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Mixed Layer Cooling in Mesoscale Oceanic Eddies during Hurricanes Katrina and Rita

Benjamin Jaimes and Lynn K. Shay
Rosenstiel School of Marine and Atmospheric Science, Division of Meteorology and Physical Oceanography, University of Miami, 4600 Rickenbacker Causeway, Miami, FL, 33149-1098.

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Corresponding author address: Benjamin Jaimes, Rosenstiel School of Marine and Atmospheric Science, Division of Meteorology and Physical Oceanography, University of Miami, 4600 Rickenbacker Causeway, Miami, FL, 33149-1098.
E-mail: bjaimes@rsmas.miami.edu
ABSTRACT

During favorable atmospheric conditions, hurricanes Katrina and Rita deepened to category 5 over the Loop Current’s (LC) bulge associated with an amplifying warm core eddy. Both hurricanes subsequently weakened to category 3 after passing over a cold core eddy (CCE) prior to making landfall. Reduced (increased) oceanic mixed layer (OML) cooling of \(\sim 1^\circ C\) (4.5\(^\circ \)C) was observed over the LC (CCE) where the storms rapidly deepened (weakened). Data acquired during and subsequent to the passage of both hurricanes indicate that the modulated velocity response in these geostrophic features was responsible for the contrasts in the upper ocean cooling levels. For similar wind forcing, the OML velocity response was about two times larger inside the CCE that interacted with Katrina than in the LC region affected by Rita, depending on the pre-storm OML thickness.

Hurricane-induced upwelling and vertical mixing were increased (reduced) in the CCE (LC). Less wind-driven kinetic energy was available to increase vertical shears for entrainment cooling in the LC, as the OML current response was weaker and energy was largely radiated into the thermocline. Estimates of downward vertical radiation of near-inertial wave energies were significantly stronger in the LC (12.1\(\times\)10\(^{-2}\) W m\(^{-2}\)) compared to that in the CCE (1.8\(\times\)10\(^{-2}\) W m\(^{-2}\)). Katrina and Rita winds provided \(O(10^{10})\) W to the global internal wave power. The vertical mixing induced by both storms was confined to the surface water mass. From a broader perspective, models must capture oceanic features to reproduce the differentiated hurricane-induced OML cooling to improve hurricane intensity forecasting.
1. Introduction

Isotherm topography and energetic geostrophic flow in mesoscale oceanic features in the Gulf of Mexico (GOM) have been shown to impact the efficiency of hurricanes to cool the oceanic mixed layer (OML). For example during hurricane Gilbert’s passage, the hurricane cooled the upper ocean by about 4°C in the Gulf Common Water due principally to vertical shear-induced mixing (Shay et al. 1998; Jacob et al. 2000). More recently, Ivan passage over cold core eddies (CCEs) indicated elevated levels of hurricane-induced cooling where sea surface temperature (SST) changes exceeded 5°C (Walker et al. 2005; Halliwell et al. 2008). These two cases of ocean response are clear examples of negative feedback originally discussed by Chang and Anthes (1978). That is, upper ocean cooling affects the OML and heat fluxes into the tropical cyclone, eventually contributing to the storm's intensity fluctuations (Hong et al. 2000; Jacob and Shay 2003; Shay and Uhlhorn 2008). By contrast, in regions of the Loop Current (LC) and warm core eddies (WCEs), the ocean only cools by about 1°C during hurricane passage (Shay et al. 2000; Shay and Uhlhorn 2008) and there is less negative feedback than commonly assumed in forecast models where the ocean is considered to be uniform and at rest. Tropical cyclones are more likely to reach a larger fraction of their maximum potential intensity over warm oceanic features of the GOM and the Caribbean Sea (DeMaria and Kaplan 1994), where subtropical water with 26°C temperatures extend to more than 100 m depth, providing a continuous source of heat for hurricanes to intensify under favorable atmospheric conditions (Hong et al. 2000; Shay et al. 2000).

The dependence of hurricane-induced OML cooling on the presence of mesoscale oceanic features therefore is a critical issue for hurricane intensity forecasting in the GOM, as at anytime the Gulf has both WCEs and CCEs features (Vukovich 2007). The LC WCEs have a
vertical scale of $O(10^3)$ m and diameters between 200 and 400 km (Mooers and Maul 1998), and the heat added by one of this eddies after separating from the LC is $\sim 7 \times 10^9$ J m$^{-2}$ (Elliot 1982). By contrast, the CCEs have a deep signature of $\sim 800$ m and diameters from 100 to 150 km (Hamilton 1992; Walker et al. 2003; Zavala-Hidalgo et al. 2003). The sharp horizontal thermal, haline and momentum gradients between these robust mesoscale features and the surrounding Gulf Common Water occur over scales of $O(10)$ km with markedly different temperature and salinity structures (Nowlin and Hubertz 1972; Shay et al. 1998). These gradient regimes usually induce the strongest air-sea fluxes during hurricane passage, with values of 1.5 to 2 kW m$^{-2}$ (Hong et al. 2000; Shay and Uhlhorn 2008). This mesoscale variability is often not captured by the satellite-derived near uniform SST distribution that prevails in the GOM during the hurricane season.

The hurricane-induced OML thermal response is principally governed by three processes: (i) sensible and latent heat loss to the tropical cyclone across the air-sea interface, (ii) upwelling of colder thermocline water due to the horizontal divergence of wind-driven OML currents during the forced stage, and (iii) turbulent entrainment of colder thermocline water due to either wind stirring (forced stage) or instability of the vertical shear of forced near-inertial oscillations (relaxation stage). In quiescent ocean regimes, vertical shear-driven entrainment generally accounts for 75 to 90% of the OML cooling (Price 1981; Greatbatch 1984; Shay et al. 1992, 2000; Hong et al. 2000; Jacob et al. 2000). By contrast, air-sea fluxes explain only 5 to 15% of the total OML cooling, while advective tendencies can be comparable to the heat loss across the air-sea interface (Jacob et al. 2000).

Several processes in mesoscale oceanic eddies can however modulate the efficiency for the storm’s wind stress to accelerate horizontal OML currents, affecting the rate of wind-
induced upwelling and vertical entrainment for layer cooling. For instance, it has been speculated that the rate of wind-driven upwelling can be affected by the geostrophic flow, as the wind stress acting on a geostrophic vortex tends to tilt its’ axis away from the vertical, but the strong geostrophic relative vorticity $\zeta_{g}$ allows the vortex to maintain vertical coherence by developing vertical velocities that balance the wind-induced horizontal advective tendency to first order (Stern 1965). Moreover, the density structure and $\zeta_{g}$ associated with the geostrophic features [i.e., shallower (deeper) isopycnals/OML in CCEs (WCEs)] set the amplitude of the OML velocity response (Gill 1984; Zervakis and Levine 1995; Jacob and Shay 2003), and impose a contrasting spatial distribution of allowable near-inertial wave frequencies excited by storms. That is, the geostrophic relative vorticity ($\zeta_{g}$) shifts the lower bound of the internal waveband from the local Coriolis frequency $f$ to an effective Coriolis frequency $f_{e}=f+\zeta_{g}/2$. This broadening of the near-inertial wave passband impacts the accumulation of momentum and vertical shear development in the upper ocean (Kunze 1985). Gill (1984) showed that deeper (shallower) OML increases (reduces) the rate of energy loss due to radiation of internal near-inertial waves from the OML into the thermocline. This energy loss can reduce the rate of OML deepening and entrainment mixing across the OML interface up to 50% (Linden 1975).

This paper evaluates the relative role of the GOM’s pre-storm geostrophic eddies in modulating the OML velocity and thermal response to hurricanes Katrina and Rita. These hurricanes propagated over the LC system during the bulging and shedding of a LC WCE, where several frontal CCEs moved along the periphery of the LC and the bulge. Both hurricanes deepened to category 5 status over the warm LC bulge (where OML cooling was reduced), then subsequently weakened to category 3 hurricane after encountering one of the
frontal CCEs (where significant cooling was observed) and unfavorable atmospheric conditions prior to making landfall.

Airborne-, mooring-, and altimeter-based data are used here (section 2) to describe the contrasting OML velocity and thermal response induced by severe hurricanes inside cyclonic and anticyclonic geostrophic ocean eddies. Both Katrina and Rita experienced rapid intensity changes as they propagated over the strong thermal gradients associated with these geostrophic features (section 3). The approach is to diagnose the three dimensional structure of the hurricane energy source available in the LC system for Katrina and Rita, and to understand the role of the LC cycle (Sturges and Leben 2000) in building up the mesoscale spatial variability encountered by the two hurricanes (section 4). The hurricane-induced temperature and velocity response induced in the upper layers of the LC bulge and CCE that interacted with the two hurricanes are described in section 5 where the foci includes OML deepening, upwelling during the forced stage, near-inertial velocity response and vertical mixing, and radiation of near-inertial internal waves form the OML into the thermocline. Implications of the modulated velocity response and negative feedback for the prediction of OML cooling are discussed in section 6. Results are summarized with concluding remarks in section 7.

2. Methodology and data resources

a. Airborne measurements

The three dimensional upper ocean thermal and salinity structure in the LC system was surveyed with Airborne eXpendable BathyThermographs (AXBT), Current Profilers (AXCP), and Conductivity-Temperature-Depth sensors (AXCTD) deployed from four aircraft flights during September 2005, as part of a joint National Oceanographic and Atmospheric
Administration (NOAA) and National Science Foundation experiment (Rogers et al. 2006; Shay 2008). Flight patterns were designed to sample the mesoscale features in the LC system: the LC bulge (amplifying WCE), the WCE that separated from the LC a few days before the passage of Rita, and two of the CCEs (CCE1 and CCE2) that moved along the LC periphery during the WCE shedding event (Fig. 1).

The first aircraft flight was conducted on 15 September (two weeks after Katrina or one week before Rita, i.e. pre-Rita), the second and third flights were conducted during Rita’s passage (22 and 23 September, respectively), and the final flight was conducted on 26 September, a few days after Rita’s passage (Fig. 1 and Table 1). Pre-Rita and post-Rita (not shown) flights followed the same pattern, while Rita-22 and Rita-23 focused on different regions along Rita's track. Data acquired during pre-Rita includes temperature profilers from AXBTs, temperature and salinity profilers from AXCTDs, and current and temperature profilers from two AXCPs deployed in the western and eastern sides of the WCE (Fig. 1). During the other three flights, only temperature profiles from AXBTs were acquired to ~350 m depth, compared to 1000 m and 1500 m for AXTCDs and AXCPs, respectively. The accuracy of the thermistor is ±0.12°C for AXCTDs (Johnson 1995), and ±0.2°C for AXBTs and AXCPs (Boyd 1987).

To evaluate these temperature profilers from the various sources, one AXBT and one AXCTD were simultaneously deployed inside the WCE during pre-Rita; a second AXCTD was deployed at this site on a subsequent leg of the flight pattern. Temperature profilers from these three probes are consistent where RMS differences are comparable to the accuracy of the AXBT thermistor (Fig. 2a). The water mass distribution from AXCTDs consists of Subtropical Water (STW) in the upper ocean layers, and Sub Antarctic Intermediate Water (SAAIW) in
deeper waters underneath the STW (Fig. 2b), in accord with previously reported observations from ship-borne measurements (e.g., Hofmann and Worley 1986). A salient characteristic of the STW (e.g., WCE) is the salinity maximum of ~36.4 to 36.6 psu in $\sigma_t$ space. This $\sigma_t$ behavior must be incorporated into numerical models, as a climatological salinity profile is insufficient to accurately initialize the model ocean with a WCE. Realistic salinity profiles to match the temperature profiles would then resolve horizontal density gradients and the corresponding geostrophic flows associated with a warm feature eddy (Nowlin and Hubertz 1972; Shay et al. 1998).

b. Mooring data

Minerals Management Service (MMS)-sponsored Acoustic Doppler Current Profilers (ADCP) and CTD moorings were deployed in the northeastern Gulf (Fig. 1 and Table 2). Continuous measurements of temperature, conductivity, and ocean currents were acquired from the mooring sensors at intervals of 0.5 and 1 hr for CTDs and ADCPs, respectively. Although the moorings were located outside the radius of maximum winds $R_{\text{max}}$ of hurricanes Katrina (~4.5 $R_{\text{max}}$ where $R_{\text{max}} = 47$ km) and Rita (~17.5 $R_{\text{max}}$ where $R_{\text{max}} = 19$ km) (Fig. 1), the CCE2 that was affected by Katrina (category 5 status) propagated over the mooring site ≈2 days after interacting with the storm [convex orientation of low-frequency isotherm fluctuations, Fig. 3a, from ~1 to 8 Inertial Periods (IP)]. The anticyclonic circulation of the LC bulge that interacted with Rita (category 5 status) extended over the mooring ≈3 days after having been affected by the storm (concave orientation of low-frequency isotherm fluctuations, Fig. 3b from ~1 to 5 IP). Altimeter-based evidence of the intrusion of the CCE2 and LC over the mooring is presented in section 3.
c. Wind stress

Wind fields used in this investigation are from the NOAA/Hurricane Research Division H*Wind product (http://www.aoml.noaa.gov/hrd/data\_sub/wind.html), which blends wind measurements from a variety of observation platforms into high-resolution objectively analyzed fields of standard 10-m surface winds (Powell et al. 1996). Based on the H*Wind product, wind stress fields for hurricanes Katrina and Rita were derived using a drag coefficient $C_d$ computed from the Large and Pond (1981) relationship, but capped at a maximum value of $2.6 \times 10^{-3}$ based on recent results indicating a saturation value of $C_d$ between 27 to 35 m s$^{-1}$ wind speeds (Powell et al. 2003; Donelan et al. 2004; Shay and Jacob 2006; Jarosz et al. 2007).

d. Geostrophic circulation

As the GOM circulation is dominated by mesoscale features in nearly geostrophic balance with horizontal velocities of approximately 1 m s$^{-1}$ (e.g. Molinari and Morrison 1988; Nowlin and Hubertz 1972), the effects of this energetic geostrophic variability need to be resolved to understand the modulated OML response to hurricane forcing. Deriving geostrophic flow from the shallow AXBTs (~350 m) measurements is, however, not trivial, as the vertical scale of the mesoscale features is ~800–1000 m. Nevertheless, the water mass homogeneity in the GOM beneath the thermocline (or below the 20°C isotherm depth, Fig. 2b), together with the fact that in this region the density is primarily a function of temperature, allows to extend the shorter AXBT temperature profiles to 1000 m by following the next approach:
1. use the reference temperature $T_r$ from the closest (deeper) AXCTD profile to extend the AXBT profile to 1000 m; the original AXBT’s upper thermal structure is preserved in the new profile $T_t$; 
2. use the reference salinity $S_r$ from the closest (deeper) AXCTD together with $T_r$ to determine the optimal polynomial fit to the $T_r$-$S_r$ relationship; 
3. utilize polynomial coefficients from above to get the target salinity $S_t$ in terms of $T_t$; 
4. estimate seawater density $\rho$ in function of $T=[T_r, T_t]$ and $S=[S_r, S_t]$; 
5. analyze $\rho$, $T$, and $S$ via objective technique with the appropriate mapping error (Mariano and Brown 1992); and, 
6. estimate geostrophic velocities relative to an assumed level of no motion at 1000 m.

The water mass distribution obtained with this method is shown in Fig 4a. The upper ocean variability arises from the original profilers. The inferred geostrophic circulation (Fig. 4b) is consistent with observations reported in the literature where the maximum surface velocities are $O(1)$ m s$^{-1}$. The magnitude of the geostrophic relative vorticity in both WCEs and CCEs is of the same order as the local Coriolis frequency ($f=6.39\times10^{-5}$ s$^{-1}$ at 26°N) (Fig. 4b), suggesting an important process for modulation of the near-inertial response (Kunze 1985, 1986).

e. Isotherm depth and sea surface height from altimetry

In situ data are complemented with altimeter-based 20°C and 26°C isotherm depths ($h_{20}$ and $h_{26}$, respectively), estimated with a two-layer approach based on the sum of mean and perturbation isotherm depths (Goni et al. 1996; Shay et al. 2000; Mainelli et al. 2008). Mean fields are from a hurricane season climatology of hydrographic measurements (Mainelli 2000), and perturbation fields are estimated from satellite-based radar altimetry measurements of the
surface height anomaly (SHA) field from NASA Jason-1 and the NOAA Geosat Follow-on-Mission (GFO). The method to produce daily maps of altimeter-based SHA and isotherm depths is described in detail by Mainelli et al. (2008). Altimeter-based isotherm depths capture the observed mesoscale features in the GOM, which agree with those determined from in situ measurements to within ~10% uncertainty, for both $h_{26}$ and $h_{20}$ (not shown). Moreover, daily maps of absolute sea surface height $\eta$ were reproduced for the entire GOM by adding the daily SHA fields to the Combined Mean Dynamic Topography Rio05 (Rio and Hernandez 2004). The correlation between four daily snapshots of $\eta$ and the absolute sea surface height from the 7-day merged AVISO product (http://www.aviso.oceanobs.com/en/data/products/index.html) was between 0.92 to 0.94 for point-wise comparisons over the entire GOM, and for the period from 31 August to 28 September 2005 (see Figs. 5 and 6).

3. Hurricanes Katrina and Rita

a. Hurricane Katrina and mesoscale ocean variability

Tropical storm Katrina emerged over the warm waters of the southeastern GOM at 0500 UTC 26 August, and quickly reached hurricane status at 0600 UTC with maximum sustained winds of 33 m s$^{-1}$. During this time period, the LC was undergoing a complex shedding event of the WCE (Fig. 7a). Weak wind shear dominated the entire GOM basin, and coupled with an efficient upper-level atmospheric outflow facilitated two periods of rapid intensification over the LC between 26 and 28 August (Knabb et al. 2005). Rapid intensification is defined as a 23-24 mb or greater pressure decrease in a 24-h period (J. Kaplan, pers. comm.). The first intensification occurred on 1200 UTC 27 August, when Katrina increased to category 3 status where surface winds exceeded 51 m s$^{-1}$ winds. By 1200 UTC 28 August, Katrina had a second
and more rapid intensification over the LC bulge during an eyewall cycle as it strengthened from a category 3 to a category 5 hurricane in less than 12 h. Late on 28 August, Katrina moved over a frontal cold cyclonic feature (growing CCE2) located on the northwestern edge of the LC (Fig. 7a). Simultaneously, the hurricane underwent eyewall erosion, while another outer ring of convection developed. Eyewall erosion continued early on 29 August, and the hurricane rapidly weakened making final landfall as a category 3 at 1100 UTC. Structural changes during eyewall erosion appear to have dominated Katrina’s rapid weakening before landfall, although colder SST over the CCE, gradually increasing wind shear, entrainment of dry air, and interactions with land could have contributed to this weakening (Knabb et al. 2005).

The amplifying CCE2 that was directly affected by Katrina’s wind stress moved over the mooring site by 31 August and it remained over this point for more than 2 weeks (Fig. 5a–b, e–f). On 21 September, the WCE was completely detached from the LC as the eddy underwent a slow westward propagation (Fig. 5c, g).

b. Hurricane Rita and mesoscale ocean variability

On 20 September 2005, tropical storm Rita approached the Florida Straits with maximum winds of 31 m s$^{-1}$. As observed during hurricane Katrina, the atmospheric environment had weak vertical wind shear (Knabb et al. 2006). By 1200 UTC 20 September, Rita became a hurricane with an intensity of 36 m s$^{-1}$, reaching category 3 intensity early on 21 September as it moved westward over the southeastern GOM. Rita’s wind reached an intensity of 74 m s$^{-1}$ by 1800 UTC, with a peak intensity of 80 m s$^{-1}$ over the LC bulge on 22 September (Fig. 7b). The inner eyewall subsequently collapsed later on 22 September as Rita weakened to
category 4 strength. Steady weakening continued on 23 September as the hurricane moved over the frontal CCE1 with increased southwesterly wind shear causing it to weaken to a category 3 hurricane with 57 m s\(^{-1}\) maximum winds by 1800 UTC. This intensity level persisted until the time of landfall, which occurred at 0800 UTC 24 September with an estimated intensity of 51 m s\(^{-1}\) (Knabb et al. 2006). On 28 September, the LC bulge that previously interacted with Rita (category 5 status) extended over the mooring array (Fig. 5d).

4. Loop Current structure

a. Isotherm topography

The LC cycle (Sturges and Leben 2000) is the predominant dynamical process that determines isotherm topography due to advection of warm STW in the eastern GOM through the Yucatan Straits, and by horizontal convergence of mass during WCE formation. These characteristics delineated the thermal structure over which Katrina and Rita moved. As shown in Fig. 7c, by March 2005 the LC started to grow as warm STW from the northwest Caribbean Sea was entrained into the current's bulge. The layer thickness with waters warmer than 26°C gradually increased, attaining peak \(h_{26}\) values between mid August and September. During this period, Katrina and Rita moved over the LC where \(h_{26}\) supported oceanic heat content levels of more than 100 kJ cm\(^{-2}\) relative to the 26°C isotherm depth (Mainelli et al. 2008; Shay 2008), as observed during hurricanes Opal (Shay et al. 2000), and Isidore and Lili (Shay and Uhlhorn 2008).

b. Ocean thermal structure and storm intensity changes
On 28 August, Katrina rapidly deepened to a category 5 hurricane (with an estimated maximum surface wind stress of 7.6 N m$^{-2}$) as it moved at a speed of ~6 m s$^{-1}$ along the LC’s western flank, over a lobe-like structure where peak $h_{26}$ depths were ~110 m (Fig. 8a). The corresponding oceanic heat content values in this warm feature were ~120 kJ cm$^{-2}$, or more than five times the threshold of ~17 kJ cm$^{-2}$ d$^{-1}$ needed to sustain a hurricane (Leipper and Volgenau 1972). Prior to landfall, Katrina crossed over the shallower isotherms of a growing CCE where $h_{26}$ depths were ~40 m (CCE2, Fig. 8a). The presence of this cooler feature contributed to the rapid weakening of Katrina prior to landfall (Knabb et al. 2006).

Three weeks later, Rita translated over the GOM. While Rita's path did not exactly follow Katrina's in the south-central part of the basin, Rita moved over the LC bulge at a speed of ~5 m s$^{-1}$ and rapidly deepened where the maximum wind stress exceeded 8 N m$^{-2}$. On 23 September, Rita moved over the eastern tip of the WCE and then began a weakening period due in part to an eyewall-replacement cycle (Knabb et al. 2006) and the previous interaction with the intensifying CCE1 located between the WCE and LC (Fig. 8b). Comparison of pre-Rita and post-Rita observations (Fig. 8c, d) reveals significant oceanic surface cooling of more than 4°C over this strong frontal regime where Rita interacted with the CCE1.

To illustrate this interaction, along-track surface pressure fluctuations are compared to along-track $h_{26}$ and SST variations. The longitude, latitude, and pressure from the National Hurricane Center best-track 6-h files were interpolated to estimate storm position and pressure at 2-hour intervals (Fig. 9). Altimeter-based $h_{26}$ fields were used to obtain an averaged value at 2-h storm positions using 9 points from -0.5° to 0.5° relative to the best-track position from -2h to 2h. Uncertainty bars represent the standard deviation estimated at each 6-hourly position based on this running 9-point averaging scheme. Generally, the normalized $h_{26}$ values vary.
inversely to pressure changes. That is, as surface pressure decreases, the $h_{26}$ tendency has an upward trend that suggests an impact to the hurricane intensity. The high variability of $h_{26}$ over relatively short distances induced large oceanic heat content gradients, which have been shown to impact the surface enthalpy fluxes due to wind speed variations over thermal gradients (Shay and Uhlhorn 2008). By contrast, along-track SSTs were essentially flat during the life cycle of Katrina and Rita within the GOM (Sun et al. 2006), and did not reveal the mesoscale variability of the LC system. In this case, radar-altimeter derived products were more closely related to intensity variations than SSTs (cf. Scharroo et al. 2005; Shay 2008).

5. Upper ocean response

a. OML cooling

In the case of hurricane Rita, the OML temperature response was differentiated along the storm's track with reduced cooling $\Delta T < 1^\circ$C in the region where Rita deepened to category 5 hurricane (LC bulge, Fig. 10a), and increased cooling $\Delta T \sim 4$ to $5^\circ$C over the region where Rita started to weaken (shedding front between the WCE and LC, Fig. 10b). Similar values of reduced OML cooling over GOM’s warm features have been documented elsewhere (Shay et al. 2000; Shay and Uhlhorn 2008), while increased OML cooling of $O(3$ to $7) ^\circ$C was observed during Ivan’s interactions with CCEs (Walker et al. 2005; Halliwell et al. 2008).

b. Upwelling during the forced stage

The comparison of pre- and in-storm temperature profiles for the case of Rita indicates that the storm’s wind stress induced downwelling of isotherms over the LC bulge, and upwelling over the region dominated by the CCE1 (Fig. 10). To investigate the effects of the
underlying geostrophic eddies on this contrasting upwelling velocity response, consider the vorticity balance in the OML resulting from the interaction of the wind stress with a geostrophic vortex, which to the lowest order is given by (Stern 1965):

\[
\frac{\partial}{\partial z} \mathbf{k} \cdot \nabla \theta_z + f \frac{\partial \mathbf{W}_b}{\partial z} = \mathbf{V}_b \cdot \nabla \zeta_g,
\]

where the first and second terms in the left hand side represent the turbulent stress and vortex stretching, respectively, while the term on the right hand side is advection of \( \nabla \zeta_g \) by the frictional velocity \( \mathbf{V}_b \) (\( \zeta_g \) is the vertical component of the geostrophic relative vorticity). After integrating from \( z = -h \) (bottom of the OML) to \( z = 0 \) (sea surface), this equation becomes

\[
\frac{1}{f} \left( \frac{\partial \tau_{xy}}{\partial x} - \frac{\partial \tau_{sx}}{\partial y} \right)_0 + w_b(0) = \frac{1}{f} \int_{-h}^{0} \mathbf{V}_b \cdot \nabla \zeta_g.
\]

In this application of Stern’s (1965) theory, \( \theta_s(z = 0) \equiv \mathbf{\tau}_s \) is the wind stress vector. By using the undisturbed Ekman relation

\[
\int_{-h}^{0} \mathbf{V}_b \cdot d\mathbf{z} = -\frac{1}{\rho_0 f} \mathbf{k} \times \mathbf{\tau}_s,
\]

the vorticity balance becomes

\[
\frac{1}{f} \left( \frac{\partial \tau_{xy}}{\partial x} - \frac{\partial \tau_{sx}}{\partial y} \right)_0 + w_b(0) = -\frac{k \times \mathbf{\tau}_s}{\rho_0 f^2} \cdot \nabla \zeta_0,
\]

or,

\[
w \equiv -w_E - w_b(0) = \frac{k \times \mathbf{\tau}_s}{\rho_0 f^2} \cdot \nabla \zeta_0, \quad (1)
\]

where \( w = -w_E - w_b \) is the total vertical velocity, \( w_E \) the vertical velocity due to Ekman pumping, and \( \mathbf{\tau}_s = \tau_{sx} \mathbf{i} + \tau_{sy} \mathbf{j} \), with

\[
\tau_{sx} = \rho_a C_d |U_{10}| U_{10}, \quad \tau_{sy} = \rho_a C_d |V_{10}| V_{10}, \quad (2)
\]

where \( \rho_a \) is the air density, \( U_{10} \) and \( V_{10} \) are the zonal and meridional 10-m wind components, \( |U_{10}| = \left( U_{10}^2 + V_{10}^2 \right)^{1/2} \), and \( C_d \) is the surface drag coefficient. For a saturation level defined as in
Powell et al. (2003), maximum values of $\tau_{sx} = \tau_{sy} \sim 3.8$ Pa were obtained with $C_d = 2.6 \times 10^{-3}$ and $U_{10} = V_{10} = 35$ m s$^{-1}$ from the H*Wind product.

The upwelling velocity ($w$) induced by Rita winds over geostrophic eddies is estimated in terms of Eq. (1) (Fig. 11). The approach is to project time-dependent wind stress fields for this hurricane over $\zeta_g$ from the pre-Rita flight (15 September). Thus, the assumption here is that the geostrophic features were stationary between September 15–23. This assumption is reasonable valid, as the features on which this investigation is focused (LC bulge and CCE2) do not change position significantly during this time period relative to hurricane wind fields (Fig. 5). However, the main point of this calculation is to highlight the sensitivity of the upwelling velocity to the projection of the wind stress vector on the oceanic geostrophic current vector, which might explain the observed contrasting upwelling regimes discussed here and elsewhere (Halliwell et al. 2008).

Compared to hurricane-induced upwelling over an ocean initially at rest (O’Brien 1967; O’Brien and Reid 1967), in the presence of geostrophic flow the vertical velocity $w$ induced by Rita according with Eq. (1) did not show a continuous upwelling maximum underneath the storm’s center, as $w$ was a strong function of the projection of the wind stress vector on the geostrophic flow [Eq. (1)]. Upwelling (downwelling) processes can typically be expected when $\tau_s$ is with (against) the geostrophic flow (Fig. 11). The orientation of the wind stress vector along the right side of the storm’s track (and within $2R_{max}$) in general coincided with the orientation of the geostrophic flow over the LC frontal zone that separated cyclonic and anticyclonic circulations (Fig. 11b). Over this region, Rita weakened from category 5 to 4 at the same time that upwelling of cooler water persisted with values of $\sim 1$ cm s$^{-1}$, consistent with upwelling velocities of about 1 cm s$^{-1}$ found during hurricanes Gilbert (Shay et al. 1998) and
Ivan (Teague et al. 2007). Observed pre- and in-storm temperature profiles (Fig. 10b) support the presence of this upwelling region along the LC front. Wind-driven downwelling velocities of approximately 1 cm s\(^{-1}\) were calculated over the WCE that was directly underneath Rita’s eye by 23 September (Fig. 11c,d) since the wind stress and geostrophic flow were anti-correlated. The comparison of the observed pre- and in-storm temperature profiles supports this downwelling over WCEs (Fig. 10a). Notice that this average upwelling/downwelling extended over broad regions ahead of the storm’s eye, indicating important SST cooling before the arrival of the storm’s center (D’Asaro 2003; Jacob and Shay 2003).

c. Near-inertial velocity response and vertical mixing

The contrasting hurricane-induced near-inertial response is a function of the underlying geostrophic flow as observed at the mooring locations. The background cyclonic eddy (CCE2) that interacted with Katrina (28 August) moved over the mooring site by 31 August. The anticyclonic circulation of the underlying LC affected by Rita (22 September) intruded over the mooring by 28 September (Fig. 5). The velocity response inside the CCE2 consisted of strong cross-track currents of approximately 80 cm s\(^{-1}\) rotating anticyclonically with depth, which is suggestive of strong vertical current shear (Fig. 12a). Similar near-inertial properties were observed in the northern portion of the LC (Fig. 12b), although in this case the velocity response and vertical shears were considerably weaker with maximum velocities of ~40 cm s\(^{-1}\). Despite differing velocity responses, the CCE2 and LC bulge were under the action of comparable wind forcing because Katrina and Rita surface winds were about 70 m s\(^{-1}\) when they impacted these eddies (Fig. 7a,b).
The evaluation of water mass at the mooring site reveals that most of the upper ocean cooling was driven by forced near-inertial processes, both inside the CCE2 (Fig. 13a) and LC bulge (Fig. 13b). Given that the near-inertial OML velocity response was stronger inside the CCE2, the vertical current shear caused more intense mixing within this oceanic cyclone. By contrast, mixing and cooling processes were reduced in the LC as both the velocity response and vertical shears were weaker, consistent with measurements in the LC during Isidore and Lili (Shay and Uhlhorn 2008). Notice that in the two cases, the shear-induced mixing was confined to neutral surfaces lighter than $\sigma_t = 27$ (within STW).

The coupling between the translation speed of the storm $U_h$ and the phase speed of the first baroclinic mode $c_1$ determines whether the upper ocean response is in the form of upwelling or a near-inertial wave wake (Geisler 1970, Nilsson 1995). To evaluate the contrasting near-inertial velocity response excited by analogous wind stress over WCEs and CCEs, consider a two-layer approach in which $c_1$ is given by

$$c_1^2 = \frac{g}{\rho_{dn}} \frac{(\rho_{dn} - \rho_{20})h_{20}h_{dn}}{\rho_{dn}(h_{20} + h_{dn})},$$

where $h_{20}$ is the $20^\circ$C isotherm depth (proxy for the thermocline in the GOM), $h_{dn}$ is the thickness of the layer extending from $h_{20}$ down to 1000 m, and $\rho_{20}$ and $\rho_{dn}$ are vertically-averaged densities upon $h_{20}$ and $h_{dn}$, respectively. Rather than the density difference between the two layers (stratification), the variability of $c_1$ in the eastern GOM was primarily a function of $h_{20}$, which was about 200 to 250 m in the LC and 100 to 120 m in CCE2. The maximum values of $c_1$ from this thermal structure were ~2.9 m s$^{-1}$ in the LC and 2.1 m s$^{-1}$ in CCE2 (Table 3).

As Katrina and Rita translation speeds were faster than $c_1$, their respective Froude numbers (defined as $Fr = U_h c_1^{-1}$) exceeded unity (Table 3), indicating a baroclinic response.
driven by near-inertial currents (Geisler 1970), consistent with the velocity response observed at the mooring site. Fr is larger in CCE2 because $c_I$ is slower than in the LC. In addition, the thermocline topography (and $c_I$ variability) determines the wave power generated by the curl of the wind stress field, and this power is always reduced when $c_I(2fR_{\text{max}})^{-1}$ is increased (Nilsson 1995). For the same $R_{\text{max}}$, the wind stress curl is more effective in generating internal near-inertial waves in CCEs than in WCEs (i.e. LC). This indicates that the spatial variability of $c_I$ contributes to the differentiated near-inertial velocity response observed in WCEs and CCEs during Katrina and Rita.

d. Radiation of near-inertial internal waves from the OML into the thermocline

Another important process of the upper ocean response to hurricane forcing is the downward flux of internal near-inertial wave energy from the OML into the thermocline, as this flux is an energy sink of turbulent kinetic energy for the OML, and it is a function of the stratification underneath the OML base (Linden 1975; Gill 1984; Nilsson 1995). Brooks (1983) estimated the vertical energy flux $F_{iw}$:

\[ F_{iw} = C_{g_z} E_d, \quad (3) \]

where $C_{g_z}$ is the vertical component of the group velocity vector, and $E_d$ the layer kinetic energy density at the time of maximum near-inertial wave response. For a constant buoyancy frequency the group velocity vector is $\mathbf{Cg} = C_{g_x} \mathbf{i} + C_{g_z} \mathbf{k}$, with $\mathbf{i}$ and $\mathbf{k}$ the unity vectors in the cross-track and vertical direction, respectively, and $C_{g_x}$ and $C_{g_z}$ given by

\[ C_{g_x} = \omega_i k^{-1} (N \omega_i^{-1} \tan \theta)^2, \quad (4a) \]
\[ C_{g_z} = \omega_i m^{-1} (N \omega_i^{-1} \tan \theta)^2, \quad (4b) \]
where \( \mathbf{k} = k_i + m \mathbf{k} \) is the wavenumber vector with \( k = 2\pi / \lambda_x \) and \( m = 2\pi / \lambda_z \); \( \omega_i \) is the near-inertial wave frequency; \( \overline{N} \) is a constant buoyancy frequency vertically-averaged over a water column of thickness \( \Delta z \) that excludes the OML; \( \tan \theta = \lambda_z / \lambda_x \) where \( \lambda_x = 4R_{\text{max}} \) and \( \lambda_z = \omega_i \Delta z \Delta t_z \) are the wavelengths in the cross-track and vertical direction, respectively; \( \Delta z \) the distance traveled by a wave (measured perpendicular to the crest line) in the vertical direction, and \( \Delta t_z \) the phase lag between the two depth levels (\( \Delta z \)). The layer kinetic energy density is:

\[
E_d = \frac{1}{2} \rho \omega^2,
\]

where \( \omega \) is the cross-track maximum near-inertial velocity response (Brooks 1983).

Current measurements from the ADCP mooring (point M, Fig. 1) were used to calculate the near-inertial wave parameters summarized in Table 4. The thickness of the layer used during this analysis was \( \Delta z = 425 \) m, and the upper and lower depth levels were taken as 75 and 500 m, respectively. Time series of cross-track velocity from these depth levels were then used to calculate time-averaged phase lags (\( \Delta t_z \)) of 8 and 12.2 hrs for the CCE2 and LC bulge, respectively. A least-squares frequency fit analysis conducted over the time span when the CCE2/LC intruded over the mooring site indicated that near-inertial response was shifted toward higher (lower) frequencies inside the CCE2 (LC bulge). In average, the corresponding near-inertial periods were 23.7 and 24.3 hrs for the CCE2 and LC, respectively. Given that the AXBTs provide a higher vertical resolution of the thermal structure, a profile from the pre-Rita flight was used to calculate a vertically-averaged buoyancy frequency \( \overline{N} = 4.7 \) c.p.h for the CCE2, and a profile from the post-Rita flight was used to calculate \( \overline{N} = 6 \) c.p.h inside the LC bulge. Characteristic velocity responses inside the CCE2 (0.6 m s\(^{-1}\)) and the LC (0.3 m s\(^{-1}\)) were obtained from data shown in Fig. 12.
Inside the CCE2, the vertical wavelength and group velocity were 144 m and 9 m day\(^{-1}\), respectively (Table 4). In the LC, these values were 214 m and 233 m day\(^{-1}\). The corresponding near-inertial wave fluxes were between \(1.8 \times 10^{-2}\) and \(12.1 \times 10^{-2}\) W m\(^{-2}\) in the CCE2 and LC, respectively. That is, the near-inertial downward energy flux into the thermocline was significantly stronger in the LC than in the CCE2. This result is consistent with the results of Gill (1984), who found that the rate of loss of energy from the OML due to internal wave radiation increase with OML depth (as in the LC). Kunze (1986) reported vertical wavelengths similar to those in the LC bulge (~250 m) for near-inertial waves inside a Gulf Stream WCE.

Given that not all the hurricane power injected by Katrina and Rita into the upper ocean was used to cool the OML, it is of interest to estimate the contribution to the global internal wave power from the near-inertial wave energy flux \(F_{iw}\) induced by these storms. As shown in Table 5, depending on the area of the hurricane’s wake, and whether the near-inertial waves propagate inside positive or negative geostrophic relative vorticity features, the contribution to the global internal wave power ranges between \(0.3 \times 10^{10}\) to \(89 \times 10^{10}\) W, which is consistent with the estimate of \(\sim 10^{10}\) W of Nilsson (1995), and with values of \(10^{11}\) W found by Shay and Jacob (2006) using velocity profiles in the wake of hurricane Gilbert. These results are 3 to 5 orders of magnitude smaller than the value of \(O(10^{15})\) W calculated by Emanuel (2001).

6. Discussion

a. OML cooling levels

The lack of systematic and comprehensive ocean measurements prevents incorporating the realistic state of the ocean in coupled hurricane models. Hence, an ocean initially at rest and constant OML thickness \(h\) are commonly assumed in these models. To evaluate the
implications of this practice, consider the OML velocity response $u_s$ that can be expected from a storm with a constant translation speed $U_h$ and wind stress $\tau_s$ at the saturation level, which to the lowest order is given by

$$\frac{\partial u_s}{\partial t} \sim U_h \frac{\partial u_s}{\partial x} = \frac{1}{\rho_o} \frac{\partial \tau_s}{\partial z},$$

which can be parameterized as (adapted from Price 1983):

$$u_s(x, y) = \frac{R_{\max} \tau_s}{\rho_o h(x, y) U_h}, \quad (5)$$

where $\tau_s = |\tau_s|$ is the magnitude of the wind stress vector as per Eq. (2). Notice that Eq. (5) indicates that for constant $U_h$ and $\tau_s$ the OML velocity response is only a function of the pre-storm mixed layer thickness $h(x, y)$, as less (more) work is required to accelerate a shallow (deep) OML current.

For example, by considering the values of $R_{\max}$ and $U_h$ (Table 3), and $h(x, y)$ from the pre-Rita data (15 September), the parametric OML velocity response is about two times larger over the shallower mixed layer of CCEs than over WCEs for similar wind conditions (Fig. 14). The parametric velocity response $u_s$ inside the CCE2 (Katrina) and LC bulge (Rita) is comparable to that observed in upper layers at the mooring site (i.e., $\sim$80 cm s$^{-1}$ and $\sim$40 cm s$^{-1}$, respectively, Fig. 12). This result indicates that, to the lowest order, the amplitude of the OML velocity response is primarily a function of the pre-storm layer thickness (cf. Zervakis and Levine 1995; Jacob and Shay 2003). An OML velocity response of $\sim$35 to 40 cm s$^{-1}$ was observed in deep OMLs of LC warm features during hurricanes Isidore and Lili in 2002 (Shay and Uhlhorn 2008).

\[ b. \textit{Negative feedback} \]
The spatial variability of both $h_{26}$ and the hurricane-induced upwelling velocity in presence of geostrophic flow add an important thermodynamic constraint on hurricane intensity, as more (less) work is required over WCEs (CCEs) to shut down the hurricane energy source and decrease the surface heat and moisture fluxes. For example, in the case of hurricane Rita, the storm experienced a rapid weakening over a 6 hr period (from 00 to 06 UTC 23 September) after passing over the LC frontal zone where upwelling velocities ranged from 1 to 2 cm s$^{-1}$. The estimated time $T_{nf}$ for the surfacing of $h_{26}$ over this region was ~2 to 3 hr (Fig. 15). At a constant translation speed of 4.7 m s$^{-1}$, Rita traveled a distance smaller than $2.5R_{\text{max}}$ over a 3-hr interval, indicating that $h_{26}$ may have reached the sea surface before the storm left the maximum upwelling regime. This suggests that the rapid surfacing of waters colder than 26$^\circ$C accelerated the negative feedback along the LC frontal zone. Given that $h_{26}$ was located at ~60 to 90 m depth along this frontal zone, isothermal displacements were $O(60)$ m, consistent with the displacements observed during hurricane Ivan (2004) over a similar ocean configuration (Walker et al. 2005).

In this context, the energy PE required to adiabatically reduce SST to a level in which the negative feedback is initiated over CCEs, Gulf Common Water, and WCEs, is respectively 2, 6, and $15 \times 10^{17}$ J (Fig. 16). To estimate the time required $\Delta T_{nf}$ for hurricanes Katrina and Rita to initiate this negative feedback, we assume (i) an ocean initially in geostrophic balance, (ii) stationary mesoscale eddies relative to hurricane passage, (iii) the momentum flux excites ageostrophic OML currents (no energy lost to dissipation, internal wave radiation into the thermocline, or to strengthen the background flow), (iv) the momentum flux into the OML is bounded by the saturation level of the sea state, and (v) the kinetic energy of the hurricane-induced ageostrophic currents is used for adiabatic lifting of isotherms (conversion to available
potential energy) such that the rate of energy transfer from the storm to the OML (hurricane work \( h_w \)) can be defined as:

\[
h_w = \int_A \frac{1}{2} \rho_s u_s^2 U_h \, dA, \quad (6)
\]

where \( u_s \) is the OML ageostrophic velocity response, and the term \( \rho_s u_s^2/2 \) is the kinetic energy injected by the storm into the OML currents. The hurricane work \( h_w \) on the OML is proportional to the product of the translation speed of the storm \( U_h \) and the square of \( u_s \) (Geisler 1970; Nilsson 1995). For slow moving or stationary storms \((U_h \to 0, \text{ or } \text{Fr} = U_h c_1^{-1} < 1)\), the product \( u_s^2 U_h \) should be replaced by \( u_s^3 \) in Eq. (6). Thus, the time required for the initiation of the negative feedback is given by \( \Delta T_{nf} = PE h_w^{-1} \), with potential energy defined in the caption of Fig. 17.

For storms similar to Katrina or Rita propagating over CCEs, the negative feedback can start anytime during the first few hours of wind action even with tropical storm force (Fig. 17) (consistent with \( T_{nf} \) in Fig. 15). By contrast, over WCEs the negative feedback could be initiated only if a series of storms with characteristics similar to Katrina propagated over the warm features during a period of 2 days, or during 10 to 14 days for conditions comparable to Rita. The storm intensity changes observed in Fig. 17 indicate that both Katrina and Rita reached hurricane category 5 over regions of reduced negative feedback, while they weakened over regions with negative feedback (i.e., Gulf Common Water).

Given the assumptions made during the derivation of \( h_w \) [Eq. (6)], the values of \( \Delta T_{nf} \) represent an upper bound of the negative feedback, as some of the ignored processes, such as diabatic mixing, horizontal advection of thermal structure, and OML kinetic energy lost to dissipation and internal wave radiation will impact these estimates. However, the spatial
variability of $\Delta T_{nf}$ underscores the importance of the presence of mesoscale oceanic eddies for oceanic feedback mechanisms to hurricanes.

7. Summary and Concluding Remarks

Rapid hurricane intensity changes were observed during Katrina and Rita passages over mesoscale ocean features in the eastern GOM. Data acquired during both storms indicated that rather than with SST, the storm’s sea-level pressure decreases correlated better with the 26°C isotherm depth and oceanic heat content, which exhibited spatial variability associated with the presence of WCEs and CCEs as found during hurricane Ivan (Walker et al. 2005; Halliwell et al. 2008). Reduced OML cooling (~1°C) was measured over the LC bulge where both storms reached category 5 status, and increased OML cooling (~4 to 5°C) was observed over the cold cyclones where the storms weakened to category 3 status prior to landfall in the northern Gulf coast. More research is needed to evaluate the relative contribution of mesoscale oceanic features to hurricane intensity fluctuations, compared with atmospheric processes.

The observed velocity response to Katrina inside the CCE2 was ~80 cm s$^{-1}$, while maximum velocities of about 40 cm s$^{-1}$ were measured inside the LC bulge that interacted with Rita, similar to direct velocity measurements during hurricanes Isidore and Lili (Shay and Uhlhorn 2008). A parametric OML velocity response from the forced equations of motion predicted velocities comparable to those measured in the upper layers of the LC and CCE2. This parametric approach indicated that the amplitude of the OML current response was mainly a function of the pre-storm OML depth associated with the geostrophic features (Jacob and Shay 2003). The oversimplified use of homogeneous OML distributions compromises the
predictions of hurricane-induced OML cooling that impact storm’s intensity in numerical forecast models.

Pre-storm geostrophic flow can impose important dynamical constraints on the wind-driven vertical velocity. Based on a simplified vorticity balance (Stern 1965), the hypothesis here is that upwelling regimes occur in regions where the wind stress vector is with surface geostrophic flows, while downwelling regimes occur over regions where the wind stress vector is against surface geostrophic currents. In this context, upwelling velocities of ~0.5 cm s$^{-1}$ were calculated for regions more than $2R_{\text{max}}$ ahead of the storm’s center for Rita, in agreement with previous studies (D’Asaro 2003; Jacob and Shay 2003).

During the relaxation stage (delayed response with no implications on storm intensity), the near-inertial velocity response and vertical shear-driven mixing were stronger (weaker) inside the CCE2 (LC bulge) that interacted with Katrina (Rita). A non-linear dependence between surface winds and the near-inertial internal wave flux was observed for both storms, consistent with an analytical function developed by Shay and Jacob (2006). The contribution of Katrina and Rita winds to the global internal wave power of $O(10^{10})$ W is limited compared with values estimated for the global energy flux from barotropic to internal tides (~$1.3 \times 10^{12}$, Sjöberg and Stigebrant 1992). These results do not support the hypothesis (Emanuel 2001) that hurricanes significantly drive the Meridional Overturning Circulation. The near-inertial wave wake of Katrina and Rita over mesoscale oceanic eddies is currently under investigation to incorporate the effects of geostrophic relative vorticity on the dispersion properties of hurricane-induced near-inertial waves.

The dependence of the hurricane-induced OML cooling on the presence of mesoscale oceanic eddies is a critical issue for hurricane intensity forecasting in the GOM, as at anytime
the Gulf is populated by both WCEs and CCEs. This mesoscale variability is often not captured by the satellite-derived nearly homogeneous SST distribution that prevail in the GOM during the hurricane season, and that is used as the hurricane energy source in the forecasting models. Surface pressure decreases are more correlated with the $h_{26}$ and ocean heat content variability as opposed to the relatively flat SSTs. Given the relative scales of the LC, WCEs, and CCEs, models must reproduce this pre-storm variability to predict the oceanic response to atmospheric forcing, and more importantly the atmospheric response to oceanic forcing. Thus, more in situ and satellite-based measurements are needed to not only capture the realistic position and structure of mesoscale ocean features for assimilation into numerical models, but also to evaluate the oceanic forcing on atmospheric processes.

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Fig. 15. Along-track vertical velocity $w$ [Eq. (1)] induced by hurricane Rita over the geostrophic flow in the LC system (along the storm’s best-track in Fig. 11). The sea level pressure lines are segments from the pressure lines in Fig. 9. Solid (dashed) line in $w$ curves indicates underlying cyclonic (anticyclonic) geostrophic relative vorticity. The $T_{nf}$ curve stands for the time required for the 26$^\circ$C isotherm to reach the sea surface in function of $w$ and $h_{26}$ (see text for details).

Fig. 16. Area-integrated energy, $PE = \frac{1}{2} \int_{A} \rho_o g h_{iso}^2 dA$ required to reduce SST respect to 30$^\circ$C in a 1$^\circ$×2$^\circ$ box by adiabatic lifting of isotherms over a vertical distance $h_{iso}$ (from airborne pre-Rita data, 15 September). For example, a 1$^\circ$C (4$^\circ$C) SST reduction means that the 29$^\circ$C (26$^\circ$C) isotherm was lifted adiabatically to the sea surface against the gravity acceleration. The rectangular area represents observed SST cooling ranges in the GOM during hurricanes Ivan (2004), and Katrina and Rita (2005). The upper limits of this rectangle are determined by the intersection of the average SST cooling observed during hurricane Ivan (~6.5$^\circ$C) with the $PE$ curve of the CCE2. The intersection of the $PE$ curves with the negative feedback line indicates that about 2, 6, and $15 \times 10^{17}$ J are required to adiabatically reduce SST to 26$^\circ$C and initiate the negative feedback over CCEs, Gulf Common Water (GCW), and LC/WCEs, respectively.
Fig. 17. Time required, $\Delta T_{nf} = PE h_w^{-1}$, for the initiation of the negative feedback for Katrina (a) and Rita (b), with $PE = \frac{1}{2} \int h_{26}^2 dA$, where $h_{26}$ is the $26^\circ C$ isotherm depth, $dA$ the grid size of the objectively analyzed map, and $h_w$ from Eq. (6). In (a) and (b), $h_{26}$ is from pre-Rita (15 September) and post-Rita (26 September) airborne data, respectively. The contour intervals are: 2 hrs from 0 to 10, and 20 hrs from 20 to 340. See caption of Fig. 16 for the meaning of $PE$. 
FIG. 1. Airborne profilers deployed in September 2005 relative to the track and intensity of hurricanes Katrina and Rita (colored lines, with color indicating intensity as per the legend) over the LC System. The light-gray shades on the sides of the storm tracks represent twice the radius of maximum winds \( R_{\text{max}} \). The contours are envelopes of anticyclonic (solid: WCE and LC) and cyclonic (dashed: CCE1 and CCE2) circulations. A set of AXBTs (not shown) was deployed after hurricane Rita (26 September), following a sampling pattern similar to pre-Rita (or post Katrina) (15 September). See Tables 1 and 2 for a description of collected data and legends. Point M indicates the position of MMS moorings used during this study, and Point C represents the drop site for profiler comparison (AXBT versus AXCTD). The horizontal line along 27°N indicates the extent of vertical sections discussed in the text.
Fig. 2. Evaluation of airborne profilers performance during the pre-Rita flight (15 September).
(a) Comparison of three drops (two AXCTDs and one AXBT) inside the WCE (point C, Fig. 1).
(b) Water mass distribution in the LC System from AXCTDs: Subtropical Water (STW), and Subantarctic Intermediate Water (SAAIW).
FIG. 3. Evolution of isotherms at the mooring site (point M, Fig. 1) from CTD measurements. (a) Katrina, and (b) Rita. The vertical lines indicate the time of closest approach of the hurricane's eye to the mooring site (29 August for Katrina, and 23 September for Rita). IP stands for inertial period.
Fig. 4. (a) Water mass distribution in the LC System during 15 September (between the passage of hurricanes Katrina and Rita), from reference AXCTDs and extended AXBTs and AXCPs (see text for details). (b) Vertical section of the geostrophic circulation of both the WCE and CCE2 at 27°N (horizontal line in Fig. 1), derived from the density field in (a); the color scale is for geostrophic relative vorticity $\zeta_g$ (positive is cyclonic, and negative is anticyclonic), normalized by the local Coriolis frequency $f=6.62\times10^{-5}$ s$^{-1}$; contours are for the meridional geostrophic velocity, with positive (negative) values indicating northward (southward) velocity.
Fig. 5. Comparison of 1-day $\eta$ (left panels) and 7-day AVISO (right panels) (see section 2e for details). Color scale is altimeter-based absolute dynamic sea surface height (SSH). The black circle in (c, g) is the position of the storm’s center. Point M represents the mooring used in this study.
Fig. 6. Comparison of 1-day $\eta$ and 7-day AVISO (from Fig. 5). The scatter plots in (a)–(d) are from point-wise comparisons between the two SSH products, conducted over a window from 94°W–85°W and from 22°N–29°N. The correlation coefficients were 0.92, 0.94, 0.92, and 0.94 for (a), (b), (c), and (d), respectively. In general, 7 to 9% of the data were not considered in the individual analysis as they exceeded the range of three standard deviations. Histograms of the difference between the two products [(e)–(h)] were calculated based on data from the scatter plots.
**Fig. 7.** Mesoscale ocean variability during Katrina and Rita. Color scale in (a, b) is altimeter-based absolute dynamic SSH from 1-day η, and circular magenta contours stand for standard 10-m wind speed from the NOAA H*Wind product. The external, intermediate, and inner wind circles in (a, b) are the lower limit of tropical storm winds (18 m s⁻¹), winds at saturated level (28 m s⁻¹), and category 1 hurricane winds (33 m s⁻¹), respectively. (c) Loop Current cycle from altimeter-based $h_{26}$ ($h_{26}$ is area-averaged within a box from 93–81°W and 22–28.5°N in the Loop Current region).
FIG. 8. Thermal structure in the LC system before (15 September: left panels) and after (26 September: right panels) the passage of hurricane Rita, from airborne profilers. (a) and (b) are the $h_{26}$ topography; the color shade shows regions with mapping error less than 40% from the objective analysis technique. (c) and (d) are zonal vertical sections of temperature across 26°N [indicated by arrows in (a) and (b)]. Color in the best-track lines stands for storm intensity as per the legend.
Fig. 9. Along-track conditions during hurricanes (a) Katrina and (b) Rita. The time series were constructed with conditions at the actual position of the storm. $h_{26}$ is altimeter-based and normalized by 60 m, while SST is normalized by 30°C. The error bars in $h_{26}$ indicate variability within a half longitudinal degree on each side of the best-track. The maxima in $h_{26}$ in (a) and (b) are associated with the LC bulge, while the minimum $h_{26}$ in (b) is associated with CCE1.
Fig. 10. Upper-ocean temperature changes induced by hurricane Rita: (a) LC bulge, (b) cyclonic circulation of the growing CCE1 (between the WCE and LC, ~90°W, 26.5°N). The temperature profiles represent cluster-averaged values from airborne data. Pre-, in-, and post-storm data are from September 15, 22, and 26, respectively.
Fig. 11. Vertical velocity \( w \) [Eq. (1)] induced by hurricane Rita over the geostrophic flow of the LC system. Positive (negative) values indicate upwelling (downwelling). Black arrows are wind stress vectors capped at the saturation level [Eq. (2)] from the NOAA’s H*Wind product. Red arrows are vectors for the oceanic geostrophic flow field, vertically averaged upon the OML. Solid (dashed) red contours are geostrophic flow lines of cyclonic (anticyclonic) circulations. The geostrophic flow field (assumed steady during hurricane passage) was derived from measurements during the pre-Rita fight (15 September).
Fig. 12. Cross-track velocity response at the mooring site (point M, Fig. 1). (a) Inside the CCE2 that interacted with Katrina (Fig. 7a), and (b) inside the LC bulge affected by Rita (Fig. 7b). The vertical lines indicate the time of closest approach of the hurricane's eye to the mooring site (29 August for Katrina; 23 September for Rita).
Fig. 13. Water mass evolution at the mooring site (point M, Fig. 1). (a) Inside the CCE2 that interacted with Katrina (Fig. 7a), and (b) inside the LC bulge affected by Rita (Fig. 7b). IP stands for inertial period (25.5 hr), and STW for Subtropical Water. In (a) and (b) the red, green, and blue colors represent pre-, in-, and post-storm (near-inertial) variability, respectively. The black line represents a reference water mass from the WCE core.
Fig. 14. OML velocity response $u_s$ at the saturation level. (a) Katrina and (b) Rita. In each panel, $u_s$ is computed from airborne pre-Rita data (15 September) with Eq. (5) by applying a spatially homogeneous wind stress at the saturation level (see Table 3 and text for more detail). The color shade shows regions with mapping error less than 40% from the objective analysis technique.
Fig. 15. Along-track vertical velocity $w$ [Eq. (1)] induced by hurricane Rita over the geostrophic flow in the LC system (along the storm’s best-track in Fig. 11). The sea level pressure lines are segments from the pressure lines in Fig. 9. Solid (dashed) line in $w$ curves indicates underlying cyclonic (anticyclonic) geostrophic relative vorticity. The $T_{nf}$ curve stands for the time required for the 26°C isotherm to reach the sea surface in function of $w$ and $h_{26}$ (see text for details).
Fig. 16. Area-integrated energy, $\int \rho_o g h_{iso}^2 dA$ required to reduce SST respect to 30°C in a 1°×2° box by adiabatic lifting of isotherms over a vertical distance $h_{iso}$ (from airborne pre-Rita data, 15 September). For example, a 1°C (4°C) SST reduction means that the 29°C (26°C) isotherm was lifted adiabatically to the sea surface against the gravity acceleration. The rectangular area represents observed SST cooling ranges in the GOM during hurricanes Ivan (2004), and Katrina and Rita (2005). The upper limits of this rectangle are determined by the intersection of the average SST cooling observed during hurricane Ivan (~6.5°C) with the $PE$ curve of the CCE2. The intersection of the $PE$ curves with the negative feedback line indicates that about 2, 6, and $15 \times 10^{17}$ J are required to adiabatically reduce SST to 26°C and initiate the negative feedback over CCEs, Gulf Common Water (GCW), and LC/WCEs, respectively.
Fig. 17. Time required, $\Delta T_{nf} = PE h_w^{-1}$, for the initiation of the negative feedback for Katrina (a) and Rita (b), with $PE = \frac{1}{2} \int \rho g h_26^2 dA$, where $h_{26}$ is the $26^\circ$C isotherm depth, $dA$ the grid size of the objectively analyzed map, and $h_w$ from Eq. (6). In (a) and (b), $h_{26}$ is from pre-Rita (15 September) and post-Rita (26 September) airborne data, respectively. The contour intervals are: 2 hrs from 0 to 10, and 20 hrs from 20 to 340. See caption of Fig. 16 for the meaning of $PE$. 
**Table 1.** Summary of profilers deployed in the LC System before (15 September), during (22, 23 September), and after (26 September) the passage of hurricane Rita. The numbers in parentheses indicate profiler failures.

<table>
<thead>
<tr>
<th>Flight name</th>
<th>Date</th>
<th>AXBT</th>
<th>AXCTD</th>
<th>AXCP</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>pre-Rita</td>
<td>15 September</td>
<td>20(1)</td>
<td>30(12)</td>
<td>2(0)</td>
<td>52 (13)</td>
</tr>
<tr>
<td>Rita-22</td>
<td>22 September</td>
<td>18(2)</td>
<td>–</td>
<td>–</td>
<td>18 (2)</td>
</tr>
<tr>
<td>Rita-23</td>
<td>23 September</td>
<td>12(2)</td>
<td>–</td>
<td>–</td>
<td>12(2)</td>
</tr>
<tr>
<td>post-Rita</td>
<td>26 September</td>
<td>56(6)</td>
<td>–</td>
<td>–</td>
<td>56 (6)</td>
</tr>
</tbody>
</table>
Table 2. Summary of mooring probes that were active in the northern Gulf of Mexico during the passage of hurricanes Katrina and Rita. T is temperature, C conductivity, and u and v the horizontal components of the velocity vector.

<table>
<thead>
<tr>
<th>Probe type</th>
<th>Lat (°N)</th>
<th>Lon (°W)</th>
<th>Deployment time</th>
<th>Parameters</th>
<th>Depth range (m)</th>
<th>Vertical sampling interval (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>ADCP</td>
<td>27.998</td>
<td>87.839</td>
<td>''</td>
<td>u, v</td>
<td>~60-500</td>
<td>~8</td>
</tr>
<tr>
<td>ADCP</td>
<td>27.606</td>
<td>87.541</td>
<td>''</td>
<td>u, v</td>
<td>~30-470</td>
<td>~8</td>
</tr>
<tr>
<td>CTD</td>
<td>28.347</td>
<td>87.547</td>
<td>''</td>
<td>T, C</td>
<td>~75-400</td>
<td>~40,100</td>
</tr>
<tr>
<td>CTD</td>
<td>27.998</td>
<td>87.839</td>
<td>''</td>
<td>T, C</td>
<td>~75-400</td>
<td>~40,100</td>
</tr>
<tr>
<td>CTD</td>
<td>27.606</td>
<td>87.541</td>
<td>''</td>
<td>T, C</td>
<td>~60-380</td>
<td>~40,100</td>
</tr>
</tbody>
</table>
TABLE 3. Storm parameters for hurricanes Katrina and Rita. The values of the first baroclinic mode phase speed and Froude number are for the LC bulge (non parenthesis values) and the CCE2 (values in parenthesis). The parameters of hurricane Ivan (2004) are included for comparative purposes.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Ivan</th>
<th>Katrina</th>
<th>Rita</th>
</tr>
</thead>
<tbody>
<tr>
<td>Radius of maximum winds</td>
<td>$R_{max}$ [km]</td>
<td>32</td>
<td>42</td>
</tr>
<tr>
<td>Speed of the hurricane</td>
<td>$U_h$ [m s$^{-1}$]</td>
<td>5.5</td>
<td>6.3</td>
</tr>
<tr>
<td>First baroclinic mode phase speed</td>
<td>$c_1$ [m s$^{-1}$]</td>
<td>–</td>
<td>2.9 (2.1)</td>
</tr>
<tr>
<td>Froude number</td>
<td>$Fr = U_h (c_1)^{-1}$</td>
<td>–</td>
<td>2.2 (3)</td>
</tr>
</tbody>
</table>
TABLE 4. Internal near-inertial waves parameters from mooring ADCP inside the CCE2 (induced by hurricane Katrina), and the LC bulge (induced by hurricane Rita). $F_{iw}$ [Eq. (3)] is the internal near-inertial wave energy flux radiated into the thermocline, and represents the fraction of eddy kinetic energy not used to entrain colder thermocline water into the OML (mixed layer energy sink). $T_i$ is the inertial period (see text for description of the other parameters).

<table>
<thead>
<tr>
<th></th>
<th>$\Delta z$</th>
<th>$\Delta T_z$</th>
<th>$T_i$</th>
<th>$\bar{N}$</th>
<th>$V$</th>
<th>$\lambda_z$</th>
<th>$C_{g z}$</th>
<th>$F_{iw}$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>(m)</td>
<td>(hr)</td>
<td>(hr)</td>
<td>(c.p.h.)</td>
<td>(m s$^{-1}$)</td>
<td>(m)</td>
<td>(m day$^{-1}$)</td>
<td>(W m$^{-2}$)</td>
</tr>
<tr>
<td>CCE2</td>
<td>425</td>
<td>8.1</td>
<td>23.7</td>
<td>4.8</td>
<td>0.6</td>
<td>144</td>
<td>8.6</td>
<td>1.8×10$^{-2}$</td>
</tr>
<tr>
<td>LC</td>
<td>425</td>
<td>12.3</td>
<td>24.3</td>
<td>6.1</td>
<td>0.3</td>
<td>214</td>
<td>233</td>
<td>12.1×10$^{-2}$</td>
</tr>
</tbody>
</table>
Table 5. Hurricane-induced global internal near-inertial wave power for oceanic conditions similar to CCE2 (non parenthesis values) and LC bulge (values in parenthesis). $W_i$ is the wave power induced by an individual storm over an arbitrary storm's wake of area $A=dx\,dy$. The values of the vertical energy flux $F_{iw}$ are from Table 4. The global wave power $W_g$ is calculated by assuming a global average of 67 storms per year (Emanuel 2001), that each individual storm was present during 20 days, and that the storms had similar characteristics than hurricanes Katrina or Rita. In an eddy-free ocean the global wave power should be somewhere in between the CCE2 and LC values.

<table>
<thead>
<tr>
<th>Wake area</th>
<th>Wake description</th>
<th>$W_i = \int \int F_{iw} , dy , dx$</th>
<th>$W_g = W_i \times 67 \times (20/366)$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>[$\times 10^{10}$ W]</td>
<td>[$\times 10^{10}$ W]</td>
</tr>
<tr>
<td>200 km×200 km</td>
<td>area of a CCE/WCE</td>
<td>0.07 (0.5)</td>
<td>0.3 (1.8)</td>
</tr>
<tr>
<td>400 km×2000 km</td>
<td>average storm wake</td>
<td>1.4 (9.7)</td>
<td>5.1 (35.5)</td>
</tr>
<tr>
<td>1000 km×2000 km</td>
<td>Emanuel (2001)</td>
<td>3.6 (24.2)</td>
<td>13.2 (88.6)</td>
</tr>
</tbody>
</table>