fields associated with oceanic features and their horizontal gradients may explain why less energy was able to leave the OML at MS10: stronger stratification was created by stronger ACW rotating (more positive) vorticity features, creating an obstacle to near-inertial motions as they attempted to propagate into the thermocline. This assertion must be supported, however, by a comparison of stratification values at MS07 and MS10. Such values are only available from HYCOM, as they were not recorded during Ivan.

Density calculated from HYCOM temperature and salinity output shows strong upwelling under the storm, followed by oscillations at the MLB with periods of 1 IP. After 5 to 10 IP, the MLB reaches an equilibrium point at 80 m depth and the density approaches an equilibrium value as well (Figure 5.2). The result suggests that the process of OML deepening occurs over a week or more (consistent with previous studies), and the effects last for a similar time period (e-folding scales of 3 to 5 IP). The MLB oscillates with 20 to 30 m displacement over 12 h. At lower depths, very little change in density occurs, although some small oscillations are present, albeit not strong enough to lower Richardson numbers to below criticality and cause mixing.
At first glance, MS07 and MS10 have very similar density values in the five IPs leading up to hurricane Ivan. However, when the profiles are differenced MS07 has a density at the MLB (30 m) that is 0.5 kg m$^{-3}$ less than that of MS10 (Figure 5.3). A lower density value at MS07 indicates weaker stratification and a more favorable environment for vertical energy propagation and shear induced mixing. This stratification change corresponds to changing background vorticity values from MS07 to MS10. Below 30 m
Figure 5.2 HYCOM density fields (kg m\(^{-3}\)) at MS07 (top) and MS10 (bottom).

differences vary. However, the MLB is the first layer to be overcome by near-inertial depth and after Ivan’s passage, the picture becomes more complicated and these motions as they propagate into the ocean interior and significant vertical mixing occurs. Thus, stratification changes there are the most important as the ocean cools and deepens.
Modes three and four were responsible for vertical shear at the MLB (30 m) according to HYCOM output, and modes two and five were responsible for vertical shear in the upper thermocline (80 m) according to both data sets (Table 4-2). There is good agreement between ADCP and HYCOM data between 50 to 490 m depths. Zero crossings of these two modes may be responsible for breaking down stratification and allowing vertical energy transfer to occur. Due to the important role shear plays at the MLB, care should be taken to observe current velocities and stratification all depths in future studies.

Figure 5.3 Difference in HYCOM density fields between MS07 and MS10 (kg m\(^{-3}\)). Negative values imply lower pre-storm stratification at MS07 and vice versa.
5.2 Deep Barotropic Response

Not often observed at lower depths, the barotropic response occurred here throughout the water column after storm passage. Many early theoretical studies have ignored this effect by assuming it is not part of the near-inertial response to hurricane forcing. Notwithstanding, it tends to be shorter in duration with a weaker magnitude than the baroclinic response. It is likely transmitted into the ocean interior quickly by virtue of hydrostatic dynamics via a free surface boundary condition that pressure is continuous across the air-sea interface (Shay and Chang, 1997). Surface height anomalies from strong storm winds are balanced by the vertically averaged mass divergence which represents interaction between the evolving baroclinic response associated with the upwelling-downwelling process and the free surface. As noted, the density profiles were not available at the ADCP sites during storm passage, so HYCOM temperature and salinity output were used along with model current velocity data to calculate depth-averaged mass divergence.

Sinusoidal oscillations of approximately one IP in the depth-averaged mass divergence indicated the presence of a near-inertial barotropic mode in the HYCOM output data. Shay and Chang (1997) modeled a similar response that was two orders of magnitude lower than that recorded here. That is, the barotropic mode and more energetic baroclinic modes interact over near-inertial time scales suggesting the barotropic mode cannot be ignored by scaling arguments. Unlike the Shay and Chang model, which was initialized in a quiescent, flat bottom ocean, HYCOM was initialized with more realistic parameters, a difference which may explain in part the change in magnitude. In terms of observed features, satellite-derived surface height anomalies were within the expected
range (10 to 20 cm), depth-averaged ADCP currents oscillated at 1 IP, and the Aanderaa current measured a near-inertial oscillation at 950 m depth as well. HYCOM depth-averaged currents agreed well with those from the ADCP moorings, although less similarity exists toward the right side of the storm (Figure 4.23). The dissimilarity may be due to different bottom topography (a steeper slope at MS10) or poorly constrained background stratification, which led to an erroneous prediction of background vorticity. Regardless, in this case the model and observations both support the existence of a barotropic response due to a changing free surface.

One interesting feature in the satellite observations is the small CCE centered at 27.5N/87°W, which intensifies by up to 45 cm s⁻¹ after Ivan’s passage (Figures 4.17 and 4.18). Such a large intensification was unexpected and occurred under the storm’s right side, which has the highest wind stress. Intensification likely occurs due to eddy interaction with the barotropic free surface trough in the storm’s wake, which extends north to the moorings. HYCOM captures the increased eddy currents, albeit about one degree west of where the actual intensification occurred.

5.3 HYCOM and ADCP Comparison

Both data sets returned similar results in most of the diagnostic analyses discussed above. However, HYCOM tends to overestimate baroclinic velocities and kinetic energy by a small amount and it underestimates the barotropic current. Spatial and temporal variations are mostly the same for both data sets and derived wave and modal characteristics are similar. However, differences in carrier frequency exist, particularly in HYCOM’s depiction of a stronger CCE from 120 to 200 m but less extreme frequency shifts at lower depths. Background vorticity features are captured well by HYCOM for
the first 7 IP, but model output quality deteriorates with time, likely due to poor altimetry data availability close to the coast and perhaps riverine output climatology that did not match current conditions in the area. Vorticity values at the bottom are particularly prone to error. Comparison of HYCOM to ADCP shear and barotropic currents suggests the model response is slightly out of phase (lag of one hour) with the observed currents. HYCOM also does not capture the intensification of a small CCE in its exact location (see Section 4.5). Altimetry and stratification data are required by the model to initialize background temperature and salinity properly, so improving data coverage and quality near the coast would give better results. Bottom topography is highly variable in the DeSoto Canyon area, another caveat that may explain discrepancies in the model output, particularly since the barotropic response is sensitive to bottom topography and HYCOM carrier frequencies and background vorticity values are less reliable at lower depths. Overall, however, HYCOM captures the essential near-inertial features left behind by Ivan and seems a trustworthy ocean model in this case (as suggested in the Halliwell et al. (2011) study).
Chapter 6: The Big Picture

Hurricanes inject energy into the ocean depths, changing the near-inertial energy budget on a global scale. Weathered oil can resurface due to hurricane-induced upwelling and may affect coastal communities, particularly those north of the Deepwater Horizon oil spill. Strong currents at the ocean bottom scour massive amounts of sediment and damage oil drilling equipment (Marks and Shay, 1998). Each of these facts is motivation to understand how and how much storm energy enters the ocean and where it ends up. Results here indicate that background stratification, as determined by Loop Current eddies and riverine outflow, must be overcome for baroclinic energy to reach the ocean interior. Each stratified layer is a barrier to be breached via shear instabilities at the layer base, a process which may coincide with zero crossings of standing modes to provide opposing velocities with depth. Background vorticity may not be constant with depth, so energy moves vertically at different rates as it encounters changing stratification. Once a layer is overcome, energy bursts through and propagates downward to the next layer like a wave, where it works to eliminate the next barrier.

A barotropic impulse caused by changes in the free surface precedes baroclinic energy propagation. Constant with depth, the impulse provides an immediate energy source throughout the water column. Although barotropic amplitudes (8 cm s$^{-1}$) may not compare with baroclinic amplitudes in the OML (up to 150 cm s$^{-1}$), they rival baroclinic amplitudes at the ocean bottom and reach great depths sooner, making them important for bottom transport and storm energy redistribution. In terms of vertically integrated transports, barotropic and baroclinic components may actually be nearly equal and must be considered in regimes such as the DeSoto Canyon where bottom slopes are steep.
Moving forward, modelers and experimentalists must consider the importance of real-time temperature and salinity measurements to correctly represent the background stratification. Documenting how stratification changes with time will provide a better idea of how much hurricane energy reaches the ocean interior. Acknowledging the importance of the barotropic impulse at all depths is also important due to its immediate impacts at depth (e.g. Aanderaa current meters at 950 m). Most of all, the view that these processes are controlled by a mixture of standing modes and propagating waves will allow for a more comprehensive view and flexibility in modeling and observing them. Pathways of weathered oil plumes, changes in global energy budgets, and damage to drilling rigs might be predicted with more accuracy if near-inertial motions are better understood in observations and better represented in oceanic models used in coupled systems. In addition oceanic models require improved initialization schemes containing kinematics and dynamics of energetic ocean features such as fronts and eddies as they impact the oceanic response to hurricane forcing. Such a model must then be properly coupled with the atmosphere to improve intensity forecasts (Marks and Shay, 1998).
References


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