Near-Inertial Wave Wake of Hurricanes Katrina and Rita over Mesoscale Oceanic Eddies

BENJAMIN JAIMES AND LYNN K. SHAY

Division of Meteorology and Physical Oceanography, Rosenstiel School of Marine and Atmospheric Science, University of Miami, Miami, Florida

(Manuscript received 1 July 2009, in final form 29 January 2010)

ABSTRACT

Tropical cyclones (TCs) Katrina and Rita moved as major hurricanes over energetic geostrophic ocean features in the Gulf of Mexico. Increased and reduced oceanic mixed layer (OML) cooling was measured following the passage of both storms over cyclonic and anticyclonic geostrophic relative vorticity $\zeta_g$, respectively. This contrasting thermal response is investigated here in terms of the evolution of the storms’ near-inertial wave wake in geostrophic eddies. Observational data and ray-tracing techniques in realistic geostrophic flow indicate that TC-forced OML near-inertial waves are trapped in regions of negative $\zeta_g$, where they rapidly propagate into the thermocline. These anticyclonic-rotating regimes coincided with the distribution of reduced OML cooling because rapid downward dispersion of near-inertial energy reduced the amount of kinetic energy available to increase vertical shears at the OML base. By contrast, TC-forced OML near-inertial waves were stalled in upper layers of cyclonic circulations, which strengthened vertical shears and entrainment cooling. Upgoing near-inertial energy propagation dominated inside a geostrophic cyclone that interacted with Katrina; the salient characteristics of these upward-propagating waves were the following: (i) they were radiated from the ocean interior because of geostrophic adjustment following upwelling–downwelling processes; (ii) rather than with the buoyancy frequency, they amplified horizontally as they encountered increasing values of $f + \zeta_g/2$ during upward propagation; and (iii) they produced episodic vertical mixing through shear instability at a critical layer underneath the OML. To improve the prediction of TC-induced OML cooling, models must capture geostrophic features and turbulence closures must represent near-inertial wave processes such as dispersion and breaking between the OML base and the thermocline.

1. Introduction

Tropical cyclones (TCs) over the Gulf of Mexico (GOM) often propagate over the Loop Current (LC) and warm core eddies (WCEs) and cold core eddies (CCEs; Hong et al. 2000; Shay et al. 2000; Jacob and Shay 2003; Walker et al. 2005; Halliwell et al. 2008; Shay and Uhlhorn 2008; Shay 2009; Jaimes and Shay 2009, hereafter JS09). These robust mesoscale oceanic features are present at any time (Sturges and Leben 2000; Vukovich 2007); are nearly geostrophically balanced; and have characteristic horizontal velocities of $O(1–2)$ m s$^{-1}$, vertical length scales of about 800 to 1000 m, and horizontal scales between 200 and 400 km in WCEs (Mooers and Maul 1998) and between 100 and 150 km in CCEs (Hamilton 1992; Zavala-Hidalgo et al. 2003). The oceanic mixed layer (OML) thicknesses are $\sim$80 and $\sim$30 m inside WCEs and CCEs, respectively, compared with mean values of $\sim$40 m in surrounding Gulf Common Water (Jaimes 2009).

This mesoscale ocean variability imposed important dynamical constraints on the OML response to Hurricanes Katrina and Rita (2005) in the eastern GOM (JS09). The JS09 study showed that, for similar forcing, the velocity response was nearly twice as large inside CCEs (cyclones) than in WCEs (anticyclones) depending on the prestorm OML thickness topography. During both storms, most of the OML cooling was driven by vertical shear instability of TC-induced near-inertial currents (near-inertial internal wave wake). More (less) wind-driven kinetic energy was available to increase vertical shears inside the cyclones (anticyclones) because the OML current response was stronger (weaker) and energy was barely (largely) radiated into the ocean’s interior. Consequently, contrasting OML cooling levels were measured...
along the storms' tracks, suggesting an important non-linear modulation of the near-inertial wave wake of Hurricanes Katrina and Rita by the geostrophic eddy field.

Within the framework of linear theory a wake develops in the ocean behind a moving TC when the storm's translation speed $U_h$ exceeds the phase speed of the first baroclinic mode $c_1$ (Froude number $F_r = U_h/c_1 > 1$; Geisler 1970; Shay et al. 1989). Following the TC-induced upwelling, the currents in the wake become more near inertial after the first half inertial period (IP), and their transport converges toward the storm track, which forces downwelling of the isotherms. A near-inertial cycle of upwelling and downwelling develops, and horizontal pressure gradients couple the wind-forced OML with the thermocline as part of the three-dimensional wave wake (Price 1981, 1983; Brooks 1983; Shay and Elsberry 1987; Shay et al. 1989, 1998). The representation of the wave wake as a sum of forced baroclinic vertical modes has been shown to be in good agreement with observational data; that is, about 70%–80% of the near-inertial energy is contained in the four gravest vertical modes, which govern the wake dynamics (Shay and Elsberry 1987; Shay et al. 1989, 1998).

Vertical shear instability [gradient Richardson number $(\text{Ri}) < \frac{1}{4}$] of near-inertial currents in the upper ocean has been found to be associated with the third and fourth forced baroclinic modes (Shay et al. 1989). This shear instability causes turbulent vertical mixing between OML and thermocline waters and produces about 85% of the TC-induced OML cooling, based on measurement (Shay et al. 1992; Jacob et al. 2000; Shay et al. 2000), theoretical studies (Greatbatch 1984), and numerical experiments (Price 1981; Hong et al. 2000). A rightward bias in the SST response is commonly observed, because the velocity response is stronger on the right of the storm track under quiescent or weak background flow conditions (Price 1981, 1983; Jacob and Shay 2003).

Although the kinetic energy (KE) supplied by the wind stress is initially confined in the OML, vertical and horizontal wave dispersion associated with near-inertial motions spread this KE (e.g., Rubenstein 1983; Gill 1984; Nilsson 1995). Vertical dispersion beneath the OML typically depends on the stratification underneath, and the largest dispersion occurs for a large buoyancy frequency $N$ and for the smallest vertical wavenumber (Klein and Treguier 1993). The vertical transfer of energy is more efficient when the OML is deeper (as in WCEs), because the initial velocity profile has a larger projection on the first baroclinic mode (Gill 1984). The rate of vertical dispersion of near-inertial KE affects the efficiency to cool the OML by reducing the amount of KE available in the layer to entrain colder thermocline waters (Linden 1975).

Even though linear theory predicts that the wake's horizontal scales are proportional to the inertial period and the storm's size and translation speed (Geisler 1970; Greatbatch 1984), direct measurements of near-inertial energy consistently show smaller coherent horizontal scales that are apparently caused by background conditions. This has important implications, because vertical dispersion of near-inertial energy strongly depends on the horizontal scale of the near-inertial motion, and it is enhanced for small horizontal scales (Gill 1984). For instance, the local variability of background vorticity causes near-inertial oscillations to lose their initial horizontal coherence. Thus, the oscillations accumulate a phase shift over short horizontal length scales on the order of tens of kilometers; as a result of the reduction of the horizontal scale, the vertical transfer of energy is more effective, because ageostrophic horizontal velocity gradients and inertial pumping become stronger (Rubenstein and Roberts 1986). Moreover, background flow divergence dampens near-inertial motions (Weller 1982), whereas background vorticity shifts the frequency of the inertial response to either above (in cyclonic background flow) or below (in anticyclonic background flow) the local inertial frequency (Mooers 1975; Olbers 1981; Weller 1982; Kunze 1985). The polarization of inertial motions by the background flow can lead to non-negligible Ekman transport divergence (inertial pumping), even with a uniform wind stress (Klein and Hua 1988).

Vertically propagating wave groups much more energetic than the local downgoing near-inertial wave field have been observed in anticyclonic circulations (Kunze and Sanford 1984; Kunze 1986; Mied et al. 1986; Kunze et al. 1995). This behavior was explained with a wave-mean flow interaction model that predicted trapping and amplification of linear waves in regions of anticyclonic vorticity (Kunze 1985). In this model, the waves' frequency is shifted from $f$ to $f_e = f + \xi_g/2$ in the presence of geostrophic background flow ($f_e$ is the effective Coriolis parameter and $\xi_g$ the background geostrophic relative vorticity), such that horizontal gradients in $f_e$ refract the waves leading to trapping in anticyclonic flows. Predictions from three-dimensional, nonlinear dynamical models (Lee and Niiler 1998) were consistent with Kunze's (1985) linear model: (i) in anticyclones, the propagation direction is downward and toward the core, such that near-inertial energy is radiated downward from the surface to the thermocline; (ii) in cyclones, wave propagation is outward from the core of the eddy, and near-inertial energy was found only in OMLs.

In this context, the goal of this study is to evaluate the modulation of the dynamical properties of the wake of Hurricanes Katrina and Rita in geostrophic eddies. Understanding the processes driving this modulation is
important, because by changing the scales and dispersion characteristics of near-inertial motions they can affect the timing, extension, and rate of hurricane-induced vertical mixing and cooling in the upper ocean, thereby limiting the contribution from the TC-induced near-inertial motions to the global ocean internal wave power (JS09). Given the strength of Katrina and Rita surface winds and their transfer of momentum to OML and thermocline currents, the result presented here should represent the upper bound of the wind-driven vertical mixing and near-inertial motions in the World Ocean.

To accomplish our goal, data acquired in the LC system prior, during, and subsequent to the passage of Katrina and Rita are used to investigate for the first time the contrasting evolution of near-inertial currents forced by major TCs over oceanic cyclones (CCEs) and anticyclones (LC and WCEs). Airborne-, mooring-, and altimetry-based measurements used here are described elsewhere (Rogers et al. 2006; Shay 2009; JS09). A description of the geostrophic variability in the LC system during Katrina and Rita is presented in section 2 using sequences of geostrophic relative vorticity derived from altimetry-based sea surface height composites. The storms’ wake in geostrophic eddies is discussed in section 3 in terms of parametric mixing and a ray-tracing technique of near-inertial waves in geostrophic shear, based on Kunze’s (1985) model and geostrophic flow fields derived from post-Katrina and pre- and post-Rita data. The near-inertial response observed at a mooring array under the influence of a CCE during Katrina and the LC bulge (amplifying WCE) during Rita are described in section 4 with a focus on the evolution of the effective Coriolis parameter $f_c$, frequency shifting, vertical energy dispersion, amplification of near-inertial waves, and amplitude of the horizontal near-inertial currents and their vertical shears. Results are discussed in section 5 and summarized with concluding remarks in section 6.

2. The LC system eddies during Katrina and Rita

Katrina and Rita propagated over the LC system during the bulging and subsequent shedding of a WCE. Two to four (sometimes more) frontal CCEs are commonly observed along the LC periphery during WCE separation sequences (Schmitz 2005). To delineate this mesoscale variability (assumed to be geostrophic) over which the two storms moved, daily maps from satellite-based measurements of the surface height anomaly (SHA) from the National Aeronautics and Space Administration (NASA) Jason-1 and the National Oceanic and Atmospheric Administration (NOAA) Geosat Follow-On (GFO) mission were obtained for August–October 2005 (JS09). The absolute sea surface height $\eta$ was reproduced for the entire GOM by adding the SHA fields to the Combined Mean Dynamic Topography (CMDT) Rio05, which represents the mean sea surface height above a geoid computed over a 7-yr period (1993–99) and results from the ocean mean geostrophic currents (Rio and Hernandez 2004). The barotropic geostrophic flow ($V_{bg} = U_{bg} + V_{bg,j}$) was then calculated from $\eta$ with $U_{bg} = -(g/f)\partial\eta/\partial y$ and $V_{bg} = (g/f)\partial\eta/\partial x$. The vertical component of the barotropic geostrophic relative vorticity is $\xi_{bg} = \partial V_{bg}/\partial x - \partial U_{bg}/\partial y$.

During Katrina’s passage as a major hurricane over the LC system, two cyclonic circulations (CCE1 and CCE2) merged together and extended between the LC bulge and WCE (Fig. 1a). Note that the incipient CCE2 was under the action of Katrina’s wind stress. Based on recent results indicating a saturation value of the drag coefficient between 27 and $35 \text{ m s}^{-1}$ wind speeds (Powell et al. 2003; Donelan et al. 2004; Shay and Jacob 2006; Jarosz et al. 2007), we define here a saturated wind stress as the stress associated with wind speeds of $28 \text{ m s}^{-1}$ or stronger. By 2 September, the WCE reattached to the LC and two well-developed cyclones bounded the region of attachment from the north (CCE2) and south (CCE1; Fig. 1b). These WCE detachments and reattachments are commonly observed during WCE shedding events from the LC (Sturges and Leben 2000; Schmitz 2005). At this point, the CCE2 that interacted with Katrina was located over the mooring site and remained at this position for more than 2 weeks (e.g., ~14 IPs at this latitude).

On 21 September, Rita moved over the GOM from the Florida Straits, when the WCE was propagating westward and had detached from the LC. Simultaneously, the CCE1 moved between the WCE and LC bulge. This configuration prevailed during 22 September, when the hurricane crossed the LC bulge at category 5 status (Fig. 1c). Rita subsequently moved over the CCE1, where the hurricane weakened to category 4 status. On 28 September, the LC bulge appears detached from the main current, and its anticyclonic circulation extended northeastward over the mooring array (Fig. 1d).

3. Near-inertial wave wake in LC system eddies

a. Cold wake

Temperature profiles acquired during the post-Katrina and post-Rita flights resolved the geostrophic eddy field over which the two storms propagated as major TCs. These datasets provide a unique opportunity to evaluate the contrasting OML cooling levels produced by severe TCs over the LC, WCEs, and CCEs. To this end, we consider the nondimensional parameter $\varepsilon = a(T_{oml} - T_B)$ (Chang and Anthes 1978), where $a = 2.9 \times 10^{-4} \text{ °C}^{-1}$ is the expansion modulus of water at $25^\circ\text{C}$ and 30 psu.
and $T_{\text{oml}}$ and $T_{T}$ are vertically averaged temperatures over the OML depth and between the OML base and the 20°C isotherm depth, respectively. This parameter can be multiplied by the gravitational acceleration constant $g$ to provide a proxy of reduced gravity $g'$. Given that $\varepsilon$ accounts for the temperature (and density) difference between OML and thermocline waters, it is used here as a cooling mixing parameter, such that smaller values of $\varepsilon$ would indicate well-mixed waters between the OML and the thermocline. Poststorm aircraft measurements were acquired about 16 and 3 IPs after the passage of Katrina and Rita, respectively.

Under relatively quiescent ocean conditions, the cold wake of a TC is typically elongated, and extends along the right side of the storm’s track over a distance of $O(10^3)$ km in the direction of storm propagation (e.g., Chang and Anthes 1978; Price 1981, 1983; Greatbatch 1984; Shay et al. 1989). However, these characteristics were not necessarily observed following the passage of Katrina over the LC system eddies, because the region of maximum cooling was located to the left of the storm’s track 2 weeks after storm passage (Fig. 2a), indicating important horizontal advection in the wake by the energetic background geostrophic flow. Notice the high correlation levels between the region of maximum cooling and the underlying cyclonic-rotating flow isolines, suggesting that the wake was trapped in a westward-propagating CCE. Similar displacement of the cold wake was observed after the passage of Hurricane Ivan (2004) over LC system CCEs (Walker et al. 2005; Halliwell et al. 2008, 2009, manuscript submitted to Mon. Wea. Rev.).
In the case of Rita, the region of maximum cooling remained on the right side of the track 3 days following the storm’s passage and was confined to an area where CCE1 and CCE2 apparently merged (Fig. 2b). Notice that the region of maximum cooling in Rita’s wake is larger than that for Katrina. This enhanced wake could have resulted because Rita moved over the wake of Katrina. The high cooling levels on the left of Rita’s track resulted from Rita moving over remnants of Katrina’s wake, which apparently were advected by the energetic geostrophic flow. Similar to Katrina’s case, Rita produced reduced cooling levels over anticyclonic features (LC bulge). Overall, the contrasting upper-ocean mixing levels associated with the passage of these intense storms exhibited the following pattern (in terms of $\varepsilon$): (i) increased mixing in CCEs; (ii) intermediate mixing along frontal regions that separate CCEs and WCEs; and (iii) reduced mixing in anticyclones (LC and WCEs).

b. Near-inertial wave ray tracing

To explore the effects of the geostrophic eddies on the near-inertial wave wake structure, a ray trace model is
used here to investigate the propagation of near-inertial waves in a geostrophically balanced flow field (Kunze 1985). In this model, the wave’s position is given by
\[
\frac{d\mathbf{r}}{dt} = C_s + \mathbf{V}_g
\]
and wavenumber vector
\[
\frac{d\mathbf{K}}{dt} = -\mathbf{V}\omega,
\]
where \(\mathbf{r} = x\mathbf{i} + y\mathbf{j} + z\mathbf{k}\), \(\mathbf{K} = k_x\mathbf{i} + k_y\mathbf{j} + k_z\mathbf{k}\), and the dispersion relation is
\[
\omega = \omega_o + (\mathbf{K} \cdot \mathbf{V}_g) = \text{constant},
\]
with
\[
\omega_o = f + \frac{N^2k_{H}^2}{2f k_z^2} + \frac{1}{k_z} \left( \frac{\partial U_g}{\partial z} k_y - \frac{\partial V_g}{\partial z} k_x \right).
\]

Kunze’s (1985) model can be used without the geostrophic straining terms in this particular case.

In accord with the theoretical expectation (Kunze 1985) and numerical experiments (Lee and Niler 1998), the model predicted that Katrina-forced near-inertial waves initially propagating downward from the sea surface \(k_z > 0\) are trapped in WCEs, reaching the seasonal thermocline \((\sim 200\text{ m})\) in about 5 IP (Fig. 3a). By contrast, waves propagating in CCEs remained at a nearly constant depth (vertical stalling) but were horizontally radiated from the eddy center toward the periphery of the feature. Notice the strong mean advection of waves along the LC’s northern portion, where the waves rapidly propagated downward in the side of negative relative vorticity. In the case of the Rita-forced near-inertial waves, the model predicted similar propagation patterns to those in Katrina, although the waves propagated over a region with stronger geostrophic horizontal straining (Fig. 3b). Several rays inside the CCE2 were short over the region of maximum vertical mixing (at about \(27^\circ-27.5^\circ\text{N, 90.5}^\circ\text{W}\); cf. Figs. 2b, 3b) as the predicted wave amplitude grew exponentially during the first IP because of strong horizontal gradients in the cyclonic geostrophic flows.

Near-inertial waves were predicted to spend more time in surface waters of CCEs, coinciding with regions of enhanced vertical mixing. This vertical stalling of downward-propagating waves in CCEs may induce an accumulation of near-inertial energy in surface layers that amplify vertical current shears. By contrast, reduced upper-ocean vertical mixing occurred over regions of anticyclonic \(\zeta\), where near-inertial waves were dispersed rapidly downward.

4. Near-inertial response

a. Effective Coriolis frequency

Hourly data from the mooring array (ADCPs) with vertical sampling interval of \(\sim 8\text{ m}\) and CTDs deployed
every ~40 m; Table 2 of JS09) allow incorporating time dependence. For this purpose, background relative vorticity \( \zeta_R \) was calculated to identify the periods over which the mooring array was under the action of the CCE2 and LC bulge that previously interacted with Katrina and Rita, respectively. To calculate \( \zeta_R \), horizontal velocity data from the three moorings (deployed in a triangular array) were first interpolated into midpoints between the moorings (squares in Fig. 1). Gaussian-weighted averaging, given by

\[
\overline{Q}(z, t) = \frac{\sum_i Q_i(z, t)e^{-r_i^2/\sigma^2}}{\sum_i e^{-r_i^2/\sigma^2}},
\]

was used to obtain interpolated values \( \overline{Q}(z, t) \) at these midpoints, where \( i = 1, 2, 3 \) is a mooring index; \( Q_i(z, t) \) is the velocity data \( [u(z, t) \text{ or } v(z, t)] \); \( r_i \) is the distances between the interpolation point and actual data points; and \( r_o \) is the maximum \( r_i \) distance. These three midpoints,

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**Fig. 3.** Near-inertial wave ray tracing based on Kunze’s (1985) model for (a) Katrina and (b) Rita. The numbers along the wave rays indicate inertial periods (one inertial period is ~25.5 h), dots are hourly positions, color is the ray’s depth level, and the flow lines are from geostrophic flow fields derived from (a) post-Katrina (15 Sep) and (b) post-Rita (26 Sep) air-borne-based data. The gray shades represent regions where \( (\zeta_R - |S_j|)/f > 0.2 \). This ratio and the flow lines were calculated from depth-averaged velocity fields.
together with the most western mooring, provide a diamond-type array (with vertices at the cardinal points N, S, E, and W) suitable to compute relative vorticity \( z \) at the diamond’s center \((x_c, y_c)\). For each depth level, hourly values of relative vorticity are computed with \( z(x_c, y_c, z, t) = \frac{(v_E - v_W)/\Delta x - (u_N - u_S)/\Delta y}{D_x^2} \left(\frac{u_N^2 - u_S^2}{D_y^2}\right) \). Finally, \( z_g \) was calculated for each depth level at intervals \( \Delta t = IP/4 \) (IP \( \sim 25.5 \) h) with

\[
\xi_g(x_c, y_c, z, t_e) = \frac{1}{M} \sum_{t=t_e-2IP}^{t_e+2IP} \xi(x_c, y_c, z, t),
\]

where \( t_e = 0IP, 0.25IP, 0.5IP, \ldots , 37.75IP, 38IP \) and \( M \) is the number of hourly data used to calculate this 5-IP running means. This 5-IP length was selected based on the estimated time that the eddies remained over the mooring and to increase the statistical significance of the mean compared with 1-IP running means aimed to resolve near-inertial waves in other contexts.

The evolution of \( \xi_g \) at the mooring array shows alternations as a function of the arrival of cyclonic and anticyclonic features (Fig. 4). Notice that the IP scaling in Fig. 4 is reset to zero at the time of closest approach of Rita to the mooring array. Thus, the first IP segment corresponds to Katrina and the second corresponds to Rita. The period from about 4 to 13 IP (Katrina segment) is associated with the arrival of the CCE2, and the negative values of \( \xi_g \) from 15 to 19 IP (13–18 September) correspond to an intrusion of the LC bulge prior to Rita’s passage. The period from about 6 to 10 IP (Rita segment) captures the intrusion of the LC bulge under the influence of Rita. For the remainder of this section, the focus is on the periods from 5 to 8 IP in the case of the CCE2 (Katrina segment, which is referred to here as CCE) and from 7 to 10 IP in the case of the LC bulge (Rita segment).

The time mean of the effective Coriolis parameter \( f_c \) calculated from \( \xi_g \) shows contrasting regimes between the CCE and LC (Fig. 5). The lower bound of the near-inertial internal wave band is shifted toward higher and lower frequencies in the CCE and LC bulge, respectively, as suggested by theoretical developments (Mooers 1975; Weller 1982; Kunze 1985) and observations (Kunze and Sanford 1984; Kunze 1986; Mied et al. 1986, 1987; Kunze et al. 1995). Note that below the 250-m depth level, \( f_c \) is nearly symmetric between the LC and CCE. However, above this depth, \( f_c \) is surface intensified in the CCE as the vorticity gradients tighten. Thus, the expectation is for \( \xi_g \) to alter the near-inertial frequency passband and vertical wavenumbers that may exist within these eddies.

b. Frequency shifting

Data from the most western mooring (closest to Katrina’s and Rita’s tracks) were used to diagnose the frequency of near-inertial oscillations inside the CCE and LC bulge. For each depth level, the velocity components were rotated to the storm coordinate system (cross- and along-track velocities). Perturbation currents were acquired by
removing background velocities from 5 to 8 IP in the CCE and from 7 to 10 IP in the LC bulge. A least squares frequency analysis involved using perturbation velocities for a series of trial frequencies (Rossby and Sanford 1976; Mayer et al. 1981; Shay and Elsberry 1987). This analysis determines a set of weights ($A_1$ and $A_2$ or velocity amplitudes) for each velocity component,

$$[u(z, t), v(z, t)] = A_1 \cos(\omega t) + A_2 \sin(\omega t) + u_r(t),$$

where $u(z, t)$ and $v(z, t)$ are observed perturbation velocities in the cross- and along-track directions, respectively; $\omega$ is the trial frequency (from 0.5$f$ to 2.5$f$); and $u_r$ is the residual velocity after removing the signal with that frequency. The overall quality of the fit (considering the two velocity components) is given in terms of the correlation coefficient

$$r = \frac{(r_u^2 + r_v^2)^{1/2}}{2^{1/2}},$$

where $[r_u, r_v] = ss_{xy}/ss_{xx}ss_{yy}$, $r_u$ and $r_v$ are the correlation coefficients between observed and modeled velocities for cross- and along-track velocities, respectively; $ss_{xy}$ the covariance matrix between the observed and modeled velocities; and $ss_{xx}$ and $ss_{yy}$ are the variance matrices of observed and modeled velocities, respectively. Because of the strength of the near-inertial response, it is assumed that a single carrier frequency exists (Shay and Elsberry 1987). For each depth, this frequency is defined as the frequency that minimizes the residual covariance (in the present treatment with the highest $r$) during the period from 5 to 8 IP.

Inside the CCE, the near-inertial response to Katrina indicates two clear patterns of frequency shifting (Fig. 6a). This small departure from $f$ makes these waves near inertial and enables them to propagate vertically in the water column. Near-inertial waves become more influenced by ambient rotation as they propagate in the direction of increasing $f$. In deeper waters (below 300 m), the optimal frequency fits occurred at higher frequencies than $f$, with a peak at $\sim 1.05f$. This blue shift is in agreement with the distribution of $f$ over this depth range (cf. Fig. 5). By contrast, in surface waters (above 300 m), the optimal frequency fits occur at lower frequencies than $f$, with peak values of $r$ (approximately 0.70–0.75) at about $0.95f$. This red shift in surface waters cannot be simply explained by the distribution of $f$, which indicates that other processes are involved. According to linear theory, in addition to $f$, the effective buoyant frequency and the vertical geostrophic shear [second and third terms on the right-hand side of Eq. (4), respectively], also contribute to the intrinsic frequency, but these contributions are between one to two orders of magnitude smaller than $f$ (Table 2) so that they should have little impact on the red shift in the frequency. However, the Eulerian frequency measured in the mooring can be importantly affected by the Doppler term $\mathbf{K} \cdot \mathbf{V}_g$ [Eq. (3)], which is about $\frac{1}{3}$ of $f$, and is large enough to shift the near-inertial response from super to subinertial frequencies.

Given that, in the Northern Hemisphere, near-inertial oscillations rotate anti-cyclonically while $\mathbf{V}_g$ is cyclonic in CCEs, the hypothesis is that the geostrophic Doppler term delays the near-inertial frequency in surfaces waters and shifts the near-inertial response to subinertial frequencies. Unfortunately, we cannot evaluate this idea because it is not possible to resolve actual horizontal near-inertial wavenumbers (including sign) with the available mooring...
The terms $U_g$ and $\Delta U_g$ are from Fig. 9a, where $\Delta U_g$ is the velocity difference between the depth levels $z = -100$ m and $z = -200$ m, and $\Delta z = 100$ m.

data. Mied et al. (1987) reported that the Doppler term had an important impact on the near-inertial wave structure inside a CCE in the North Atlantic subtropical zone.

In the case of the near-inertial response inside the LC bulge (Fig. 6b), the frequency fits are better than in the CCE above the 100-m depth level. In the upper 250 m, the fits are slightly skewed toward lower frequencies than $f_i$ between 250- and 430-m depths, there is a slight shift toward higher frequencies than $f_i$. The blue shift in the LC is nearly homogeneous with depth in the upper 400 m, in agreement with the distribution of $f_i$ (Fig. 5).

c. Vertical wavenumber spectrum

Rotary spectra of vertical wavelengths were computed to delineate the vertical propagation of near-inertial energy (Gonella 1972). These vertical wavelengths were calculated with the procedure of Leaman and Sanford (1975) by the following: (i) vertical averages of the horizontal velocity components from each profile of the mooring data were removed; (ii) time averages of the horizontal velocity components were removed at each depth level; (iii) velocity components were scaled based on the Wentzel–Kramers–Brillouin–Jeffreys (WKBJ) approximation, with

$$u_n(z) = u(z)\left[\overline{N}(z)/N_o\right]^{1/2}$$

and

$$v_n(z) = v(z)\left[\overline{N}(z)/N_o\right]^{1/2},$$

where $u$ and $v$ are the original horizontal velocity components, $u_n$ and $v_n$ are the scaled velocities, $\overline{N}(z)$ is the mean buoyancy frequency profile, and $N_o$ is a reference buoyancy frequency equal to $3$ cph depending on the basin; (iv) vertical coordinate was stroched according to $dz_n = [\overline{N}(z)/N_o] dz$, where $z$ is the original vertical coordinate (meters) and $z_n$ is the “stroched” vertical coordinate [stroched meters (sm); Table 3]; (v) original profiles were interpolated to equally spaced $z_n$ levels; and (vi) hourly stroched profiles were time averaged from 5 to 7 IP for the CCE that interacted with Katrina and from 8 to 10 IP for the LC bulge affected by Rita. Rotary spectra were independently calculated for the CCE and LC bulge.

The predominance of the anti-clockwise (ACW) rotating over the clockwise (CW) rotating part of the rotary spectrum is the most striking aspect of the near-inertial response inside the CCE (Figs. 7a,b), denoting upward energy propagation (Leaman and Sanford 1975; Leaman 1976). Inside this eddy, upgoing energy propagation is about 10 times larger than downward propagation for wavelengths from $\sim 100$ to 250 sm; for shorter wavelengths, the spectra is nearly equally partitioned between ACW and CW components. Notice that the energy peak of downgoing energy is narrower and skewed toward longer wavelengths, with a peak from about 200 to 300 sm. The rapid decay at larger wavelengths in both the ACW and CW spectra is an artifact, because wavelengths larger than 780 sm ($\sim 500$ m; Table 3) were not resolved. A region of cyclonic vorticity with more upgoing than downgoing near-inertial energy at the same wavelengths was observed in a Sargasso Sea front (Mied et al. 1986).

In the case of the LC bulge, downward energy propagation dominates practically at all wavelengths (Fig. 7b), which is consistent with energy partitions from rotary spectra calculated from current profilers deployed in the GOM during and subsequent to Hurricane Gilbert (1988), which indicated that the CW rotating component dominated in 83% of the profilers, and the average ratio of CW to ACW energies was $\sim 3.6$, indicative of a preference for downward energy propagation from the wind-forced OML near-inertial currents into the thermocline, particularly in the Gulf Common Water (Shay and Jacob 2006).

Comparison of downward energy propagation (CW component) between the LC bulge and CCE indicates

<table>
<thead>
<tr>
<th>$U_g$ (m s$^{-1}$)</th>
<th>$\Delta U_g$ (m s$^{-1}$)</th>
<th>$\Delta z$ (m)</th>
<th>$K_H$ (10$^{-5}$ m$^{-1}$)</th>
<th>$K_z$ (10$^{-2}$ m$^{-1}$)</th>
<th>$f_i$ (10$^{-5}$ s$^{-1}$)</th>
<th>$B = N^2K_H^2/2K_z^2$</th>
<th>$S = \frac{K_H\Delta U_g}{K_z\Delta z}$</th>
<th>$D = K_HU_g$</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.5</td>
<td>0.15</td>
<td>100</td>
<td>3.93</td>
<td>3.5</td>
<td>6.9–7.1</td>
<td>0.06</td>
<td>10$^{-5}$ s$^{-1}$</td>
<td>10$^{-5}$ s$^{-1}$</td>
</tr>
</tbody>
</table>

TABLE 3. Comparison of real ($z_o$) and stretched ($z_s$) depth levels at the mooring site.

<table>
<thead>
<tr>
<th>$z_o$ (m)</th>
<th>$z_s$ (sm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>70.4</td>
<td>103.9</td>
</tr>
<tr>
<td>102.4</td>
<td>188.2</td>
</tr>
<tr>
<td>150.4</td>
<td>276.4</td>
</tr>
<tr>
<td>198.4</td>
<td>351.8</td>
</tr>
<tr>
<td>254.4</td>
<td>438.0</td>
</tr>
<tr>
<td>302.4</td>
<td>507.7</td>
</tr>
<tr>
<td>350.4</td>
<td>576.9</td>
</tr>
<tr>
<td>398.4</td>
<td>644.3</td>
</tr>
<tr>
<td>454.4</td>
<td>722.9</td>
</tr>
<tr>
<td>494.4</td>
<td>778.5</td>
</tr>
</tbody>
</table>
that this propagation was about 2 times more energetic inside the geostrophic anticyclone for wavelengths from about 120 to 300 sm, even though the near-inertial velocity response was about 2 times larger inside the CCE (JS09). In the CCE, the ratio of CW to ACW energies was 0.2, whereas in the LC bulge this ratio was 2.1 (Table 4). This confirms that inside the CCE only a small fraction of the KE supplied by the hurricane (18%) is exported downward into the thermocline, while 82% of this energy remains in upper layers where it is available to increase vertical shears and cooling through vertical mixing, which is consistent with the ray-tracing analysis discussed earlier and results of Lee and Niiler (1998). By contrast, in the LC bulge, less KE was available for vertical entrainment mixing events in upper layers via shear instability, because 68% of the wind-forced near-inertial energy was radiated into thermocline waters, similar to Kunze (1985, 1986).

d. Amplification of near-inertial waves in the CCE

From a theoretical perspective, near-inertial waves become more influenced by ambient rotation as they propagate in the direction of increasing \(f_e\) (Mooers 1975; Olbers 1981; Kunze 1985). In this context, of particular interest is to delineate the role of the vertical distribution of \(f_e\) (Fig. 5) on upgoing near-inertial energy in the CCE. Notice that, in this eddy, waves traveling downward from the sea surface become less inertial in character and therefore can freely propagate away from the cold eddy (cf. Fig. 3a).

Near-inertial waves propagating inside the CCE above the 250-m depth and toward the sea surface become even more near inertial because they encounter a rapidly increasing \(f_e\) (Fig. 5). Under these circumstances, the theory predicts that the vertical wavenumber must shrink for the waves to continue satisfying their dispersion relation [Eq. (3)]. Consequently, the vertical group velocity also must diminish, becoming zero at the critical layer depth where the wave’s intrinsic frequency \(\omega_o\) equals \(f_e\) (Kunze 1985). To satisfy the wave action principle (Bretherton and Garrett 1969), the reduction of the vertical scale must be compensated by wave amplification in the horizontal, until a level in which vertical shears become unstable, enhancing turbulent vertical mixing at the critical depth. The amplification of the near-inertial currents in the CCE can thus be calculated with the weights \((A_1, A_2)\) associated with the carrier frequency [Eq. (5)]. The

---

Table 4. Summary of vertical energy fluxes associated with the near-inertial waves induced by Katrina and Rita in the CCE and LC bulge, respectively. These numbers are integrated values of the product \(E(m) \times C_g(m)\) (Leaman 1976), with \(E(m)\) being the spectral energy of the \(m\)th wavenumber (from the rotary spectrum; Fig. 7a) and \(C_g(m) \approx -2f / m^2\) being the vertical group velocity of the \(m\)th wavenumber (Rossby and Sanford 1976); \(\delta\) is the departure from the local Coriolis frequency (+0.05 for downgoing energy and –0.05 for upgoing energy, based in Fig. 6). Upgoing (downgoing) energy fluxes were calculated from the ACW (CW) components of the rotary spectrum. Values in parenthesis for the upgoing and downgoing energy fluxes represent the fraction of the total vertical energy flux for the particular geostrophic feature.

<table>
<thead>
<tr>
<th></th>
<th>Upgoing ((10^{-2}) W m(^{-2}))</th>
<th>Downgoing ((10^{-2}) W m(^{-2}))</th>
<th>CW/ACW</th>
</tr>
</thead>
<tbody>
<tr>
<td>CCE</td>
<td>25.4 (82%)</td>
<td>–5.4 (18%)</td>
<td>0.2</td>
</tr>
<tr>
<td>LC</td>
<td>3.7 (32%)</td>
<td>–7.9 (68%)</td>
<td>2.1</td>
</tr>
</tbody>
</table>
weights \( (A_1, A_2) \) associated with this carrier frequency provide a proxy to the canonical amplitude of the near-inertial response \( A_m = (A_0^2 + A_1^2)^{1/2} \), where \( (A_0, A_1) = (A_0^2 + A_1^2)^{1/2} \), with \( A_0 \) and \( A_1 \) being the velocity amplitudes in the cross- and along-track directions, respectively.

As shown in Fig. 8, near-inertial amplitudes are surface intensified in the upper 250 m of the CCE, coinciding with the increase of \( f_e \). The maximum amplitude occurs at \( \sim 125 \)-m depth; above this depth, the amplitude rapidly decays about 50% over a distance of approximately 50 m. Below 300-m depth, the amplitude is slightly bottom intensified where \( f_e \) grows gradually with depth.

For waves propagating vertically in a uniform, time-independent medium, the intrinsic frequency and vertical wavenumber are constant along a ray; that is, \( -\partial \omega_{\infty}/\partial z = 0 \) and \( dk_e/\partial t = 0 \). However, in a nonuniformly moving medium (e.g., CCE), the intrinsic frequency \( \omega_{\infty} \) varies along a ray (Bretherton and Garrett 1969) so that \( dk_e/\partial t = \pm \omega_{\infty}/\partial z \) or \( d\lambda_e/\partial t = \partial \omega_{\infty}/\partial z \) for upgoing waves \( (m > 0) \). As shown in Fig. 8, the increase of \( \omega_{\infty} \) from the depth to the surface coincides with wave amplification, indicating that the reduction of the vertical wavelength is compensated by amplification of the horizontal velocity for waves propagating upward in the upper 250 m, in accord with theoretical predictions.

e. Horizontal near-inertial currents and vertical shears

The near-inertial response inside the CCE shows a CW rotation of horizontal currents with depth, with two regimes separated at about the 250-m depth (Fig. 9a). In the near-surface regime, the rotating helices of the velocity vector (envelops of the stick vectors) resemble a downward phase propagation that indicates upward energy propagation (Leaman and Sanford 1975). By contrast, beneath 250-m depth, the phase propagates upward (downward energy propagation). Notice that, in the near-surface regime, the current amplitudes increased from the depth to the surface (see also Fig. 8), reaching maximum levels of \( O(60) \) cm \( s^{-1} \) between 100- and 200-m depths. At these depth levels, the maximum cooling occurred (see, e.g., the change in color of the stick vectors near the 100-m depth from 6 to 7 IP, indicating a cooling of approximately \( 4^\circ-5^\circ C \)). The vertical distribution of the gradient Richardson number indicates that this maximum cooling was driven by shear instability (Fig. 9b).

Inside the LC bulge, the near-inertial response exhibited CW velocity rotation with depth (Fig. 10a), though the amplitude was about one-half the value observed in the CCE. Other important differences with respect to the CCE were as follows: (i) in the LC bulge, there was no significant amplification of the velocity response with depth (notice the nearly homogeneous vertical distribution of \( f_e \) in the LC, Fig. 6); (ii) upward phase propagation (downward energy propagation) predominated over the water column; and (iii) cooling was negligible over the full water column, because vertical shears were weak and Ri was above criticality (Fig. 10b).

5. Discussion

a. Contribution of buoyancy on the amplitude of the near-inertial waves

In addition to geostrophic relative vorticity, the density stratification can modulate the forced near-inertial oscillations, because variations in \( N \) cause the waves to change their horizontal kinetic energy and vertical wavenumber as they propagate vertically through the water column (Leaman and Sanford 1975). Thus, it is important to delineate the role of the vertical distribution of \( N \) on the change of amplitude of the Katrina- and Rita-forced waves as they propagated vertically inside the CCE and LC bulge, respectively. For this purpose, the original velocity data from the most western mooring were used to estimate the vertical distribution of time-averaged kinetic energy of the perturbations \( \mathbf{K'} \). Perturbation velocities were obtained with steps (i) and (ii) of section 4c (i.e., non-WKBJ-normalized velocities), by removing vertical and time averages from 5 to 8 IP and from 7 to 10 IP for the CCE and LC bulge, respectively. Perturbation kinetic energies \( (K') \) were then calculated from these perturbation velocities at 1-h intervals. Finally, \( K' \) was time averaged over the periods described earlier.
FIG. 9. Near-inertial response to Katrina in the CCE as observed at the mooring site (27.998°N, 87.839°W): (a) horizontal velocity response and (b) square of the vertical shear of horizontal currents $S^2 = (\partial u/\partial z)^2 + (\partial v/\partial z)^2$. The black curves in each profile in (b) are $4N^2$, with $N^2 = -(g/p)(\partial p/\partial z)$ being the square of a reference buoyancy frequency from a density profile acquired inside the CCE during a pre-Rita flight. The gradient Richardson number is $\text{Ri} = S^2/N^2$, and the theoretical limit for shear instability is $\text{Ri} < \frac{1}{4}$. 
In the case of the LC bulge, $K_9$ scales with the buoyancy frequency (Fig. 11), indicating that in this feature its background relative vorticity played a negligible role in modulating the near-inertial oscillations’ amplitude (see the nearly homogeneous vertical distribution of $f_r$ inside the LC bulge; Fig. 5). However, in the CCE, there is no apparent direct link between the variability in $N$ and the amplitude of $K'$. This supports the hypothesis that, inside this CCE, the near-inertial oscillations’ amplitude was modulated by the relative vorticity of the basic-state flow. In this eddy, there is a reduced level of $K'$ at the depth of maximum buoyancy frequency (100-m depth). This reduction of $K'$ is associated with the maximum in vertical shear instability, which dampens...
near-inertial energy of upgoing waves, via increased turbulent vertical mixing.

b. Surface maximum of cyclonic relative vorticity in the CCE

To explain the surface intensification of $\zeta_g$ inside the CCE, consider Fig. 4, which indicates two periods of cyclonic $\zeta_g$ at the mooring. The first period, corresponding to the CCE that interacted with Katrina, exhibits a baroclinic current structure between about 5 to 12 IP, with a strong vertical gradient from approximately 100- to 150-m depth range. Over this 50-m depth interval, $\zeta_g$ intensifies from 0.04 to 0.1 in the upward direction. This reveals that, during the forced stage, the input of cyclonic vorticity from Katrina’s wind stress strengthened the cyclonic circulation in the upper layers of the CCE. This effect had profound effects on the amplification of upward-propagating near-inertial waves during the relaxation stage. That is, the wind stress curl increased $f_\sigma$, which in turn strengthened vertical shears and cooling mixing.

c. Geostrophic adjustment and near-inertial wave generation

The direct generation at the OML of downward-propagating near-inertial waves by a fast-moving TC (Fr > 1; Geisler 1970) is generally well understood. However, near-inertial wave generation in the ocean’s interior associated with the TC passage has not been systematically addressed, partly because of the inherent sampling problems. In the previous section, we discussed such waves and their effects on vertical mixing and subsequent upper-ocean cooling. The observations here suggest that these upward-propagating waves may have been generated by geostrophic adjustment, in analogy with processes studied in other contexts (Rossby 1938; Gill 1982). The focus is on the CCE affected by Katrina because this fast-moving storm produced a stronger near-inertial response of upward-propagating waves in a water column extending from about 120- to 250-m depth levels (Fig. 9a).

In analogy to the upwelling–downwelling regimes induced by Rita over the LC system (JS09), the scenario is that, during Katrina passage, wind-forced surface waters diverge and denser water is upwelled along the track. This oceanic flow is enhanced when wind stress is in the direction of the cyclonically rotating geostrophic flow. By contrast, there is horizontal convergence and downwelling of lighter water in regions where the wind stress is against $\mathbf{V}_g$. Density anomalies associated with these contrasting upwelling–downwelling regimes extend to the thermocline and had to be removed by horizontal mass redistribution.

As shown in Fig. 12, the reduction of background kinetic energy $\mathbf{K}$ (crests in isolines of $\mathbf{K}$) between 5 to 9 IP (and between approximately 150- and 250-m depth levels) coincided with radiation of near-inertial wave perturbation kinetic energy. From 9 to 11 IP, the amplitude of the crests in $\mathbf{K}$ was reduced at the same time that the
production of $K'$ was dampened. Notice the tendency for $K$ to be homogenized in the vertical as time evolves (i.e., reduced energy levels at the surface and increased energy at depth), which is an indication that the sub-surface horizontal pressure gradients associated with the hurricane-induced vorticity and density anomalies were removed via radiation of near-inertial internal waves. Notice that this simultaneous increase of $K'$ and decrease of $K$ occur in a water column (approximately 100–200-m depth) below the OML base that extended to about 60-m depth.

6. Summary and concluding remarks

Katrina and Rita (2005) moved over the LC system as major hurricanes (category 5). These TCs produced contrasting OML cooling levels depending on the distribution of geostrophic relative vorticity $\zeta_g$ in the upper ocean. Cooling was increased and reduced over cyclonic and anticyclonic $\zeta_g$, respectively. To explain this differentiated cooling, airborne- and mooring-based measurements were used here to investigate the three-dimensional near-inertial wave wake induced by both TCs inside cyclonic and anticyclonic geostrophic features of the LC system.

Ray-tracing techniques in realistic geostrophic flow predicted that forced near-inertial waves initially moving downward from the OML are trapped in regions of negative $\zeta_g$, where radiation into the thermocline predominates. These anticyclonic regimes coincided with distribution of reduced OML cooling, which suggests that rapid downward dispersion of near-inertial energy reduces vertical shears in upper layers of anticyclonic features, as observed in other oceanic regions (e.g., Kunze and Sanford 1984; Kunze 1986). By contrast, downward-propagating near-inertial waves are stalled in regions of cyclonic circulation (cf. Lee and Niiler 1998), where vertical shears and entrainment cooling are increased. This contrasting wave dispersion might explain an important fraction of the reduced hurricane-induced OML cooling observed in anticyclonic features (Shay et al. 2000; Jacob and Shay 2003; Shay 2009; Shay and Uhlhorn 2008) and increased cooling in cyclonic circulation (Walker et al. 2005; Halliwell et al. 2008; JS09). Direct Eulerian and Lagrangian observations are needed to better depict the three-dimensional dispersion in the wake of TCs in the energetic geostrophic circulations.

Rather than with the buoyancy frequency $N$, near-inertial velocities at the mooring amplify as function of $f_e = f + \zeta_g/2$ inside the CCE that interacted with Katrina. This is due to the premise that WKBJ scaling often relates to scales on the order of the eddy field itself or smaller (Kunze 1985) compared to the response to large-scale atmospheric forcing. Thus, the presence of mesoscale ocean features can have an important influence in shaping the internal wave spectrum as wind-forced near-inertial waves propagate vertically in the water column; this issue requires more attention. Evidence of internal motions beyond $1.5f$ was found neither in the upper 500 m of the CCE nor in the LC bulge, as suggested by numerical studies that have found a peak at $2f$ for other regions (e.g., Niwa and Hibiya 1997; Danioux and Klein 2008).

Upgoing near-inertial energy propagation dominates inside the CCE that interacted with Katrina, in agreement with near-inertial energy propagation observed in a cyclonic eddy in the Sargasso Sea front (Mied et al. 1986). In the CCE, amplification of these upward-propagating waves produced a critical layer (increased turbulent mixing) just underneath the OML base. This can be an important mechanism for OML deepening in regions of intense cyclonic $\zeta_g$ in the World Ocean. Vertical wave energy fluxes were 4–10 times larger than in unforced cases (Leaman and Sanford 1975) but were comparable to those found in frontal regimes (Kunze and Sanford 1984) and during Hurricane Gilbert (Shay and Jacob 2006). This vertical energy propagation is highly nonlinear (Shay and Jacob 2006; JS09), and more investigation and observations are needed to improve the representation of this process in numerical models.

From a broader perspective, numerical models must be initialized with geostrophic features to improve the representation of wind-induced vertical mixing and near-inertial wave dispersion. These two processes affect the dispersion of tracers and larvae in the upper ocean and SST distributions commonly used in climatic predictions. Of particular importance is to separate near-inertial waves from turbulence in turbulent kinetic energy equations, because waves are affected by the earth’s rotation and $\zeta_g$, whereas turbulence is not.

Acknowledgments. B. Jaimes acknowledges the Fulbright-Garcia Robles Commission (United States and Mexico) and Consejo Nacional de Ciencia y Tecnología (Mexico) for their support during his initial years in the Ph.D. program at the University of Miami. Research effort has been supported by the NSF through Grants ATM-01-08218, 04-44525, and NASA-66520Y and the NOAA Joint Hurricane Testbed program. The authors are grateful for the efforts of the pilots, technicians, engineers, and scientists at NOAA’s Aircraft Operation Center (Dr. Jim McFadden) and HRD (Dr. Frank Marks, Dr. Rob Rogers, Dr. Peter Black, and Dr. Eric Uhlhorn). Drs. Kevin Leaman and William Johns of the University of Miami provided insightful discussion on various parts of this work. Valuable comments from two anonymous
reviewers helped to clarify and improve this paper. Dr. John Lilibrige (NOAA/NESDIS) provided the GFO radar altimetry data for the period of Katrina and Rita. Jodi Brewster (RSMAS/MPO) processed the satellite-based data. The mooring data were supported by MMS to Evans-Hamilton, Inc. under the OCS Contract 1435-01-04-CT-34239. Rio05 was produced by CLS Space Oceanography Division.

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