Mean and Near-Inertial Ocean Current Response to Hurricane Gilbert

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ABSTRACT

The three-dimensional hurricane-induced ocean response is determined from velocity and temperature profiles acquired in the western Gulf of Mexico between 14 and 19 September 1988 during the passage of Hurricane Gilbert. The asymmetric wind structure of Gilbert indicated a wind stress of 4.2 N m$^{-2}$ at a radius of maximum winds ($R_{\text{max}}$) of 60 km. Using observed temperature profiles and climatological temperature–salinity relationships, the background and storm-induced geostrophic currents (re: 750 m) were 0.1 m s$^{-1}$ and 0.2 m s$^{-1}$, respectively. A Loop Current warm core ring (LCWCR) was also located to the right of the storm track at 4–5 $R_{\text{max}}$, where anticyclonically rotating near-surface and 100-m currents decreased from 0.9 m s$^{-1}$ to 0.6 m s$^{-1}$ at depth. The relative vorticity in the LCWCR was shifted below the local Coriolis parameter by about 6%.

In a storm-based coordinate system, alongtrack residual velocity profiles from 0 to 4 $R_{\text{max}}$ were fit to a dynamical model by least squares to isolate the near-inertial content over an $\epsilon$-folding timescale of four inertial periods ($\text{IP} = 30$ h). Observed frequency shifts in the mixed layer currents ranged from 1.03 to 1.05 $f$ in agreement with both the backrotated velocity profiles at 1.04$f$ relative to the storm profile (where maximum correlation coefficients were 0.8) and the predicted frequency shift from the mixed-layer Burger number. This frequency was increasingly blue shifted in the upper 100 m to 1.1$f$, decreasing toward $f$ within the thermocline. Near-inertial currents rotated anticyclonically by 90°±180° in the upper ocean, providing the velocity shear for layer cooling and deepening observed on the right-hand side of the track. A summation of the first four baroclinic modes described up to 77% of this near-inertial current variability during the first 1.75 IP. However, the variance explained by this modal summation decreased to a minimum of 36% after 2.9 IP following passage due to phase separation between the first baroclinic mode and higher-order modes in the mixed layer. Although the response was complicated by the LCWCR, the evolving three-dimensional current structure can be described by linear, near-inertial wave dynamics.

1. Introduction

Strong atmospheric forcing events, such as tropical cyclones and the upper-ocean response, represent one of the more extreme examples of mesoscale ocean–atmosphere interaction. Previous observational studies of the upper ocean’s momentum response to tropical cyclones have been limited to fortuitous encounters with moored instrumentation deployed in support of experimental programs with other objectives. Although these measurements have provided insight into the dominant evolution of the response at fixed vertical levels (e.g., Shay and Elsberry 1987), the three-dimensional evolution of the near-inertial wave wake has only been inferred from numerical and analytical simulations for assumed quiescent ocean conditions. In the presence of background flows, the excitation and spreading of the near-inertial wave wake requires a framework to understand the observed evolutionary phases of the response.

For the first time, the three-dimensional momentum response in the directly forced regime of a hurricane was documented from a series of airborne expendable current profiles (AXCPs) successfully deployed in Hurricanes Josephine (1984), Norbert (1984), and Gloria (1985) (Sanford et al. 1987). Since only one snapshot of the current and temperature structure was acquired in these storms, the interpretation of the evolving near-inertial response depended on analytical study where the baroclinic, hurricane-induced velocity structure was simulated based on an idealized hurricane (Shay et al. 1989). They found that over 70% of the ocean current variability in the directly forced region of Norbert (1984) could be ascribed to a sum of the first few baroclinic modes. In a detailed numerical study of Norbert, Price et al. (1994) found good agreement between modeled and observed velocity profiles for the case of weak background flows. By contrast, the North Atlantic subtropical front altered the response as inferred from drifter trajectories in Josephine and Gloria (Black et al. 1988).
In an earlier study, Weller (1982) found a marked difference between the amplitudes and phases of the mixed layer near-inertial currents from two surface moorings due to vorticity and divergence associated with the background current structure. Mooers (1975) and Kunze and Sanford (1984) demonstrated that near-inertial motions are modulated by interactions with current shear associated with baroclinic fronts, jets, and eddies. Kunze (1985) subsequently derived a dispersion relation based on a wave-mean flow interaction model to predict near-inertial wave propagation in geostrophic current shear. The frequency of near-inertial waves is shifted above (below) the effective Coriolis frequency in regions of positive (negative) vorticity of the low-frequency motions. Based upon the WKBJ approximation, low-frequency motions should have a minimal impact on forced near-inertial motions because of their long wavelengths compared to the scale of the background geostrophic vorticity fields, particularly with respect to the effective Coriolis frequency (D'Asaro 1995). The interactions between forced near-inertial motions and the geostrophic flow field have not been understood during hurricane passage even in a qualitative sense.

Near-inertial currents forced by tropical cyclones are characterized by low-order, forced baroclinic modes (Geisler 1970; Shay et al. 1989). There is a characteristic timescale for the phase of each baroclinic mode to separate from the wind-forced mixed layer when the horizontal scale of the wind stress (typically 60 km) is greater than the deformation radius associated with the first baroclinic mode \([O(40 \text{ km})]\) (Gill 1984). This baroclinic timescale required for a phase difference of \(\pi/2\) to develop for each mode \(n\) is given by

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

where \(k\) is the horizontal wavenumber of the wind stress, \(f\) is the Coriolis parameter, and \(c_n\) is the phase speed of the \(n\)th baroclinic mode. These predicted timescales for modal separation increase with mode number due to decreasing modal phase speeds, and the timescales may also represent the onset of rapid energy propagation from the mixed layer into the thermocline (Zervakis and Levine 1995).

Near-inertial current shears across the mixed layer base drive the bulk Richardson number to below critical values, causing the mixed layer to deepen and cool (Price 1981). Near-inertial motions rotate anticyclonically with depth by \(90^\circ-180^\circ\) over \([100 \text{ m}]\) vertical scales and are associated with high-order, short vertical wavelength baroclinic modes (Niiler and Kraus 1977). A kinematical description of the evolution of such three-dimensional processes has remained unresolved due to the lack of high-resolution observations acquired during hurricanes.

An experiment was conducted in the western Gulf of Mexico from NOAA WP-3D research aircraft by successfully deploying 76 AXCPs and 51 airborne expendable bathythermographs (AXBTs) during the passage of Gilbert from 14 to 19 September 1988 (Shay et al. 1992). This experimental effort measured the current and temperature structure prior (Prestorm), during (Storm), one (Wake 1), and three (Wake 2) days following Gilbert’s passage (Fig. 1). Prior to these airborne experiments, the trajectory of a drifter north of the storm track indicated the presence of a Loop Current warm core ring, designated F, (LCWCR F). Thus, the objective here is to gain an understanding of the role of this strong baroclinic structure on the near-inertial response to Gilbert. To achieve this objective, temperature and salinity relationships from historical data and observed temperature profiles are used to estimate the densities, buoyancy frequency profiles, and the geostrophic flows at each profiler site (section 3). Upon removal of the geostrophic currents (section 4), the near-inertial response is isolated by fitting alongtrack residual profiles to a near-inertial model using the approach of Rossby and Sanford (1976). Mixed layer current response, the evolving current shears across the entrainment zone, and the impact of vorticity upon the forced near-inertial motions are described within the context of these analyzed results. Simulated profiles from an analytical model (Shay et al. 1989) based on summing the first few forced baroclinic modes are compared to near-inertial current profiles in section 5, followed by a summary and concluding remarks (section 6).

2. Observations

The primary dataset was acquired during five flights to examine the hurricane-induced ocean response, designated as Prestorm, Storm (two flights), Wake 1, and Wake 2, consisting of (31, 0), (20, 17), (0, 31), and (0, 28) (AXBTs, AXCPs) for these experimental periods, respectively (Shay et al. 1992) (Fig. 1). These data are complemented by measurements from the Minerals Management Service (MMS) Gulf of Mexico Physical Oceanography Program (SAIC 1989), including expendable bathythermographs (XBTs), conductivity–temperature–depth (CTDs), and AXBTs deployed from research vessels and aircraft six weeks prior (yearday 200) and four weeks (yearday 300) following Gilbert. Moored current meter measurements along 92ºW are also used to examine the response at several vertical levels.

a. Velocity and temperature profiles

Velocity and temperature profiles at 3-m intervals were processed following the procedures described in Sanford et al. (1982), and the times, positions, and failure rates of the profilers have been reported elsewhere [Tables 3, 4, and 5 in Shay et al. (1992)]. In addition to the strong baroclinic velocity signals, these velocity
Fig. 1. Locations of AXCPs (circles) and AXBTs (boxes), tracks of SAIC drifting buoys (dashed), NDBC buoys (triangles), and MMS moorings EE, FF, and GG (open circles in panel a) relative to the track of Hurricane Gilbert (solid) in the Gulf of Mexico from (a) 14 Sep, (b) 16 Sep, (c) 17 Sep, and (d) 19 Sep 1988. The 200-m bottom contour is dotted (from Shay et al. 1992).

profilers detected surface-wave-induced orbital velocities in the mixed layer that were estimated by fitting a three-layer model of Sanford et al. (1987) to the velocity profiles. These higher-frequency motions were removed from the current profiles (referred to as observed current profiles).

The Wake 1 experiment occurred between 1 to 1.7 inertial periods (IP) following passage and encompassed the second upwelling event (Fig. 2). Over this time interval, upper-ocean currents with magnitudes of 1–1.2 m s\(^{-1}\) indicated a divergence of the flow away from the storm track, which induced an upwelling of cooler water along Gilbert’s track. The observed current vectors rotated anticyclonically in the upper layers when viewing the currents from the western to the eastern part of the experimental domain. This predominant anticyclonic rotation is indicative of strongly forced near-inertial motions. On the right-hand side of the storm track, warmer temperatures penetrated deeper into the water column in the LCWCR F in comparison to the hurricane-induced mixed-layer cooling of 3.5°–4°C along the track, in accord with the conceptual model of Black (1983). This observed temperature pattern indicated a wavelike structure over periods of the same order as the local inertial period, which is consistent with strong near-inertial motions (Shay et al. 1992).
b. Hurricane Gilbert

Gilbert’s winds weakened as it moved into the Gulf of Mexico in a northwestward direction at an average speed of 5.6 ± 0.3 m s⁻¹. Central pressures were in the 940-mb range and maximum winds were 50–55 m s⁻¹ at the primary radius of maximum winds ($R_{\text{max}}$) of 60 km. There was a secondary wind maximum at 90 km (1.5$R_{\text{max}}$) due to a contracting eyewall (Black and Willoughby 1992). These variations caused uncertainties in exactly defining $R_{\text{max}}$ since it decreased from about 60 km just off the Yucatan Peninsula to 45 km prior to landfall in Mexico on 17 September.

Wind stress vectors (Fig. 3) at 10-km intervals were constructed from the objectively analyzed flight-level winds reduced to 10 m using a boundary layer model (Powell 1980) and the Large and Pond (1981) drag coefficient formulation. Maximum wind stress was 4.2 N m⁻² at $R_{\text{max}}$ and 3.3 N m⁻² at 1.5$R_{\text{max}}$, respectively, where the estimated friction velocity ($u^*_{\text{f}}$) was 3.5–4 m² s⁻² (Fig. 3b). The stress divergence field (Fig. 3c) indicated maximum values of $-0.9 \times 10^{-4}$ s⁻¹ just in front of the storm and more typical values of $-0.3 \times 10^{-4}$ s⁻¹ elsewhere. The strongest wind stress curl in the cyclonically rotating vortex extended over $\pm 2 R_{\text{max}}$ with a maximum of $1.7 \times 10^{-4}$ s⁻¹ at the center and values of $0.4 \times 10^{-4}$ s⁻¹ at 1.5$R_{\text{max}}$ (Fig. 3d), suggesting an asymmetry in Gilbert’s wind stress field.

c. Storm coordinate system and air–sea variables

The oceanic wavelength of the response induced by a moving tropical cyclone is proportional to the product of the storm translation speed ($U_h$) and the local inertial period (IP) (Geisler 1970). Based upon a 5.6 m s⁻¹ translation speed and an IP of 30 h, the predicted wavelength ($\lambda$) is estimated to be 586 km with an uncertainty of
FIG. 3. (a) Wind stress vectors (dyn cm⁻²), (b) surface friction velocity (m² s⁻²), (c) wind stress divergence (×10⁻⁶ s⁻¹), and (d) wind stress curl (×10⁻⁶ s⁻¹) from flight-level winds on 15–16 Sep 1988 in the Gulf of Mexico in the vicinity of the AXCP and AXBT measurements (courtesy of NOAA's HRD).

±30 km (Table 1). Velocity profiles were placed into a storm coordinate system relative to 0600 UTC 16 September at 22°57'N and 94°44'W, allowing conversion of alongtrack distance to time. These storm-based profiler positions in the along- and cross-track directions were then nondimensionalized in terms of the inertial wavelength (Λ) and the radius of maximum winds (Rmax) in Shay et al. (1992).

Price (1983) defined a mixed-layer Burger number as

\[ M = \frac{(1 + S^{-2})g' h}{(2R_{\text{max}} f)^2}, \]  

where \( S \) is the nondimensional storm speed \( U_s/(2R_{\text{max}} f) \), \( h \) is the mixed layer depth, and \( g' \) is reduced gravity. The thermocline Burger number \( (T = h^{-1} bM) \), where \( b \) is the thermocline thickness, is a factor of 5 larger than the mixed layer Burger number \( (M = 0.08) \) or approximately 0.4. Both Burger numbers represent the dynamical coupling between the pressure gradients in the two layers. The blue shift in the mixed-layer inertial frequency is proportional to \( M/2 \) (Table 1), equating to a near-inertial frequency of 1.04\( f \). Key air–sea variables and Price (1983) scaling arguments were used to place the observations into a nondimensional framework. Isopycnal displacements (\( \eta \)) scale as \( \tau_{\text{max}}/(\rho_f f U_s) \) or about 13 m, and the geostrophic speeds \( (V_g) \) are proportional to \( g' \eta (f R_{\text{max}}) \) or 0.11 m s⁻¹. The
Table 1. Air–sea parameters, nondimensional numbers, and scales in Hurricane Gilbert based on scaling arguments of Price (1983). Note that the maximum stress represents the symmetric part of the wind stress field based on a fit to a Rankine vortex for the purpose of scaling.

<table>
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<th>Parameter</th>
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<td>Radius of maximum winds (km)</td>
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<tr>
<td>Maximum wind stress (N m^{-2})</td>
<td>\tau_{max}</td>
</tr>
<tr>
<td>Speed of the hurricane (m s^{-1})</td>
<td>U_h</td>
</tr>
<tr>
<td>Wavelength (km)</td>
<td>\lambda</td>
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<td>First mode phase speed (m s^{-1})</td>
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<tr>
<td>First mode deformation radius (km)</td>
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<td>Inertial period (d)</td>
<td>IP</td>
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<tr>
<td>Reduced gravity (m s^{-2})</td>
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<td>Mixed layer depth (m)</td>
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Nondimensional numbers

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<td>Nondimensional storm speed (S)</td>
<td>(U_h/2R_{max})</td>
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<tr>
<td>Mixed layer Burger number (M)</td>
<td>(1 + S^{-2})g' h/(2R_{max})^2</td>
</tr>
<tr>
<td>Thermocline Burger number (T)</td>
<td>(h/b)M</td>
</tr>
<tr>
<td>Nondimensional forcing (Fo)</td>
<td>(2R_{max}/c_1)</td>
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Scales

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<tr>
<td>Wind-driven velocity (V_is) (m s^{-1})</td>
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</tr>
<tr>
<td>Thermocline velocity (V_{th}) (m s^{-1})</td>
<td>(h/b)V_{th}</td>
</tr>
<tr>
<td>Isopycnal displacements (\eta) (m)</td>
<td>\tau_{max} / (f R_{max})</td>
</tr>
<tr>
<td>Geostrophic velocity (V_{gs}) (cm s^{-1})</td>
<td>g' \eta / (f R_{max})</td>
</tr>
<tr>
<td>Frequency shift (\epsilon)</td>
<td>M/2</td>
</tr>
</tbody>
</table>

wind-driven ocean velocity (V_is) scales as \( \tau_{max} R_{max} / (\rho h U_h) \), representing a speed of 1.07 m s^{-1}, while the predicted thermocline (V_{th}) speed is 0.22 m s^{-1} based on the expression \( V_{th} h/b \).

d. Temperature and salinity structure

Historical temperature–salinity (T–S) relationships are needed to estimate the density, dynamic height, and geostrophic flow from observed temperature profiles (see the appendix). These data consisted of XBTs and AXBTs from yeardays 187–217 and yeardays 293–296, respectively; CTD profiles from yeardays 295–309; and CTD profiles from 1989 yeardays 25–35 (SAIC 1989). Given the one month difference between the 1988 CTD dataset and Gilbert observations, and the 1-yr decay timescale for LCWCWRs (Elliot 1982), the inferred T–S relationships were representative of these fields for the Gilbert experiments.

In the western Gulf of Mexico, the two dominant T–S curves (Fig. 4) correspond to Gulf common water and Loop Current water (Wust 1964). The T–S relationship for Gulf common water spans a narrow range of salinities for tem-

![Fig. 4. Temperature and salinity diagrams for the (a) gulf common water and (b) LCWCWR water in the western Gulf of Mexico. Spline fits are given as solid lines.](image-url)
peratures exceeding 18°C. Temperatures decrease linearly by 1°C per 1.4 parts per thousand (ppt) change below the 18°C Water. Based upon water mass definitions (Wust 1964; Nowlin 1972), Elliot (1982) characterized Loop Current ring water as high-salinity subtropical water from the Caribbean (S = 36.6 psu at 22°C) surrounded by gulf common water (S = 36.4 psu at 22°C). The salinity maximum of LCWCRs is 36.70 psu (Fig. 4b) occurring at a temperature of 22°C (Cooper et al. 1990). This high saline water is absent at temperatures below 18°C where the T–S relationships between the two water masses converge.

e. Baroclinic modes

A representative profile from the western Gulf of Mexico was used to determine the buoyancy frequency profile (Fig. 5). A maximum frequency of 12 cph was located beneath the mixed layer of 40 m. Below this maximum, a region of frequencies >3 cph was concentrated in the seasonal thermocline over an approximate scale (b) of 200 m that decayed exponentially with depth, approaching 0.1 cph at 1000 m. The baroclinic modes of the vertical velocity eigenfunction (φn) for near-inertial motions [σ ≤ k n]

were estimated from

\[ \frac{d^2 \phi_n}{dz^2} + \frac{N^2(z)K_n^2}{\sigma^2 - f^2} \phi_n = 0, \tag{3} \]

where φn is the nth eigenfunction, \( N^2(z) \) is a spatially averaged buoyancy frequency profile, \( K_n = (k^2 + F)^{1/2} \) is the total horizontal wavenumber for each n mode, and \( \sigma \) is the wave frequency, which was shifted above \( f \) by 4% based on the mixed layer Burger number. Note that phase speed of the wave \( c_n \) is given by \( \sigma f / 2 \). Equation (3) was solved numerically by imposing rigid-free surface and bottom boundary conditions of \( \phi_n = 0 \) at \( z = 0, -D \) (where \( D = 1000 m \)). Since the baroclinic velocities were measured by the profilers, the imposition of a rigid lid for the surface boundary condition and neglect of the barotropic component in the near-inertial band is justified (Shay and Chang 1997). Baroclinic phase speeds, horizontal wavelengths, deformation radii, and baroclinic timescales [from (1)] are listed in Table 2. As mode number increased, these baroclinic timescales increased allowing lower modes to separate from the mixed layer faster than higher-order baroclinic modes due to decreasing phase speeds (Gill 1984).

The normalized vertical (φn) and horizontal velocity eigenfunctions (dφn/dz) for the first four baroclinic modes are shown in Figs. 5b–c. For mode 1, the largest horizontal velocity was in the mixed layer; however, as mode number increased, larger values were displaced downward. That is, the higher-order modal eigenfunctions not only increased relative to their mixed layer value, but the amplitudes contained more vertical shear that contributes to upper-ocean mixing events (Niiler and Kraus 1977).

f. Current meter moorings

Moored current meter observations were acquired along 92°W extending from the Louisiana shelf to the central Gulf of Mexico (Fig. 1a). These moorings were instrumented with General Oceanics Niskin Winged and Aanderaa RCM 4 and 5 current meters sampling current speed, direction, and temperature at intervals of 30 and 60 min (Hamilton 1990). LCWCR F passed directly over the GG mooring at nearly the same time Gilbert passed through the western Gulf of Mexico. Although moorings EE and GG were located 5–8 \( R_{max} \) north of the storm track, they were within Gilbert’s broad wind field (Fig. 3).

To isolate the near-inertial response at the moorings, tidal components were removed from the time series. Detided time series at EE and GG were bandpass filtered between 22 and 36 h with full power between 28 and 32 h, revealing energetic near-inertial oscillations with amplitudes of 0.25, 0.12, and 0.06 m s\(^{-1}\) at GG, and 0.16, 0.03, and 0.05 m s\(^{-1}\) at EE at 100, 300, and 725 m, respectively (Fig. 6). In addition to the initial current increase near day 0600 UTC 16 September \( (t = 0) \), the currents increased between 8 to 9 IP following hurricane passage as was also found in the Frederic observations (Shay and Elsberry 1987).

Prior to Gilbert, 40-h low-pass filtered current meter records from mooring GG indicated a surface-intensified flow regime of 0.70 m s\(^{-1}\) at 100 m decreasing to 0.08 m s\(^{-1}\) at 725 m within the LCWCR F (Fig. 7). Over the first few IP following passage, these speeds increased to 0.80 m s\(^{-1}\) at 100 m, 0.2 m s\(^{-1}\) at 305 m, and 0.08 m s\(^{-1}\) at 725 m, and remained constant for the next 3–4 IP. Since the vertical extent of the LCWCR current structure exceeded 725 m, but did not extend to 1650 m, and, since the nominal maximum depth of the AXBTs was 750 m, the reference level for the geostrophic currents was chosen to be 750 m.

3. Low-frequency dynamics

Geostrophic currents are calculated using observed temperature profiles and climatological T–S relationships to estimate the salinity profiles and then the density profiles. These geostrophic components are removed from the observed profiles to form residual profiles,

\[ V_i = V_o - V_{gst} - V_{gb}, \tag{4} \]

where \( V_i \) is the observed current (measured minus the surface-wave-induced component), \( V_{gst} \) represents the geostrophic term associated with horizontal density gradients induced by the hurricane, and \( V_{gb} \) is the ambient geostrophic current or time-averaged flow. This kinematical approach is useful in understanding both the geostrophic and near-inertial wave wakes excited by the hurricane passage.

a. Geostrophic flow

The predominant circulation feature was the LCWCR F, which was extensively sampled as part of the MMS field program (SAIC 1989). Upon separation from the Loop Current, LCWCR F was visible in satellite images in early May and in the OA maps of the 8°, 15°, and 18°C isotherm depths for yeardays 200 and 300 (Fig. 8). The center of the LCWCR F was located at 24°30’N,
Fig. 5. (a) Buoyancy frequency (cph: solid) and corresponding temperature profile (°C: dashed) and the amplitudes of the (b) vertical ($\phi_z$) and (c) horizontal velocity ($d\phi_z/dz$) eigenfunctions for baroclinic modes 1 (solid), 2 (dashed), 3 (dotted), and 4 (chain-dotted) from a CTD station in the western Gulf of Mexico at about 23°N.

94°15'W based upon these isothermal maps. These maps, drifter tracks (see Fig. 1) and satellite images (not shown) indicated that this ring propagated through the experimental domain in a west to southwest direction at 3–4 km d$^{-1}$, eventually decaying along the Texas shelf in 1989.

An extensive set of CTD data was acquired from yeardays 291 to 305 within one month following the Gilbert experiments. This span of data, designated as yearday 300, was used to estimate geostrophic flow fields (Fig. 9) as described in the appendix. The maximum geostrophic flow of 1 m s$^{-1}$ was located between depths 20 to 70 m within the LCWCR F, consistent with previous studies where peak current speeds ranged from
0.8 to 2 m s\(^{-1}\). Even at 200 m, currents within the LCWCR F were 0.5 m s\(^{-1}\), consistent with the low-pass filtered current meter records at GG (see Fig. 7). Within this baroclinic structure, the dynamic height was 1.3 dyn m at 30 m and 0.7 dyn m at 200 m. On the periphery of this baroclinic feature, the horizontal gradients in these dynamic heights decreased as reflected in weaker currents of 0.1–0.2 m s\(^{-1}\).

The relative vorticity fields (Fig. 9) were calculated from the second derivatives of the splined dynamic height fields and were normalized by the local Coriolis parameter. Notice the two cyclonically rotating rings located on the periphery of the LCWCR F. This pairing of an anticyclonic ring with two cyclonic rings was previously found from survey-based observations acquired in 1978 by Merrell and Morrison (1981). The magnitude of the relative vorticity in the upper 200 m for yearday 300 was 4%–6% below \(f\), indicating the possibility of frequency shifts in the near-inertial motions as they interact with the geostrophic flow (Kunze 1985). The relative vorticity normalized by \(f\) represents a form of the Rossby number (Ro), suggesting that the advective acceleration compared to the Coriolis acceleration may have been initially weak as Ro < 0.1. However, a cautionary note is that the observational grids used were relatively coarse resolution for relative vorticity estimation. Consequently, these smoothed relative vorticity estimates may not necessarily capture the largest amplitude features associated with smaller-scale variability.

Temperature profiles from yeardays 187–217 (two months prior to Gilbert) encompassed a larger portion of the western Gulf of Mexico. These profiles were converted to dynamic height fields and mapped to a regular grid for yearday 200 (Fig. 10). LCWCR F was evident as a high in dynamic height of 1.3 dyn m with peak geostrophic currents of 0.9 m s\(^{-1}\). The normalized vorticity in the LCWCR was between 4% and 6% less than the local Coriolis parameter in agreement with the yearday 300 results. These fields suggested a complex geostrophic flow structure in the LCWCR F with cyclonically rotating vortices along its periphery.

<table>
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<th>Mode</th>
<th>(c_n) (m s(^{-1}))</th>
<th>(\alpha^{-1}) (km)</th>
<th>(\lambda) (km)</th>
<th>(\int_1^{200} \phi_{V} , dz) (m)</th>
<th>(t_f) (ips)</th>
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<td>146.0</td>
</tr>
<tr>
<td>10</td>
<td>0.33</td>
<td>5</td>
<td>33</td>
<td>3567</td>
<td>183.0</td>
</tr>
</tbody>
</table>

Geostrophic flows and normalized relative vorticity fields from the Wake 1 experiment (yearday 262) indicated similar energetic flows within this background eddy field (Fig. 11). Maximum geostrophic currents in the upper 30–50 m were about 1 m s\(^{-1}\) in the LCWCR. In the core of the LCWCR F where cooling was observed (\(28^\circ-27^\circ\)C), the anticyclonic relative vorticities decreased to 5%–6% below \(f\) and pools of cyclonic relative vorticities were about 2% above \(f\). At 80 and 200 m, relative vorticities remained 5%–6% below \(f\) where geostrophic flows were 0.4–0.7 m s\(^{-1}\), consistent with observed mean flows at the GG mooring. Closer to Gilbert’s track, the relative vorticities ranged between ±2%. The largest geostrophic current from the Storm surveys exceeded 0.9 m s\(^{-1}\) at a depth of 80 m, whereas geostrophic currents in the Wake 2 experiment were similar to those in the Wake 1 experiment. Peak observed currents measured by the profilers were 1.1, 1.4, and 1 m s\(^{-1}\) in the mixed layer for the Storm, Wake 1, and Wake 2 flights, respectively (Shay et al. 1992). Thus, the geostrophic flows were strong and complex, exhibiting asymmetric structure in the LCWCR F and mesoscale structures that affected the momentum response to Gilbert.

b. Geostrophic wake

The geostrophic flows (\(V_g\)) obtained from the dynamic height fields contained two components:

\[
V_g = V_{gst} + V_{gb},
\]

where \(V_{gst}\) represents the transient geostrophic term associated with the hurricane-induced upwelling and downwelling of the density gradients and \(V_{gb}\) is the background geostrophic current. These background geostrophic profiles were estimated from 8 to 12 alongtrack profiles of the geostrophic currents (\(V_g\)) at each of the AXCP sites. A subset of these profiles was randomly sampled to form a mean and standard deviation of \(V_{gb}\) at 3-m intervals using bootstrap techniques (Efron and Gong 1983). These background current profiles indicate marked agreement between the track and 4\(R_{max}\) (Fig. 12). Both velocity components had maximum normalized values of 0.8 in the upper ocean and large uncertainty limits along the track, whereas from 1 to 4\(R_{max}\) the background geostrophic flow of 0.1 m s\(^{-1}\) was toward the northwest direction. Beyond the distance of 4\(R_{max}\), the current profiles were within the LCWCR regime as the \(u\) component with normalized values of 2 (0.2 m s\(^{-1}\)) indicated a net westward flow, whereas this component reversed direction at 5\(R_{max}\). Similar trends were also evident in the \(v\) component. The corresponding time-averaged vorticities were typically within ±2% of \(f\), suggesting a minimal influence on the response (Fig. 12c). However, the averaging was done over both the anticyclonic and cyclonic regimes between 4 and 6\(R_{max}\), which eliminated any large anomaly in the relative vorticities, particularly within the LCWCR.
Storm-induced geostrophic current profiles \( (V_{gs}) \) at 2\( R_{max} \) were determined by subtracting the background flows in Fig. 12 from the total geostrophic current \( (V_g) \) at each AXCP site. These geostrophic flow variations were due to the upwelling of the isopycnals (isotherms) at approximately 0.25 and 1.25 IPs, and downwelling of isopycnals (isotherms) at about 2.75 IP. These upwelling and downwelling events coincided with weak geostrophic currents between 0.3, 1.2, and 2.7 IPs following passage due to the baroclinic adjustment process (Rossby 1938). As time evolved, more energy leaked into the geostrophic flows as compared to the initial...
energy (Fig. 13c). For example, 5% of the initial kinetic energy was due to the geostrophy after 0.3 IP, reaching a limit of about 15%–20% after 3 IP. The sampling period for the Gilbert observations was too short to attain geostrophic equilibrium, which usually is expected to occur 10–12 IP following passage (Price 1983).

4. Near-inertial response

a. Residual profiles

To isolate the salient features of the near-inertial current response, the total geostrophic flows ($V_g$) at each AXCP profiler site were removed from the observed velocity profiles in the storm-based coordinate system to form residual profiles. Alongtrack profiles at $2R_{max}$ for the observed ($V_o$), geostrophic ($V_g$), and backrotated, residual velocity profiles ($V_r$) by a frequency of $1.04f$ were normalized based upon the scaling in Table 1 (Fig. 14). Maximum normalized velocities ranged from 0.9 to 1.3 (i.e., 1–1.4 m s$^{-1}$) prior to the removal of the geostrophic current (Fig. 14a). The mixed layer vector rotated anticyclonically over inertial periods (IP $\approx$ 30 h) in time [or inertial wavelengths ($\Lambda \approx 586$ km) in alongtrack distance] starting at the storm profile and exponentially decayed over approximately 4 IP. During Storm and Wake 1 experiments (0.3–1.7 IP), mixed layer currents were directed away from the track (divergence), whereas during Wake 2 (2.6–3.2 IP) upper-ocean velocities were directed toward the track (conver-
gence). This time-dependent upper-ocean current divergence and convergence forced by the translating wind stress pattern caused upwelling and downwelling of the isopycnals (isotherms) (Price 1983). By contrast, the total geostrophic currents were less intense having amplitudes of about 2 (0.22 m s\(^{-1}\)), but were nearly twice the predicted geostrophic currents induced by the isopycnal displacements (Table 1). Since the background flows were 0.1 m s\(^{-1}\) at 2\(R_{\text{max}}\) (Figs. 9, 10), the additional current was within the predicted range for the storm-induced geostrophic component of about 0.11 m s\(^{-1}\).

After the removal of geostrophic components from the profiles, backrotated current profiles (Fig. 14c) were 0.8–1.1, equating to maximum flows of 1.2 m s\(^{-1}\). The consistent alignment of the backrotated velocity vectors (i.e., in nearly the same direction as the storm current profile) suggests that the frequency of 1.04\(f\) was close to the oscillating frequency in the mixed layer as predicted from the mixed layer Burger number. The corresponding complex correlation coefficients for these backrotated velocity profiles relative to the storm profile exceeded 0.5 at both 2 and
Fig. 9. Objectively analyzed temperatures (°C) and the 750-m geostrophic current vectors (cm s⁻¹) (left panels) and the streamfunction (ψ) (dyn m) (right panels) superimposed on the normalized relative vorticity field (% of f) at 30, 50, 80, and 200 m on yearday 300 based upon survey data acquired as part of the MMS sponsored program in the western Gulf of Mexico (SAIC 1989).

However, along the track, these correlation coefficients were <0.5, presumably due to the strong wind field asymmetry and a rightward bias in the response (Chang and Anthes 1978). These coefficients decreased in the last two profiles where the backrotated velocities were not necessarily aligned in the same directions as the other profiles. These last two profiles may represent a second wake regime differing
from the directly forced regime, occurring about 0.7 IP after a baroclinic timescale of 2.1 IP.

b. Near-inertial profiles

Given this space–time variability of the residual velocity profiles in the alongtrack velocity sections from −1 to 5 \( R_{\text{max}} \), the near-inertial content of these profiles was determined using least squares in storm-based coordinates (Rossby and Sanford 1976), including an exponential decay timescale of 4 IP in the mixed layer. For each trial frequency starting at 0.8\( f \) and ending at 1.2\( f \), the residual profiles were least squares fit to an expression of the form

\[
\hat{v}(t, \theta) = \sum_{n=1}^{N} A_n \exp\left(-\frac{t}{\tau_n}\right) \cos \left(2\pi f_n t + \phi_n\right)
\]

where \( A_n, \tau_n, f_n, \) and \( \phi_n \) are the amplitude, decay timescale, frequency, and phase of the \( n\)-th harmonic, respectively.
Fig. 11. Geostrophic vectors (cm s\(^{-1}\)) at 750 m superimposed on the normalized relative vorticities (% of \(f\)) from objectively analyzing the Wake 1 (yearday 262) of Hurricane Gilbert AXCP data at (a) 30 m, (b) 50 m, (c) 80 m, and (d) 200 m in the western Gulf of Mexico. Normalized relative vorticity field above and below \(f\) are depicted by the red and purple on the color bar for each panel.

\[ U(z, t) = [A_u(z, t) \cos(\sigma\delta t + \theta) + B_u(z, t) \sin(\sigma\delta t + \theta)]e^{-(\omega_1t)} + u(z, t), \]  
\[ V(z, t) = [A_v(z, t) \sin(\sigma\delta t + \theta) + B_v(z, t) \cos(\sigma\delta t + \theta)]e^{-(\omega_1t)} + v(z, t), \]  

where \(A\) and \(B\) represent Fourier amplitudes for \(u\) and \(v\) components, respectively.
Fig. 12. Time-averaged geostrophic profiles ($V_{gb}$) in the alongtrack direction from $-1$ to $6R_{max}$ for the (a) $u$ component, (b) $v$ component, and (c) relative vorticity normalized by $f$. The velocities were normalized by $V_g$ (0.11 m s$^{-1}$), and stippled regions represent the 95% confidence intervals estimated from bootstrap techniques.

$v$ components, $\delta t = (X_s - \delta x)/U_h$ represents the time interval and distance ($\delta x$) of the profiler relative to the storm position at $X_s$, $\theta$ is the phase angle, and $u_r$ and $v_r$ represent unresolved currents (Marquardt 1963). The carrier frequency is defined as the frequency minimizing the covariance between the unresolved components ($\langle u_r, v_r \rangle$). To examine the issue of sampling variability, sensitivity tests were conducted based upon a near-inertial wave of known frequency and amplitude of 1 and 0.2 m s$^{-1}$ and a random noise component of 10% of these input signals. This waveform was subsampled at the increments when Gilbert data were acquired (typically 10–12 profiles) that yielded an uncertainty level of about $\pm 0.02f$.

Mixed layer currents at 0, 2, and $4R_{max}$ for the frequency band from 1.03 to 1.05 $f$ were consistent with the predicted mixed layer Burger number and backrotated residual profiles (Fig. 15). Maximum normalized amplitudes increased from 0.8 to 1.2 from the track to $4R_{max}$. Amplitudes over the near-inertial cycles were in
Fig. 13. Storm-induced geostrophic velocities ($V_{gst}$) at $2R_{max}$ for the (a) $u$ component, (b) $v$ component, and (c) the ratio of the geostrophic and observed kinetic energies. The velocities in panels a and b are normalized by $V_{gst} (0.11 \text{ m s}^{-1})$.

Phase at these cross-track distances with a maximum of 3$^\circ$ phase change, as the mixed layer velocity components were in quadrature. These velocity components indicated an asymmetric response skewed to the right of the storm track, in agreement with the observed thermal response (Shay et al. 1992). This result indicates that the wavelike mixed layer cooling patterns were due to forced near-inertial motions. Rms amplitudes (i.e., $\langle u^2, v^2 \rangle$) were 15 cm s$^{-1}$ (Table 3), in agreement with uncertainties found in other hurricane profiles (Sanford et al. 1987).

Velocity profiles at $2R_{max}$ indicated that a large fraction of the thermocline variability was due to near-inertial motions (Fig. 16). There were marked consistencies between the residual and near-inertial profiles where the maximum amplitudes were 0.9 (1.1 m s$^{-1}$) in the mixed layer. By removing the geostrophic velocity profiles, rms amplitudes associated with the covariance of the unresolved currents (Fig. 16b) decreased from 15 cm s$^{-1}$ in the mixed layer, to 5 cm s$^{-1}$ in the upper thermocline, and to 2 cm s$^{-1}$ at depth. If the geostrophic current profiles were not removed, rms differences increased by almost a factor of 2 in the thermocline, indicating that their removal was
necessary in the analysis as manifested in the high correlation coefficients between the residual and near-inertial profiles exceeding 0.8 (Fig. 16c). Thus, this approach was effective in isolating the near-inertial current response from the residual profiles despite the coarse sampling grid.

c. Thermocline response

The near-inertial motions across the mixed layer base and the thermocline are an important aspect of these evolving profiles. The normalized near-inertial currents ranged between 1 and 2 (0.3–0.5 m s\(^{-1}\)) at 0, 2, and 4
Fig. 15. Near-inertial mixed layer currents in storm coordinates at (a) 0\(R_{\text{max}}\), (b) 2\(R_{\text{max}}\), and (c) 4\(R_{\text{max}}\) for the \(u\) (solid) and \(v\) (dotted) currents relative to the residual observations \(u\) component (circles) and \(v\) component (squares). The model included an \(e\)-fold timescale for the mixed layer currents. The velocities were scaled by 1.07 m s\(^{-1}\) and the near-inertial frequency was about 1.04 \(f\), which minimized the covariance between the observations and simulations.

### Table 3

<table>
<thead>
<tr>
<th>(R_{\text{max}}) (cm s(^{-1}))</th>
<th>(A_u) (cm s(^{-1}))</th>
<th>(\phi_u) (°)</th>
<th>(A_v) (cm s(^{-1}))</th>
<th>(\phi_v) (°)</th>
<th>(a_{uv}) (cm s(^{-1}))</th>
</tr>
</thead>
<tbody>
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<td>0</td>
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<td>0.0</td>
<td>126.3</td>
<td>0.0</td>
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<td>-2.5</td>
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</tr>
<tr>
<td>4</td>
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<td>0.4</td>
<td>54.3</td>
<td>1.0</td>
<td>1.04</td>
</tr>
</tbody>
</table>

\(R_{\text{max}}\) and rotated anticyclonically with depth (Fig. 17), indicative of energy transfer from the wind-forced mixed layer to the thermocline (Leaman and Sanford 1975). The depth over which the current vector rotated 90° relative to the mean mixed layer current direction indicated an upwelling phase at about 1.2 and 3.2 IP along the track when the mixed layer was thin (Fig. 17a). The corresponding 180° directional change in the thermocline relative to the mixed layer current occurred at 150-m depth. At 2\(R_{\text{max}}\), this current vector rotation was displaced to 250 m after the first IP (Fig. 17b). These directional changes of 90° and 180° relative to the near-inertial current in the mixed layer were nearly coincident (ca. 70 m) in depth after the initial maximum of 250 m at 4\(R_{\text{max}}\), presumably due to the deeper eddy structures (Fig. 17c). These results indicate intense near-inertial currents that were highly coherent in the vertical. The maximum buoyancy frequency of about 10 cph occurred between 50 and 60 m at 0\(R_{\text{max}}\) and increased to 100 m at 2 and 4\(R_{\text{max}}\). These maxima in the buoyancy frequency coincided with the maximum in the near-inertial currents (neglecting the mixed layer), suggesting the importance of this zone for vertical mixing processes. That is, variations in the near-inertial rotating current vector with depth enhances mixing processes to the right of the storm track, as manifested in the deeper mixed layers of 60–80 m. These strong vertical shears forced the bulk Richardson number to values below unity causing the mixed layer to cool and deepen (Shay et al. 1992) in accord with numerical models (Price 1981).

### d. Effects of vorticity

To assess the role of the geostrophic relative vorticity, the \(a_{uv}\) profiles were determined from (6) and (7) for 0, 2, and 4\(R_{\text{max}}\) by minimizing the residual covariance (Fig. 18). In the upper 100 m, there was a consistent blue shift of the near-inertial frequency with depth. However, this near-inertial frequency was shifted to values less than \(f\) within the thermocline, suggesting a strong influence of the background vorticity field on the near-inertial response at depth. It decreased to 0.9\(f\) at 110, 160, and 220 m at both 0 and 2\(R_{\text{max}}\). There was also an increase in the upper 100 m to a maximum of about 1.16\(f\), followed by a decrease in the frequency toward \(f\) at 4\(R_{\text{max}}\). Along all three alongtrack sections, near-inertial frequencies indicated a significant down-
ward shift toward $f$ at 100 m. These deviations in the frequency shifts with depth were above the uncertainty level of $0.02f$ due to sampling variability and have been observed previously during other strong forcing events (Shay and Elsberry 1987; Zervakis and Levine 1995).

The minimum normalized vorticity field ($\zeta/f$) was shifted to $0.94f$ within the LCWCR between 4 and 6 $R_{\text{max}}$, while closer to the track of Gilbert, the magnitude of the vorticity decreased to a maximum of $\pm 0.02f$. In a time-averaged sense (averaging in the alongtrack direction), normalized vorticities were less, approaching $\pm 0.01f$ with a minimum of $0.02$ below $f$ at $4R_{\text{max}}$. These vorticity profiles may have been underestimated due in part to the coarse observational grids. For example, relative vorticities in smaller-scale warm core rings have approached $-0.5f$ (Kunze 1986) and $-0.3f$ (Shay et al. 1996). Such large variability in the vorticity field significantly affects near-inertial motions by altering the internal wave pass band based upon the effective Coriolis frequency ($f_{\text{eff}} = f + \zeta/2$ for $\zeta \ll f$). Here, $f_{\text{eff}}$ was $0.97f$ in the LCWCR, whereas closer to the track $f_{\text{eff}}$ was about $0.99f$, implying negligible effects on the hurricane-induced response. Although the wavelength of the forced near-inertial motions was large (586 km) compared to the scales of the vorticity fields (200 km), the vorticity field (i.e., $f_{\text{eff}}$), appear to have influenced the near-inertial motions in the thermocline. However, this scientific issue remains an open question because the sampling did not resolve the smaller-scale vorticity structures that surrounded LCWCR F.

**Fig. 16.** (a) Residual and (b) near-inertial velocity profiles (including averaged rms velocity differences) (cm s$^{-1}$) and (c) correlation coefficients between residual and near-inertial profiles at $2R_{\text{max}}$. Velocity profiles are scaled by 1.07 m s$^{-1}$. 
FIG. 17. Depths of the mixed layer (solid) and the mixed layer and thermocline near-inertial current vector differed by 90° (dashed) and 180° (dotted) (upper panels) and the time-averaged buoyancy frequency profile (shaded) and velocity profiles (lower panels) for (a) 0R_max, (b) 2R_max, and (c) 4R_max. Velocity profiles are scaled by 0.44 m s⁻¹ or 2V_th.

**e. Time series**

The near-inertial frequencies were also determined by fitting the bandpass filtered currents from the current meter arrays to a series of trial frequencies ranging from 0.8 to 1.2 f. At mooring EE, located at 7±8 R_max (Fig. 19), the blue shift in the carrier frequency of 1.04 f at 100 m increased to 1.1 f at 725 m. This frequency shift agreed with the mixed layer Burger number of 1.04 f and results from the least squares approach. At mooring GG (not shown), the results from 100 m indicated a bimodal distribution in the near-inertial frequency at 0.87 and 1.06 f due to the combined effects of the low-frequency vorticity and surface-intensified flows, respectively. At the deeper current meters, near-inertial motions had a carrier frequency of 1.02±1.04 f. Previous numerical studies have suggested a monotonically increasing frequency shift beneath the mixed layer predicted by the thermocline Burger number (Price 1983). These time series and profiler data do not support this conclusion, presum-
Fig. 18. Near-inertial frequency profiles scaled by the local Coriolis parameter ($\zeta/f$: solid) versus the envelope of variability associated with the time-averaged normalized vorticity by the local Coriolis parameter ($\zeta/f$: stippled) from the Rossby and Sanford (1976) model in storm coordinates at (a) $0R_{\text{max}}$, (b) $2R_{\text{max}}$, and (c) $4R_{\text{max}}$. Stippled regions for the normalized vorticity represent the 95% confidence intervals estimated from bootstrap techniques.
Fig. 19. Variance reduction curves (%) as a function of normalized frequency ($\sigma f$) for the $u$ (solid) and $v$ (dashed) components for (a) 100 m and (b) 725 m. Anticyclonic rotating (c) amplitudes normalized by $V_{th}$ and (d) phases (cycles) relative to the 100-m level at the point of closest approach at 0600 UTC 16 Sep 1988 at MMS mooring EE based upon complex demodulation. The abscissa is scaled as in Fig. 6.
ably due to the dynamical influence of the strong background flows.

Bandpass filtered series at EE were demodulated at these carrier frequencies to form anticyclonic-rotating amplitudes and phases at each level that represent the near-inertial response (Figs. 19c,d). At the point of closest approach (\( t = 0 \)), normalized maximum amplitudes were 0.75 at 100 m, while at depths of 305 and 725 m, they ranged from 0.1 to 0.2, respectively. At 1 and 4.5 IP, the phases at 100 and 725 m were nearly constant over the eight IPs following passage, indicative of the stable carrier frequencies found above, and a phase locking in the vertical. As time evolved, wave packets with differing frequencies and wavenumbers arrived at 305-m depth, which may have caused phase changes at that level as observed in Wake 2 profiles.

5. Simulated near-inertial current structure

a. Forced dynamical modes

Shay et al. (1989) derived a three-dimensional velocity field based on the forced dynamical modes in a stratified fluid subjected to a moving hurricane. The free mode problem is solved from (3) by imposing rigid boundary conditions for a given buoyancy frequency profile \( \bar{N}(z) \). After calculating these near-inertial baroclinic modes, the linear equations are expanded in terms of the dynamical modes for the horizontal velocities and pressures

\[
u, U, v, (x, y - U_h t), z = \sum_{j=1}^{n} u_j, v_j, \rho_j (x, y - U_h t) \frac{d \phi_j}{dz}, \tag{8}\]

and for the vertical velocities and densities

\[
u, \rho (x, y - U_h t, z) = \sum_{j=1}^{n} \nu_j, \rho_j (x, y - U_h t) \phi_j, \tag{9}\]

where \( \phi_j \) and \( d \phi_j / dz \) represent the vertical eigenfunction for the vertical and horizontal velocities, respectively, for a northward moving storm. As a result of this expansion and simplification, the wind stress (\( \tau_x, \tau_y \)) is superimposed onto the \( n \)th baroclinic modes

\[
X_n, Y_n = \frac{\tau_x, \tau_y}{\rho \int_0^h \phi_{nz} dz}, \tag{10}\]

where \( \phi_{nz} = d \phi_n / dz \) (Fig. 5c) and the mixed layer depth \( h \) is replaced by an equivalent forcing depth \( \int \phi_{nz}^2 dz \) (see Table 2) (Kundu and Thomson 1985). Since equivalent forcing increases with mode number, the projection of the modes onto the wind stress decreases, and is the rationale for the low baroclinic-mode ocean response to moving storms.

The resulting equations are solved using a Green's function, leading to a first-order Bessel function for a moving hurricane (Geisler 1970). Each set of horizontal structure coefficients in (8) and (9) are found by convolving the Green's function with the modal wind stress forcing pattern as per (10). For a steadily moving storm to the north (+y direction), the \( u \) component of the structural coefficients is

\[
u_u = \frac{1}{2U_h \alpha^2} \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} \nabla \times \tau_n (r') d\xi' d\eta', \tag{11}\]

the \( v \) component is

\[
u_v = \frac{1}{2U_h \alpha^2} \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} \left( \int_0^\infty \frac{\partial}{\partial x} \times \tau_n (r') dy \right) J_\nu (r') d\xi' d\eta', \tag{12}\]

and the vertical velocity is

\[
u_w = -\frac{1}{2U_h} \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} \nabla \times \tau_n (r') d\xi' d\eta' \tag{13}\]

where the Bessel function represents the kernel in the convolution and \( n \) is the mode number. The argument of the Bessel function is given by

\[
r'^2 = \frac{1}{\alpha^2} \left[ \left( \frac{U_h^2}{c_s^2} - 1 \right) (y' - y)^2 - (x' - x)^2 \right], \tag{14}\]

where \( \alpha^{-1} \) represents the deformation radius for the \( n \)th baroclinic mode (Table 2), and \( \nabla \times \tau_n = \partial Y_n / \partial x - \partial X_n / \partial y \) is the vertical component of the wind stress curl. The limits of integration extend from \( x' = \pm 4R_{max} \) and from 0 to \( 4R_{max} \) in \( y' \) as a result of the radiation boundary condition, including the positive and negative wind stress curl fields (Fig. 3). In the storm’s wake, near-inertial pumping induced by the moving wind stress curl pattern sets up a baroclinic ridge associated with the upwelling and downwelling phases of the response allowing for the excitation of near-inertial motions in the thermocline. The remaining terms represent the Ekman components \( Y_n / f, -X_n / f \), an additional source of vorticity, and the \( y \) component of the divergence.

Amplitudes of the Bessel function \( J_n \) for the first four baroclinic modes decayed rapidly over the first 0.5 IP following passage (Fig. 20). The first baroclinic mode decreased from 1.0 to 0.4 within 0.5 IP, and
Fig. 20. Time–space behavior of the amplitudes of the Bessel function ($J_0$) as a function of baroclinic mode number 1 (solid), mode 2 (dotted), mode 3 (dashed), and mode 4 (chain–dashed). Time or alongtrack distance is normalized by the inertial period or wavelength relative to storm passage.

Subsequently increased to 0.3 after 1 IP. Given this behavior, the transition between the near-field and far-field regimes may have occurred between 0.5 to 1 IP following passage. As time evolves, the first baroclinic mode separated from the other modes in the mixed layer between 2.1 and 2.3 IP, and was directly out of phase with the higher-order modes by 2.9 IP (Gill 1984). These predicted timescales agreed reasonably well with the observed near-inertial current variability during the Wake 2 experiment.

b. Mixed layer and thermocline response

Based upon a summation of the first ten baroclinic modes, mixed layer horizontal velocities had a maximum velocity of 1.2 m s$^{-1}$ at $R_{\text{max}}$ and indicated a net horizontal divergence and convergence pattern (Fig. 21). This rightward displacement was due to the imposed asymmetry in the wind field (Greatbatch 1983). The spreading of the velocity field with time (or equivalently alongtrack distance) of the near-inertial wave wake, filled a wedge following

$$\tan^{-1}\left(\frac{U_h}{C_n^2} - 1\right)^{1/2}$$

dependency (Geisler 1970). At the mixed layer base, the maximum upward vertical velocity was $7 \times 10^{-3}$ m s$^{-1}$ during the upwelling phase due to the mixed layer current divergence occurring at $\pi/4$ and $2\pi$ increments thereafter (Fig. 21b). Maximum horizontal velocities at the mixed layer base were about 0.82 m s$^{-1}$ (not shown). Over the next half-cycle, vertical velocities were downward as the mixed layer currents converged toward the storm track at $3\pi/4$ IP. In the thermocline (Fig. 21c), currents were reduced to a value of about 0.20 m s$^{-1}$ that scaled well with the predicted thermocline velocity scale (Table 1). This result suggests the maximum response is predictable based on the air–sea variables ($U_h$, $\tau_{\text{max}}$, and $R_{\text{max}}$), which set the initial oceanic response scales. The thermocline current was opposite to the mixed layer flow, indicating an anticyclonic rotation of the current vector with depth. Vertical velocities in the thermocline were in the same sense as those at the base of the mixed layer, with maximum amplitudes of $2 \times 10^{-3}$ m s$^{-1}$ (Fig. 21d). These fields suggest that the analytical simulations agreed with numerical studies except for the rightward bias due to nonlinear interactions (Chang and Anthes 1978).

Alongtrack sections at $2R_{\text{max}}$ indicated a flow reversal between the mixed layer, through the transition zone, and into the thermocline (Fig. 22). During the storm passage, currents were excited in the upper layer with maximum velocities in the cross-track and alongtrack directions of 1 m s$^{-1}$ and 0.4 m s$^{-1}$, respectively. Cross- and alongtrack current components were in quadrature for these forced near-inertial motions, and the structural variations were consistent with those
found in the Norbert current profiles (Shay et al. 1989). Thus, the three-dimensional structure of the simulated velocities indicated sufficient veracity to warrant a comparison to the near-inertial profiles.

c. Comparison to current profiles

Observed near-inertial profiles were rotated 68° anticyclonically to facilitate a direct comparison to the simulated profiles based upon the forced baroclinic modes. These velocity profiles at 2R_{max} were compared to the profiles from the forced analytical model based on four and ten modes (Fig. 23), revealing marked similarities between the profiles in the directly forced regime and the three-dimensional wake. The simulated vertical structure was consistent with the observed current profiles with the four-mode model (Fig. 23b). Although adding six more baroclinic modes did not necessarily improve the correlations and variance estimates (Fig. 23c), the ten-mode model contained more structure at the mixed layer base. Complex correlation coefficients between near-inertial and simulated profiles ex-
Fig. 23. (a) Near-inertial and simulated velocity structure at 2R_{max} for the (b) four and (c) ten mode models in the upper 750 m and (d) the complex correlation coefficients for two models. Explained variances (%) are given above the corresponding bars. Velocities are normalized using 1.07 m s\(^{-1}\) and times for each AXCP deployment are scaled by the inertial wavelength (\(L\)) relative to the storm center.

ceeded 0.7 for the first six profiles encompassing both the Storm and Wake 1 experiments, where the maximum correlation coefficient was 0.88 at 0.3 IP. During Wake 2, correlation coefficients were generally less ranging from 0.4 to 0.7, except for profiles from 2.7 to 2.9 IP due to the phasing between mode 1 and the other baroclinic modes beginning at 2.1 IP following passage (Gill 1984). Apart from these nodal points in Wake 2, near-inertial current variability was well approximated by a profile constructed from the first four modes, which described 66%–77% of the observed current variance. A large fraction of the simulated current variability was induced by the wind stress curl through the near-inertial pumping by the divergent and convergent mixed layer currents. One possibility was to include more baroclinic modes to capture the relative importance of the higher-order baroclinic modes, but summing 30 baroclinic modes did not significantly increase the explained vari-
ance. Other discrepancies between the simulated and observed profiles were due to relatively rapid modal decay rates, and assumed, fixed vertical structure (i.e., constant mixed layer depth). Notwithstanding, comparisons of observed to simulated profiles provided a first-order description of the forced near-inertial response in a complicated regime where $Ro < 0.1$.

6. Conclusions

Given uncertainties of storm track prediction and probe failure during the prestorm flight, the experimental objective was successfully achieved. In the absence of the flight-launched profiles, the oceanic response would have been inferred from only two current meter moorings and two National Data Buoy Center air–sea interaction buoys located 5–8 $R_{\text{max}}$ from the track. However, when mooring and ship-based data were combined with synoptic snapshots from airborne measurements, the three-dimensional evolution indicated energetic near-inertial flows and advective effects in an eddy field. This synergism of various observational techniques also proved effective in the Ocean Storms Experiment (D’Asaro et al. 1995).

Temperature profiles and climatological salinity relationships resolved the mesoscale geostrophic currents relative to 750 m within the LCWCR and gulf common water. The geostrophic wake, forced by the gradients in the isopycnal displacements, were isolated from the profiler data, accounting for 10%–20% of the variance in the gulf common water, and within the LCWCR, geostrophy accounted for 30%–40% of the observed current variance. Geostrophic velocities were removed from the observed velocity profiles, to form a time–space series of the residual velocity profiles.

The relative vorticity in the LCWCR had a minimum of 6% below $f$ and may have shifted the near-inertial frequencies below $f$ in the thermocline. Even though the kinematical effect of the LCWCR was removed from the observed profiles in this low Rossby number flow, its dynamical effect on the response altered the near-inertial frequency within the thermocline. Smaller-scale variability in these fields was not well resolved because of the coarse observational grids. Given the WKBJ constraints on the $f_{\text{crit}}$ where near-inertial wavelengths are required to be less than the scale of the baroclinic feature (Kunze 1985), the relative vorticity is important in the forced problem even when wavelengths are $O(500 \text{ km})$. The relevant scale appears to be that of the wind stress curl (i.e., $\pm 1.5–2 R_{\text{max}}$) compared to the diameter of the LCWCR of $O(200 \text{ km})$. Given the lack of direct pre-storm current measurements and higher-resolution sampling, these questions remain open and important to understanding the role of geostrophic flows on the response.

A least-squares fit of a dynamical model in storm-based coordinates to alongtrack residual profiles was effective in isolating the forced, large-scale near-inertial current variability (Rossby and Sanford 1976). Rms differences were within the envelope of expected differences based on previous studies. This result suggests that a fraction of the observed variations were due to other processes such as unresolved high-frequency internal waves and tides. Observed near-inertial frequency shifts in the mixed layer currents ranged from 1.03 to 1.05$f$, in agreement with the predicted mixed layer Burger number and bandpass-filtered mooring data at mooring EE. Near-inertial frequency shifts in the upper 100 m were increasingly blue shifted with depth, but in the thermocline there was a tendency for a shift in frequency back toward $f$. The variability at the mixed layer base and across the entrainment zone into the thermocline was associated with strong near-inertial currents that scaled well with twice the predicted thermocline velocity scale. These near-inertial shears extended into the thermocline and were responsible for decreasing the bulk Richardson numbers to below unity. These processes induced vertical mixing, which cooled and deepened the mixed layer (Shay et al. 1992; Jacob et al. 1996). During the Wake 2 experiment, there was a clear separation between the first baroclinic mode and the higher modes as suggested by Gill (1984), which may have triggered the onset of shorter wavelength wave packets emanating from the mixed layer (Zervakis and Levine 1995). This second dynamical regime became evident in the profiles where wave–wave interactions and planetary vorticity ($\beta$) may have played an important role in the momentum response.

Near-inertial velocity profiles were compared to a summation of four and ten forced dynamical modes (Shay et al. 1989). The behavior of the Bessel function delineated the near-field (directly forced) and far-field (wave wake) between 0.5 and 1 IP following passage. Over the first 1.75 IP, a summation of the first four baroclinic near-inertial modes described up to 77% of the near-inertial current variance. The variance explained by these modal estimates decreased to a relative minimum of 36% after 2.9 IP due to the phase separation between the first baroclinic mode and the higher-order modes as predicted by theory (Gill 1984). Additional modes contributed to the current differences at the mixed layer base and upper thermocline, but did not contribute significantly to the variance. Over the first two IP, a large fraction of the evolving three-dimensional current structure in the mixed layer and thermocline was described by linear near-inertial processes consistent with a low Rossby number flow. After this time, unresolved processes with shorter horizontal length scales may become important, providing another mechanism for the transfer of energy from the oceanic mixed layer into the thermocline. This issue needs to be explored with higher-resolution sampling in the wake structure for a longer period of time.

This study complements the thermal response study of Shay et al. (1992), providing the near-inertial variability to examine the roles of advection and vertical
mixing on the thermal response described in Jacob et al. (1996). The importance of these studies lies in the utilization of the observed velocity field that previous studies have neglected because of the lack of simultaneous current structure information together with temperature profiles. By only treating the temperature structure, numerical solutions may be tuned to maximize the agreement with observed temperature profiles. Concurrent velocity profiles, particularly across the entrainment zone at the mixed layer base, place more constraints on model physics implying that thermal, haline, and momentum responses have to be balanced, and must play an integral role in the model validation.

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APPENDIX

Objective Analysis

A conversion scheme for temperature to dynamic height was formulated to account for the spatially varying salinity and temperature relationship. The CTD data were sorted for salinity values in the temperature range 18°–25.5°C, and were tagged as either gulf common water or LCWCR water based on a salinity threshold of 36.6 psu, which was an obvious choice based on Fig. 4 and its use as a water mass identifier (Elliot 1982). One-dimensional spline fits to all salinity and temperature samples from the gulf common water and to the center LCWCR profile were denoted as \( S_{\text{com}} \) and \( S_{\text{ring}} \), respectively. For each sample, salinity was estimated from

\[
S = w_{\text{com}} S_{\text{com}} + w_{\text{ring}} S_{\text{ring}}, \tag{A1}
\]

where \( w_{\text{com}} + w_{\text{ring}} = 1 \) representing the relative weights. Dynamic heights, relative to the surface, were calculated from temperature profiles and estimated salinity profiles. At each of the 75 levels, bi-cubic splines were fit to each dynamic height field and were used for calculating its derivative (Inoue 1986; Mariano and Brown 1992).

The estimated geostrophic flow fields, based upon temperature profiles and historical CTD data, required tuning two different steps. First, a spline smoothness parameter was necessary to determine the scale cutoff in differentiating the dynamic height field. The relative merits between least-squares smoothing splines versus interpolating splines for the purpose of derivative estimation was determined by the roughness parameter \( \rho \) ranging from small (0.001–1) to large values (100–1000). This roughness and the other spline parameters [tension \( \tau \), number of knots, data error level] were found by an empirical search over a wide range of parameter values (Kim 1991). The optimal spline parameters were chosen to be the median value of the parameter range bracketing the normalized velocity error of 0.4–0.6. Spline parameters were \( \rho = 500, \tau = 0.9 \), fifteen knot locations, and an error level of 0.1. Due to probe failures, which require extrapolation (Inoue 1986), a larger tension \( \tau \) is optimal here than that used by Kim.

The truth data available for tuning these parameters were the 40-h low-pass filtered current meter measurements at MMS mooring GG (see Fig. 7). The normalized geostrophic velocity \( \left( u_g, v_g \right) \) error, relative to the low-pass filtered current meter estimates \( \left( u_{\text{cm}}, v_{\text{cm}} \right) \), was calculated by averaging over both depth and the two velocity components,

\[
e = (1/4) \sum_{\text{depth}} \left( \frac{(u_g - u_{\text{cm}})}{u_{\text{cm}}} \right) \left( \frac{(v_g - v_{\text{cm}})}{v_{\text{cm}}} \right), \tag{A2}
\]

This expression was minimized by an empirical search over the spline parameter space and for the two weighting functions, incorporating both linear and Gaussian spatial dependence. The derivative calculations were insensitive to both the weighting function and spline roughness parameter for reasonable values of the latter. Based on the shape of LCWCR F (Fig. 11) and an assumed similar shape for modeling rings, the weighting function was chosen to be Gaussian. The normalized mapping error was 0.6 at the mooring GG location. The normalized velocity error ranged from 0.4 to 0.86 for 20 different sets of parameters for yeardays 262 and 264.

A technique for gridding irregularly distributed data was needed to take advantage of the profiler data. The parameter matrix objective analysis algorithm of Mariano and Brown (1992) uses an anisotropic, time-dependent correlation function with nine parameters \( \left( C_1 - C_9 \right) \), and a time-dependent trend surface for efficient analysis of dynamically heterogeneous and nonstationary fields. These nine correlation parameters were strongly influenced by LCWCR F and were similar to those used in Shay et al. (1992): \( \left( C_9 = 0 \right) \), e-folding scales \( \left( C_6, C_7 \right) \) of 1.8° long and 1.2° lat, respectively, and phase speeds \( \left( C_2, C_3 \right) \) of –3 and –1 km d−1 in the x and y directions, respectively. These correlation parameters were estimated from the depth of selected isotherms 8°C, 15°C, and 18°C maps using a contour-based procedure of Mariano and Chin (1996) because most of the available data consisted of temperature profiles.

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where the signal-to-noise ratio was the largest. The mapping noise level was assumed to be 10% corresponding to \( C_i = 0.9 \) associated with subgrid-scale variability. These parameter estimates were compared to historical data acquired from five LCWCRs in the western Gulf of Mexico (Elliott 1982). The estimated mean long scale for Loop Current rings was 183 km with a range of 102–244 km, a mean phase speed of 2.1 km d\(^{-1}\) toward the west-southwest, and a peak phase speed of 4–5 km d\(^{-1}\). Cooper et al. (1990) surveyed two rings with radii of 100–150 km and found peak phase speeds of 6 km d\(^{-1}\) toward the west, whereas Vukovich and Crissman (1986) estimated a westward phase speed of 5 km d\(^{-1}\) from AVHRR data. Thus, the correlation parameter estimates lie within the range of previous results providing confidence in their values.

At each grid point in the domain, dynamic heights were calculated from CTD profiles by setting the dynamic height to zero at 750 m and integrating upward with the OA estimates at each depth interval corrected by the initial dynamic height value at 750 m. The derivatives of the dynamic height field \((\phi)\) required for the geostrophic velocities,

\[
\begin{align*}
   u_g &= -\frac{g}{f} \frac{\partial \phi}{\partial y}, \\
   v_g &= -\frac{g}{f} \frac{\partial \phi}{\partial x}
\end{align*}
\]

were calculated from least squares bi-cubic splines fit to each dynamic height map. The relative velocity profiles were corrected by the average velocity measured on the current meter moorings at 720-m depth, equating to the OA estimates at each depth interval corrected by the initial dynamic height value at 750 m. The derivatives of the dynamic height field \((\phi)\) required for the geostrophic velocities,

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