Part I: Observations

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ABSTRACT

The ocean mixed layer response to a tropical cyclone within and immediately adjacent to the Gulf of Mexico Loop Current is examined. In the first of a two-part study, a comprehensive set of temperature, salinity, and current profiles acquired from aircraft-deployed expendable probes is utilized to analyze the three-dimensional oceanic energy evolution in response to Hurricane Lili’s (2002) passage. Mixed layer temperature analyses show that the Loop Current cooled $\Delta T > 1^\circ C$ in response to the storm, in contrast to typically observed larger decreases of $3^\circ - 5^\circ C$. Correspondingly, vertical current shear associated with mixed layer currents, which is responsible for entrainment mixing of cooler water, was found to be up to 50% weaker, on average, than observed in previous studies within the directly forced region. The Loop Current, which separates the warmer, lighter Caribbean Subtropical Water from the cooler, heavier Gulf Common Water, was found to decrease in intensity by $|\Delta u| = 2^0.18 \pm 0.25 \, m \, s^{-1}$ over an approximately 10-day period within the mixed layer. Contrary to previous ocean response studies, which have assumed approximately horizontally homogeneous ocean structure prior to storm passage, a kinetic energy loss of $5.8 \pm 6.4 \, kJ \, m^{-2}$, or approximately $\frac{1}{2}$ wind stress-scaled energy unit, was observed. By examining near-surface currents derived from satellite altimetry data, the Loop Current is found to vary similarly in magnitude over such time scales, suggesting storm-generated energy is rapidly removed by the preexisting Loop Current. In a future study, the simulated mixed layer evolution to a Hurricane Lili–like storm within an idealized preexisting baroclinic current is analyzed to help understand the complex air–sea interaction and resulting energetic response.

1. Introduction

Hurricanes, and more generally tropical cyclones (TCs), are among the most intense organized vortical systems observed in the atmosphere. These cyclones derive their energy primarily from the release of latent heat upon condensation of water vapor (Ooyama 1969). Thus, it is necessary that a large moisture source be present, such as the ocean, and that the sea surface temperature (SST) is sufficiently warm to maintain a moist enthalpy flux from the sea to the atmospheric boundary layer, as was first recognized by Palmen (1948). Numerous other studies have examined the relationship between the intensity of TCs and the SST. Malkus and Riehl (1960) derived a relationship between the decrease in central pressure and the increase in equivalent potential temperature in the eyewall region (due to imported enthalpy from the ocean). Further studies by Emanuel (1986) and Betts and Simpson (1987) confirmed this relationship to be approximately constant. More recent research has studied not only the influence of SST on TC intensity but also the relationship between intensity and the upper-ocean thermal energy (Shay et al. 2000).

Because the underlying ocean significantly modulates TC intensity, much attention has been drawn toward gaining a better understanding of the physical interaction between the atmosphere and ocean during these events. Unfortunately, because of limited observational data at the air–sea interface in high-wind conditions, the
understanding of momentum and energy transfer processes has not progressed enough to significantly increase forecast accuracy. The recent Office of Naval Research sponsored Coupled Boundary Layer Air–Sea Transfer (CBLAST) field experiment in 2003–04 (Black et al. 2007) was designed to address questions regarding poorly understood air–sea exchange processes in TCs such as the maturity of the sea state (Donelan et al. 1993), sea spray (Fairall et al. 1994), and boundary layer roll vortices (Foster 2005) and their possible impacts on TC intensity. In sum, the various conclusions generally now agree that the bulk enthalpy and momentum exchange coefficients at high wind speeds are not as large as previously assumed (Drennan et al. 2007; French et al. 2007).

Further complicating TC air–sea interactions is the sometimes large variability in the upper ocean typical of large-scale ocean current systems (e.g., Gulf Stream, Kuroshio) that exist near continental western boundaries where TCs often make landfall. A particularly interesting region to examine the effects of variable ocean structure on TC intensity is the Gulf of Mexico (GOM). Studies of Hurricanes Gilbert (Shay et al. 1992), Opal (Shay et al. 2000), Ivan (Walker et al. 2005), Katrina, and Rita (Shay 2009; Jaimes and Shay 2009, 2010) have demonstrated the potential influence GOM upper-ocean variability can have on TC intensity. As revealed by microwave satellite imagery, SST cooling (indicated by the arrow in Fig. 1) was associated with Hurricane Lili’s (2002) passage through the GOM. However, this cooler water is clearly confined to an area outside of the GOM Loop Current (LC), which is located in the extreme southeast portion of the GOM. While Lili traversed the GOM, it underwent a period of rapid intensification from Saffir–Simpson category 1 to 4, immediately followed by a weakening to category-1 intensity, over a 2.5-day period (Pasch et al. 2004). Particularly relevant to understanding air–sea energy exchange over the LC, the resulting modulation of SST cooling, and ultimately feedback to TC intensity is the large horizontal transport and its impact on shear-induced mixing. Adding to uncertainties in forecasting hurricane intensity in this region is the often-complex atmospheric environment, which also exerts an influence on intensity (Bosart et al. 2000).

During 2002, life cycles of Hurricanes Isidore and Lili in the northwest Caribbean Sea and GOM were extensively observed (Shay and Uhlhorn 2008), which indicated deep, warm ocean structures cooled relatively little and provided a relative positive thermal feedback to these storms. This is in contrast to more typical negative feedback on hurricane intensity when shear-induced mixing at the ocean mixed layer (OML) base cools and deepens this layer and causes a cold ocean wake where air–sea fluxes decrease (Chang and Anthes 1978; Price 1981; Shay et al. 1992; Schade and Emanuel 1999; Bender and Ginis 2000). These observations demand a detailed understanding of the physical mechanisms responsible for preventing the expected cooling. In this first of a two-part study, the analysis of these observations in Lili is expanded beyond primarily the thermal response previously reported by the authors. Particular attention is given to analyzed fields of mixed layer currents and resulting mechanical energy. In section 2, mass and current field objective analyses are constructed from observations obtained in the GOM Loop Current region around the time of Lili’s passage. From these data, mixed layer depth and geostrophic current fields are estimated. In section 3, the evolution of mixed layer mechanical energy is computed from the analyses, and section 4 examines the variability of the Loop Current system in the southeast GOM to help understand the observational results. A summary of results and conclusions are presented in section 5. In a future study, details of the energy interactions are examined with the aid of an idealized numerical model.

2. Observed fields

As reported in Shay and Uhlhorn (2008), a large set of temperature, salinity, and current vertical profiles was obtained prior to, during, and after passage of Hurricane Lili (2002) across the southeast GOM during a joint
National Oceanic and Atmospheric Administration–National Science Foundation (NOAA–NSF) aircraft-based field research experiment. These data underwent an extensive processing and quality-control effort to analyze details about the upper-ocean evolution in response to a Saffir–Simpson category-3 storm. Figures 2–6 show aircraft expendable bathythermograph (AXBT) probe, aircraft expendable conductivity–temperature–depth (AXCTD) probe, and aircraft expendable current probe (AXCP) profile locations over the course of the experiment. The fairly dense horizontal coverage of vertical profiles provided an opportunity to compute optimally interpolated objective analyses of three-dimensional (3D) synoptic fields for relevant quantities to examine the mass, momentum, and energy evolution over a 10-day period. Details regarding experimental goals, methodology, and data processing are provided in the above referenced article.

Fig. 2. Locations of the ocean probes deployed prior to the passage of Hurricane Lili used to estimate the initial temperature conditions. AXBTs (circles), AXCTDs (×‘s), and AXCPs (pluses). Lili’s observed “best track” is indicated by the solid line, and marked at 6-h intervals. The storm travels from southeast to northwest. The solid box indicates the analysis region for this study, and the dotted contours identify the 200- and 1000-m isobaths.
a. Mass distribution

1) TEMPERATURE AND SALINITY FIELDS

Temperature and salinity structures are known to be unique for the Caribbean Subtropical Water (STW) and Gulf Common Water (GCW) (e.g., Wüst 1964; Nowlin and Hubertz 1972; Behringer et al. 1977). Prestorm AXCTD profiles at two locations within the experimental domain separated by ~300 km reveal a distinct difference in measured temperature and salinity and the derived mass density (Fig. 7). Relevant to the upper-ocean energetics is the LC driven by this density anomaly, which is confined primarily to the upper 200 m.

Based on the observed upper-ocean profiles of temperature and salinity, objective analyses for these quantities from the prestorm observations (18–23 September) are shown for the surface and 100-, 200-, and 300-m depths in Fig. 8. Fields are computed at yearday 265.0, corresponding to 0000 UTC 22 September; therefore, observations obtained later in the prestorm experimental period (and closer in time to storm passage) are weighted more heavily. Analyses clearly identify the location of the northern extension of the STW into the southern GOM. This generally warmer and more saline core is separated from the GCW by the LC system. Although the surface temperature and salinity show relatively little horizontal variability, the subsurface fields indicate the marked contrast between water masses. The strong temperature gradient becomes apparent at ~100-m depth, whereas the largest salinity gradient is found at ~200 m. These gradients gradually become weaker at greater depths.

An SST analysis is shown in Fig. 9a for profiles obtained during the storm passage on 2 October. Based on nearly uniform SSTs found in the prestorm analysis, evidence of ~2°C surface cooling (Fig. 9b) is found in the extreme northwest portion of the observation domain at the boundary between the LC and GCW and along the storm track, which is within expected limits of inner-core SST cooling in TCs (Cione and Uhlhorn 2003). As opposed to the typical situation where larger cooling is observed to the rear of a storm, the cooling occurs forward of the storm in the GCW while temperatures remain close to prestorm values in the LC and STW.

Several temperature profiles obtained poststorm, mostly from AXCTD and AXBT, show evidence of near-surface warming and restratification because of strong solar insolation and weak surface winds soon after storm passage, as observed by Hazelworth (1968). To obtain more realistic poststorm surface temperatures that are representative of the actual thermal response to the TC, a subjective correction (see example in Fig. 10) is performed. The mean excess SST for all profiles is found to be 0.31 ± 0.19°C and occurs over an average ~10-m layer depth. Assuming an average ~70 W m⁻² (Price and Morzel 2006) radiative energy flux absorbed over a 10-m layer for a 48-h duration after storm passage, a 0.28°C temperature increase would result, suggesting the observed warming is reasonable.
Poststorm temperature and salinity analyses shown in Fig. 11 are generated from profiles acquired on 4 October, about 60 h (~2 inertial periods) after the passage of Lili. Poststorm adjusted surface cooling of ~2.5°–3°C was found in the GCW to the right of the storm track (Shay and Uhlhorn 2008), slightly less than the 3°–5°C cooling typically observed in TC cold wakes (Black 1983). Otherwise, over most of the domain there is relatively little evidence of significant surface cooling. The subsurface horizontal temperature gradient appears to be generally weaker than observed in the prestorm fields, especially at 200- and 300-m depths (cf. Figs. 11c,d with Figs. 8c,d). In addition, both the temperature and salinity fields have shifted to the east of their initial positions, because the axis of the STW extension is now oriented north–south. It does not appear likely that this large pattern change was a direct result of the storm.

2) DENSITY FIELDS

The 3D mass density distribution \( \rho(x, y, z) \) is responsible for driving the LC flow under geostrophic balance constraints. The prestorm and poststorm density field analyses shown in Fig. 12 are computed from the objectively analyzed temperature \( T \) and salinity \( S \) using the Fofonoff and Millard (1983) seawater equation of state \( \rho = \rho(S, T, p) \). The density distribution is expectedly (anti-) correlated with the temperature fields, with warm/light water in the southeast part of the domain and cold/heavy water to the northwest. However, the density gradient becomes weaker at greater depths (Figs. 12d,h), a result of the higher salinity in the STW, which compensates to homogenize the density field at depths below 200 m. Poststorm density fields show an increase of ~0.6 kg m\(^{-3}\) associated with the observed
2.5°C cooling found in the northwest and right of storm track, and the poststorm ridge axis shifts from its pre-storm position, as was found in the \( T \) and \( S \) fields.

### b. Currents

#### 1) Observed Currents

Critical to fully understanding LC response to TC forcing, it was an experimental goal to observe the pre-existing mesoscale current field using AXCPs, which might be responsible for modulating the upper-ocean energy budget. As described in Shay and Uhlhorn (2008), surface wave orbital velocity-induced currents are initially removed from the AXCP-measured current profiles. This procedure (Sanford et al. 1987) also estimates the mean OML current velocity and the current shear directly below the OML for each profile, which are of primary interest for analyzing the evolving current in response to the storm. Based on these three-layer statistical model fits to current profiles, the mean OML current vector \( \mathbf{V} \) is estimated for each AXCP. The pre-storm, near-surface analyzed velocity field (Fig. 13) clearly reveals the LC location relative to Lili’s track and indicates peak current speeds of \( \approx 1.1 \text{ m s}^{-1} \), although a few individual profiles show peak speeds as high as \( \approx 1.6 \text{ m s}^{-1} \).

Based on observed pre-storm currents, the LC is defined here to be confined \( \mathbf{e} \cdot \mathbf{V}_{\text{max}} \) analysis isotachs in Fig. 13. This definition limits the LC to the geographic location in which pre-storm OML current speed exceeds \( 0.38 \text{ m s}^{-1} \). Observations and analysis grid points lying to the warm side of the LC fall into the Caribbean STW; accordingly, data on the cold side are assumed to lie within the GCW.

#### 2) Geostrophic Currents

The strongest observed currents are found to be well correlated with large horizontal density gradient indicating the LC is primarily driven by the mass-field anomaly. The three-dimensional geostrophic velocity

![Fig. 7.](image-url)
\( \mathbf{V}_g \) is computed by vertically integrating the Boussinesq-approximated thermal wind equations,

\[
\int_{z_0}^{z} \left\{ \frac{\partial \mathbf{V}_g}{\partial z} = -\frac{g}{f_0 \rho_0(z)} \left[ \mathbf{k} \times \nabla \rho(x, y, z) \right] \right\} dz',
\]

where \( g \) is the gravity acceleration and \( f_0 \) is the Coriolis parameter. The integration is carried out subject to the boundary condition \( \mathbf{V}_g(z_0) = 0 \), where \( z_0 \) is the smaller of 750 m (Jacob et al. 2000) and the actual depth. The density fields derived from the \( T \) and \( S \) analyses are used to calculate the three-dimensional \( \mathbf{V}_g \) field for prestorm (19–23 September) conditions (Fig. 14).

A vertical cross section in the along-track (cross stream) direction (Fig. 15) clearly identifies LC structure. As determined by horizontal density gradients, the
LC is mostly confined to the upper 300 m. Maximum surface geostrophic currents speeds are found in the LC to be \(\approx 0.75 \text{ m s}^{-1}\), and speeds \(\geq 0.4 \text{ m s}^{-1}\) (~1 e folding from the peak observed value) are confined to a cross-stream width of \(~100 \text{ km}\). Peak geostrophic currents are \(\approx 0.3 \text{ m s}^{-1}\) weaker than the observed maximum in situ currents, for which there are multiple possible reasons. The surface \(V_g\) is computed relative to 750-m depth; however, observed currents at large depths are generally found to flow counter to the surface direction at speeds of \(~0.2 \text{ m s}^{-1}\) (Maul 1977; Ochoa et al. 2001), which would explain a surface \(V_g\) underestimate and suggests the 750-m motionless-level assumption is incorrect. Another potential reason for a low bias exists in the balance assumption itself. The LC’s large anticyclonic curvature (i.e., small radius of curvature) suggests gradient balance may be more appropriate for describing flow. In fact, for an anticyclone with radius of curvature \(R = -100 \text{ km}\) at 24\(^\circ\) latitude, \(|V_g|\) is approximately 20% less than the gradient current (Holton 1992, chapter 3).

3) AGEOSTROPHIC CURRENTS

The LC response is quantified in terms of observed currents \(V\), geostrophic currents \(V_g\), and ageostrophic current \(V_a = V - V_g\). The estimated OML mean \(V_g\) at the AXCP location is estimated by bilinear interpolation. The ageostrophic current in the OML serves as a proxy for the forced, near-inertial motion, which scales directly with the storm’s intensity when the ocean is initially in a state of rest. The analyzed prestorm and poststorm OML mean observed, geostrophic, and ageostrophic currents and their respective changes are shown in Fig. 16. Observed currents (Figs. 16a–c) generally show a slight decrease in intensity within the LC and little change in the STW. As a small increase is found in the GCW. Geostrophic currents (Figs. 16d–f) show weakening of \(~0.4 \text{ m s}^{-1}\) in the LC, especially at the inflow from the south. Overall, the ageostrophic current response is somewhat mixed and confused with possibly large errors, because of subtracting two quantities of similar magnitude, especially in the LC. Clearly, a classic near-inertial current wake, as described by linear theory (Geisler 1970), is not evident in these observations.

![Fig. 10. Example temperature profile from AXCTD, showing near-surface warming confined to a shallow layer. The shaded area represents an estimate of poststorm warming: in this example, +0.4°C.](image)

![Fig. 11. As in Fig. 8, but for poststorm profiles obtained on 4 Oct.](image)
Based on measurements of OML $V$ from AXCP current profiles (Shay and Uhlhorn 2008), statistics for magnitudes are calculated in each of the three physical regions for prestorm and poststorm profiles, and additionally changes are also computed (Table 1). Because of the fairly small number of profiles individually for each region, all profiles are included in the statistics regardless of radial distance from the storm track.

All values in the Caribbean STW show only small average change between prestorm and poststorm conditions. An average increase in $|V_a|$ of $0.10 \pm 0.16 \text{ m s}^{-1}$ may be compared with the wind-driven current scale value of $0.20 \text{ m s}^{-1}$, which assumes the ocean is initially at rest. The GCW shows a somewhat unexpectedly weak response in $|V_a|$ near the storm track, with an average increase of $<0.1 \text{ m s}^{-1}$. However, this is consistent with the fairly weak poststorm shear as compared with previous observational results (Shay and Uhlhorn 2008). Because Lili was a rather small, fast-moving storm, it is not surprising that a strong response was not found. Of particular interest is the overall observed decrease in the LC speed from prestorm to poststorm, which is found for both the observed and geostrophic currents. This result contradicts most previous studies of upper-ocean current response to TCs, which have often observed strong near-inertial currents after storm passage.

OML current statistics are computed for a cross-storm-track swath from $-2$ to $+4R_{\text{max}}$ from the analyzed fields (Table 2). Because the objective analyses tend to filter high-frequency variability, results are more clear than for the individual profiles. Analysis field statistics corroborate the results from individual profiles, highlighting the general decrease in current magnitude within the LC. Contrasting the weak current response found for Lili, Shay et al. (1998) found...
wind-forced, near-inertial currents within the GCW in Hurricane Gilbert > 1 m s⁻¹.

c. OML depth

Numerous criteria to estimate the OML depth \( h \) from profiles have been established in the literature, suggesting a general lack of agreement on a single definition. Thresholds in both temperature decrease and density increase from surface values have been applied. Temperature-based criteria vary from \( \Delta T = 0.1°C \) (Martin 1985) to \( \Delta T = 1.0°C \) (Hastenrath and Merle 1987), where \( \Delta T = SST - T(h) \). A common density definition in terms of \( \sigma_r = (\rho - 10^{3}) \text{kg m}^{-3} \) is \( \Delta \sigma_r = -0.125 \text{kg m}^{-3} \) (Levitus 1982). Still, other authors have applied vertical temperature gradient thresholds (e.g., Lukas and Lindstrom 1991; Shay et al. 1992).

Within this definitive range, \( h \) can vary several tens of meters (Fig. 17). In their analysis of GOM temperature profiles in Hurricane Gilbert, Shay et al. (1992) used a gradient criterion of \( dT/dz = 0.1°C \text{m}^{-1} \). Based on Lili prestorm temperature analyses, mixed layer depths could not be computed in the STW using this method, because of the very weak stratification, where lapse rates never exceeded 0.08°C m⁻¹ anywhere in the profile. Depths in the GCW, however, were found to be 30–40 m using this threshold, in agreement with OML depth estimates in Hurricane Gilbert (Shay et al. 1992). Lukas and Lindstrom (1991) used a \( dT/dz = 0.05°C \text{m}^{-1} \) definition in their analysis of western equatorial Pacific depths, which when applied here give OML depths ranging from 150 m in the STW to 40 m in the GCW. This method produces average \( \Delta T \) values as defined above, of 1.8°C ± 0.6°C, which appear to be far too large.

Based on the observed current profiles, a subjective estimate of the well-mixed layer depth was made for AXCP profiles (\( Z_1 \) values tabulated in Shay and Uhlhorn 2008). For each of the poststorm current profiles, where the top of the shear layer was most clearly identified, \( \Delta T = SST - T(Z_1) \) is calculated. A mean \( \Delta T = 0.7°C ± 0.4°C \) is found for 22 profiles. This value compares reasonably well with results reported in Kara et al. (2000), which suggest that a \( \Delta T = 0.8°C \) criterion best represents the depth of turbulence penetration into the thermocline or the bottom of the entrainment zone below the OML. This \( \Delta T = 0.8°C \) threshold is applied to the objectively analyzed temperature fields to estimate the OML depth.

Prestorm and poststorm OML depths (Fig. 18) closely resemble the subsurface distribution of temperature and mass in the LC vicinity, ranging from >90 m in the LC to <40 m in the GCW. Based on temperature analysis errors, typical OML depth errors are approximately 5–7 m for both prestorm and poststorm fields over most of the experimental domain, except for 15–20-m errors in the extreme S corner in the poststorm analysis due to locally poor data coverage. Because the rate of change of OML temperature is inversely proportional to the OML depth (e.g., Pollard et al. 1973), greater surface cooling is expectedly found in the northwest portion of the observational domain, as depicted previously in Fig. 9.
3. Energy response

a. Air–sea parameters and scaling

Parameters that describe the expected energy response are developed, as proposed by Price (1981). Shay and Uhlhorn (2008) estimated storm-forcing scales (Table 3) based on observations in Hurricane Lili, and important scales from the prestorm upper-ocean fields are restated in Table 4. From these parameters, a set of dimensional response scales is estimated to describe OML energetics (Table 5). Because of large variation across the experimental domain, we expect a highly complex response, and parameters are developed separately for each of the three dynamic regions. Considering the significant pre-existing current at least as large the expected wind-driven current, isolating the near-inertial response from the highly variable geostrophic background using previous methods (e.g., Shay et al. 1998; Jacob et al. 2000) becomes less practical, as was found in the observation analyses of ageostrophic currents in the previous section.

Fig. 16. Analyzed OML currents (a),(d),(g) prestorm; (b),(e),(h) poststorm; and (c),(f),(i) changes (a)–(c) \( V \), (d)–(f) \( V_g \), and (g)–(i) \( V_a \). In (c),(f),(i), red (blue) contours are positive (negative) change. Contour intervals are 0.2 m s\(^{-1}\).
b. OML mechanical energy

The rotating Cartesian, hydrostatic, and Boussinesq equations of motion are given by

\[
\rho_0 \frac{d\mathbf{V}}{dt} = -\rho_0 f_0 \mathbf{k} \times \mathbf{V} - \mathbf{V} \cdot \nabla p' - \rho_0 g' \mathbf{k} + \frac{\partial \tau}{\partial z}. \tag{2}
\]

An approximate form of the mechanical energy equation for the OML is derived by scalar multiplying Eq. (2) by \(\mathbf{V}\) and vertically integrating over the OML depth,

\[
\frac{d(K + P)}{dt} = -\mathbf{V} \cdot \nabla_h p' - p'w + \mathbf{V} \cdot \tau, \tag{3}
\]

where \(K\) and \(P\) are the kinetic and potential energy, respectively, with dimensions of energy per unit area of OML given by

<table>
<thead>
<tr>
<th>Locations</th>
<th>Variable</th>
<th>Pre</th>
<th>Post</th>
<th>(\Delta)</th>
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<tbody>
<tr>
<td>STW</td>
<td>(</td>
<td>V</td>
<td>)</td>
<td>0.35 ± 0.23</td>
</tr>
<tr>
<td></td>
<td>(</td>
<td>V_g</td>
<td>)</td>
<td>0.43 ± 0.20</td>
</tr>
<tr>
<td></td>
<td>(</td>
<td>V_a</td>
<td>)</td>
<td>0.27 ± 0.14</td>
</tr>
<tr>
<td>LC</td>
<td>(</td>
<td>V</td>
<td>)</td>
<td>0.90 ± 0.40</td>
</tr>
<tr>
<td></td>
<td>(</td>
<td>V_g</td>
<td>)</td>
<td>0.55 ± 0.23</td>
</tr>
<tr>
<td></td>
<td>(</td>
<td>V_a</td>
<td>)</td>
<td>0.63 ± 0.43</td>
</tr>
<tr>
<td>GCW</td>
<td>(</td>
<td>V</td>
<td>)</td>
<td>0.34 ± 0.31</td>
</tr>
<tr>
<td></td>
<td>(</td>
<td>V_g</td>
<td>)</td>
<td>0.32 ± 0.18</td>
</tr>
<tr>
<td></td>
<td>(</td>
<td>V_a</td>
<td>)</td>
<td>0.26 ± 0.15</td>
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</tbody>
</table>

FIG. 17. Example (left) density, (middle) temperature, and (right) salinity profiles illustrating OML depth estimate variability under a number of cited criteria.
The first and second terms on the right-hand side of Eq. (3) are the work done by current against the pressure gradient, and the third term is net surface energy flux to the OML (wind stress minus shear stress). Initially, nearly all wind energy input will go toward increasing the mechanical energy, and it will take several inertial periods before significant energy is transferred to the thermocline by the wave-energy flux (Gill 1984). Furthermore, with a preestablished OML, most of the energy is expected to go to increasing kinetic energy of the current (Pollard et al. 1973).

In an approximate one-dimensional system, internal wave fluxes cannot be supported and correspond to purely inertial motion. Although the Loop Current or any highly variable oceanic regime can hardly be accurately modeled by such a system, comparisons of observations here can be made with previous studies, which have often assumed approximately horizontally homogeneous initial conditions, primarily aimed at understanding mixing and OML deepening from wind forcing. The remaining source/sink flux terms are examined in simulations in a future study, because they are particularly difficult to directly compute from the observations obtained herein.

From observed current and mass–density analyses, OML mechanical energy is computed from the prestorm and poststorm fields, as shown in Fig. 19. LC prestorm kinetic energy (Fig. 19a) is approximately 20–25 kJ m\(^{-2}\), decreasing to less than 5 kJ m\(^{-2}\) outside. In accord with the observed decrease in current magnitude in the LC form prestorm to poststorm conditions, \(K\) curiously also weakens to approximately 10–15 kJ m\(^{-2}\) (Fig. 19b). A small increase in \(K\) is found in the GCW, as the expected response to the storm. The potential energy fields generally reflect the analyzed OML depth distribution. Based on scaling arguments, a slight increase in \(P\) would be expected; however, observational analyses suggest much more complex modification, possibly due to mesoscale oceanic variability independent of the storm.

The energy distribution in the LC and surrounding STW and GCW is further quantified as a function of normalized cross-storm-track distance \((x/R_{\text{max}})\), as shown in Figs. 20 and 21. Error statistics are computed by a Monte Carlo method based on analyzed field errors, as

### Table 3. Storm scaling parameters based on observations in Hurricane Lili.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
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</thead>
<tbody>
<tr>
<td>Radius of max stress (R_{\text{max}})</td>
<td>20 km</td>
</tr>
<tr>
<td>Max stress (t_{\text{max}})</td>
<td>7.0 Pa</td>
</tr>
<tr>
<td>Storm speed (V_s)</td>
<td>7.0 m s(^{-1})</td>
</tr>
<tr>
<td>Inertial period (2\pi/f_0)</td>
<td>29.4 h at 24°N</td>
</tr>
<tr>
<td>Inertial wavelength (V_s/f_0)</td>
<td>740 km</td>
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### Table 4. Upper-ocean scaling parameters based on prestorm observations.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>STW</th>
<th>LC</th>
<th>GCW</th>
</tr>
</thead>
<tbody>
<tr>
<td>OML reference density (\rho_0) (kg m(^{-3}))</td>
<td>1023</td>
<td>1023</td>
<td>1023</td>
</tr>
<tr>
<td>OML depth (h_0) (m)</td>
<td>90</td>
<td>60</td>
<td>40</td>
</tr>
<tr>
<td>Geostrophic current (</td>
<td>V_g</td>
<td>) (m s(^{-1}))</td>
<td>0.2</td>
</tr>
<tr>
<td>Thermocline stratification (g') (×10(^2) m s(^{-2}))</td>
<td>1.0</td>
<td>1.5</td>
<td>2.5</td>
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</tbody>
</table>
described in Shay and Uhlhorn (2008). In Fig. 20c, the LC shows an overall decrease in $K$ of 5–10 kJ m$^{-2}$, which appears fairly significant considering the error bars on the mean change. The GCW shows a slight increase in $K$ but no clear indication of strong response close to the peak forcing location at $R_{\text{max}}$. Large changes in $P$, especially in the STW, are found (Fig. 21c), although because of relatively large uncertainty in OML depth, $P$ errors are expectedly highly magnified.

Summary statistics for prestorm, poststorm, and changes in observed $K$ and $P$ are presented in Tables 6 and 7, respectively. Between $2$ and $14R_{\text{max}}$, the OML mean $K$ is found to decrease an average $5.8 \pm 6.4$ kJ m$^{-2}$ in response to the storm, corresponding to the observed

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Scale</th>
<th>STW</th>
<th>LC</th>
<th>GCW</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wind-driven current</td>
<td>$\delta U = \tau_{\text{max}}R_{\text{max}}/\rho_0h_0V_s$ (m s$^{-1}$)</td>
<td>0.22</td>
<td>0.33</td>
<td>0.49</td>
</tr>
<tr>
<td>OML deepening</td>
<td>$\delta h = (\tau_{\text{max}}/\rho_0)^{1/2}\delta U/g'$ (m)</td>
<td>1.8</td>
<td>1.8</td>
<td>1.6</td>
</tr>
<tr>
<td>Kinetic energy increase</td>
<td>$\rho_0V_0\delta U^2$ (kJ m$^{-2}$)</td>
<td>4.3</td>
<td>6.5</td>
<td>9.8</td>
</tr>
<tr>
<td>Potential energy increase</td>
<td>$\rho_0g\delta h^2$ (kJ m$^{-2}$)</td>
<td>0.03</td>
<td>0.05</td>
<td>0.07</td>
</tr>
</tbody>
</table>

Table 5. Dependent variable scales of the upper-ocean response.

Fig. 19. (a),(c) Prestorm and (b),(d) poststorm OML (a),(b) kinetic energy and (c),(d) potential energy fields. Values are in units of kJ m$^{-2}$.
weakened OML current. Furthermore, large changes in \( P \) are found, which do not appear to be in direct response to the storm forcing. By adding the results for changes in \( K \) and \( P \), a general increase in OML mechanical energy is found; however, kinetic energy response should far exceed potential energy for wind-induced upper-ocean mixing (at least 3 times greater; Pollard et al. 1973).

Referring to the energy scales developed based on the known forcing, the overall loss of \( K \) near the storm track is around 0.9 input units (1 unit = 6.5 kJ m\(^{-2}\)). However, in terms of total mechanical energy \( (K + P) \), an increase of around 0.6 units is found in the LC, suggesting that around -0.4 units are gained through advective transport and wave flux. Although the error estimates are rather large and therefore energy changes are not statistically significant, results point to the need for capturing these other processes to properly estimate the energy budget of the OML response in a region of large horizontal variability. Furthermore, a large-scale current system such as the LC is also highly variable in space in time, independent of strong surface forcing events. These energy fluctuations in the LC and surrounding ocean at short (weekly) time scales may be far greater than changes induced by TCs, suggesting accurate and frequent reinitialization of upper-ocean structure in similar regimes is necessary for coupled numerical models.

4. Background Loop Current variability

In light of the somewhat unexpected result that the OML LC energy is observed to decrease after storm passage, it appears necessary to perform an analysis of LC variability independent of TC forcing. With little existing sustained and continuous in situ data about LC variability (Sturges 1992) and most certainly not at the time scales of interest here (days to weeks), satellite radar altimetry data become the most viable option to address this uncertainty. An obvious limitation of altimetric data is the absence of vertical structure information; the sea surface elevation and associated balanced current merely respond to the vertically integrated mass distribution. Particular attention here is given to LC energetic variability through the Hurricane Lili experimental domain and not to the overall structure as it periodically penetrates, retreats, and sheds eddies throughout the GOM (e.g., Sturges and Leben 2000).

a. Altimetric analysis

The sea surface elevation, as measured by a radar altimeter, responds to the mass–density distribution within the ocean and below the ocean floor. After subtracting the solid-earth-induced topography (the geoid),

| Table 6. Summary statistics (means and standard deviations) for \( K \) (kJ m\(^{-2}\)) between -2 and +4R\(_{max}\). |
|-----------------|-----------------|-----------------|-----------------|
| Location        | Pre             | Post            | \( \Delta \)    |
| STW             | 7.2 ± 3.6       | 5.8 ± 2.3       | -1.5 ± 3.5      |
| LC              | 18.7 ± 8.4      | 12.9 ± 5.3      | -5.8 ± 6.4      |
| GCW             | 3.2 ± 3.5       | 4.6 ± 2.5       | +1.4 ± 2.2      |

| Table 7. Summary statistics (means and standard deviations) for \( P \) (kJ m\(^{-2}\)) between -2 and +4R\(_{max}\). |
|-----------------|-----------------|-----------------|-----------------|
| Location        | Pre             | Post            | \( \Delta \)    |
| STW             | 23.0 ± 8.6      | 35.6 ± 17.0     | +12.5 ± 24.6    |
| LC              | 12.4 ± 4.8      | 21.9 ± 9.8      | +9.5 ± 11.0     |
| GCW             | 8.9 ± 2.6       | 15.6 ± 6.9      | +6.7 ± 6.9      |

Fig. 20. (a) Prestorm, (b) poststorm, and (c) change in OML kinetic energy (kJ m\(^{-2}\)) for three analyzed regions as a function of cross-storm-track distance.

Fig. 21. (a) Prestorm, (b) poststorm, and (c) change in OML potential energy (kJ m\(^{-2}\)) for three analyzed regions as a function of cross-storm-track distance.
the sea surface dynamic topography, which is related to surface currents via mass–momentum balance, can be measured. The 0.5°-resolution Rio-05 mean dynamic topography (MDT) product (Rio and Hernandez 2004) serves as a reference topography and is developed by averaging over 7 yr of multisensor altimeter data (Fig. 22a). The mean LC position is clearly identified in this field, and, because the product averages multiple modes of extension into the GOM, the LC appears to be stretched in the northwest–southeast direction relative to its instantaneous and typically much narrower channel.

To estimate the total instantaneous (10-day mean) surface elevation field, sea height anomaly (SHA) fields are added to the MDT. Altimeter SHA ground tracks are obtained from Naval Research Laboratory (NRL)/ Naval Oceanographic Office (NAVOCEANO) for all available satellite altimeters operating during 1997–2006 [Ocean Topography Experiment (TOPEX)/Poseidon, European Remote Sensing Satellite-1 (ERS-1), ERS-2, Jason, and Geosat Follow-On (GFO)]. Because of relatively infrequent revisit periods, available SHA tracks are placed into 10-day bins, and the global mean SHA over the 10-day period is removed separately for each satellite to eliminate possible cross-platform bias. Finally, optimal interpolation is used to produce a 0.25° grid of SHAs (Fig. 22b) and is added to the MDT (interpolated to 0.25°) to produce dynamic height $h(x,y)$ fields every 10 days. The sea height in Fig. 22c is a 10-day average field centered around yearday 265 in 2002, or near the time of Lili’s passage.

b. Derived currents and spectra

Currents driven by the mass distribution as inferred from the $\eta(x,y)$ field are estimated from geostrophic balance, $V_g(x,y) = -\frac{g}{f_0} V_H \eta(x,y)$. (6)

To analyze the time series of currents within the LC flowing through the Lili domain, the LC is defined here as containing current magnitudes $\geq (1/e)|V_{\text{max}}|$, as for the in situ data, where $|V_{\text{max}}|$ is the peak current speed found in the domain and must be at least 0.5 m s$^{-1}$ to consider the LC as flowing through the experimental domain. Because this domain lies nearly due north of the Yucatan Straits, some portion of the LC is always found to intersect this area of interest. The LC location, as depicted in Fig. 23, appears well correlated with that determined by the in situ analysis (Fig. 13). In addition, currents estimated from geostrophic balance indicate marked consistency with measured near-surface currents, with peak values of $\sim 0.9$ m s$^{-1}$ and similar LC channel width, indicating the height gradient is fairly well resolved.

Mean current speed and variability (1σ) are computed at grid points simultaneously falling within the Lili domain and within the LC, as defined, every 10 days for the entire 10-yr record (Fig. 24, top). Over the entire record, the mean current speed is $0.72 \pm 0.12$ m s$^{-1}$, although significant local excursions clearly exist. Given this observed variability, changes in currents over the time period that the in situ measurements were acquired seems quite plausible, as indicated in the 1-yr time series centered on the time Lili passed (Fig. 24, bottom). To examine this, the autocorrelation of the derived current time series is shown in Fig. 25 for lags from zero to 2 yr. Evidently, there is a fairly rapid decorrelation of the current speed such that less than 50% of the variance is retained after around 20 days. No significant correlation...
is found beyond around 6 months, although there is apparently a small return associated with a relatively weak annual cycle.

The OML mechanical energy cannot be directly computed from the altimeter data because there is no direct estimate of the OML depth. However, the in situ observations of OML depth and associated errors can be combined with the altimeter data to estimate the variability of $K$ in the LC over the relatively long-period record examined here. Using the estimate for the mean geostrophic surface current derived from the $\eta$ fields and a mean OML depth of $60 \pm 10$ m based on the observed profiles, the mean OML LC kinetic energy is $16.5 \pm 6.1$ kJ m$^{-2}$, which compares quite well with the values presented in Table 6, determined from observations obtained only around 10 days apart.

Finally, energy spectra for the LC are computed from the 10-yr record, specifically to examine intraannual variability. First, the record is broken into ten 64 data point records, each with 50% overlap. Each individual record is detrended and multiplied by a Hanning data window before computing the Fourier transform. The 10 resulting energy spectra are then averaged to yield spectral estimates with approximately 40 degrees of freedom (Fig. 26). Energy peaks are found centered at 0.0109 and 0.0281 cycles per day (cpd), which correspond to periods of approximately 90 and 35 days, respectively. Of particular interest here is the higher-frequency, shorter-period energy peak, which suggests that LC structure should be temporally resolved on at least a monthly cycle for accurate coupled atmosphere–ocean tropical cyclone forecasts.

5. Summary and conclusions

For the first time, the Gulf of Mexico Loop Current response to a major hurricane has been directly observed with a comprehensive set of in situ profilers. The combination of current, temperature, and salinity three-dimensional synopses before, during, and after Hurricane
Lili’s passage in 2002 provided an adequate description of upper-ocean mechanical and thermal energy evolution. Contrary to most previous poststorm ocean temperature response studies, which generally showed 3–5°C SST cooling, in situ and satellite observations indicated SST cooling of \(1\)\(^\circ\)C after Lili’s passage over the LC. This limited surface cooling may have allowed a large ocean-to-atmosphere enthalpy flux to be maintained, possibly positively contributing to Lili undergoing rapid intensification in the southeast Gulf of Mexico.

Although Lili was nearly a Saffir–Simpson category-3 hurricane at the time of observation with maximum surface winds of 50 m s\(^{-1}\), its small size (\(R_{\text{max}} = 20\) km) and rapid translation speed (\(V_s = 7\) m s\(^{-1}\)) might have somewhat limited the energetic response. However, the observational analyses revealed an initially unexpected result that the LC lost kinetic energy in the region traversed by Lili. Previous studies of TC ocean response have generally found strong near-inertial currents after storm passage, and scaling arguments suggested a 16.5 kJ m\(^{-2}\) energy response within the LC, based on storm parameters and the initial observed OML depth. In contrast, LC mixed layer mean kinetic energy was observed to weaken by 5.8 kJ m\(^{-2}\) between \(-2\) and \(+4R_{\text{max}}\). As ageostrophic currents within the LC were found to increase slightly, much of this kinetic loss appears to correlate with a decrease in OML geostrophic current, suggesting a change in kinetic energy independent of storm forcing.

To place the observations within the context of background LC variability, upper-ocean currents were derived from satellite altimetry sea surface topography data. Derived current speeds in the LC were found to vary ±0.12 m s\(^{-1}\) (1σ) over 10-day periods. OML mean currents estimated from in situ profiles indicated a decrease of 0.12 m s\(^{-1}\) from prestorm to poststorm in the LC. Thus, it appears plausible that the experimental measurements merely revealed background LC variability and that any response to Hurricane Lili was either mitigated in or quickly removed from the LC region by the time poststorm observations were made around two local inertial periods after storm passage.

Particularly relevant to the motivating research goal of improving TC intensity forecasts is more accurately specifying the total (mechanical and thermal) energy exchange across the air–sea interface. The two most important quantities representing these processes are the wind stress and the enthalpy flux. Unfortunately, these quantities represent turbulent processes and, for practical purposes, must be parameterized in terms of mean variables in numerical weather forecast models.

![Figure 26. Variance-preserving energy spectrum of LC current magnitude within Lili observational domain. Dashed lines are the 95% confidence interval based on a \(x^2\) test.](imageURL)

where \(\delta T\) is the discrete temperature jump between the OML and the seasonal thermocline below and \(\beta\) is a non-dimensional number proportional to the bulk Richardson number. Under approximately horizontally homogeneous initial conditions, this process is typically the dominant OML cooling mechanism. As the ocean-to-atmosphere enthalpy flux in a TC is highly sensitive to the SST (or, more precisely, the air–sea enthalpy difference), variability in subsurface structure, such as observed here, may have a profound effect on a TC’s intensity.

These results shed light on the possibility that the LC, or any other large current system for that matter, simply does not respond significantly to TC forcing (at least for storms of Lili-like structure). Although the time scales over which the altimetry-based current variations are found to be somewhat longer than the prestorm to poststorm in situ synopses, it is apparent that structural changes on the order of weeks, at a minimum, must be taken into account in further studies. It is rather fortunate the LC was found to weaken over this particular experiment’s time period, as a strengthening LC could very well have occurred, leading to a different and possibly
erroneous conclusion about storm/current interactions. Given the tremendous complexity in accurately predicting TC intensity, high-resolution coupled atmosphere–wave–ocean numerical models will be increasingly utilized by operational centers to issue public forecasts. It is imperative that, in conjunction with improvements to the forecast models, in situ datasets continue to be obtained under a broad range of conditions to help address outstanding questions about physical and dynamical processes in the TC environment.

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