Loop Current Response to Hurricanes Isidore and Lili

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(Manuscript received 8 February 2007, in final form 17 December 2007)

ABSTRACT

Recent hurricane activity over the Gulf of Mexico basin has underscored the importance of the Loop Current (LC) and its deep, warm thermal structure on hurricane intensity. During Hurricanes Isidore and Lili in 2002, research flights were conducted from both National Oceanic and Atmospheric Administration (NOAA) WP-3D aircraft to observe pre-, in- and poststorm ocean conditions using airborne expendable ocean profilers to measure temperature, salinity, and current structure. Atmospheric thermodynamic and wind profiles and remotely sensed surface winds were concurrently acquired as each storm moved over the LC.

Observed upper-ocean cooling was about 1°C as Isidore moved across the Yucatan Straits at a speed of 4 m s⁻¹. Given prestorm ocean heat content (OHC) levels exceeding 100 kJ cm⁻² in the LC (current velocities ≥1 m s⁻¹), significant cooling and deepening of the ocean mixed layer (OML) did not occur in the straits. Estimated surface enthalpy flux at Isidore’s eyewall was 1.8 kW m⁻², where the maximum observed wind was 49 m s⁻¹. Spatially integrating these surface enthalpy fluxes suggested a maximum surface heat loss of 9.5 kJ cm⁻² at the eyewall. Over the Yucatan Shelf, observed ocean cooling of 4.5°C was caused by upwelling processes induced by wind stress and an offshore wind-driven transport. During Hurricane Lili, ocean cooling in the LC was ~1°C but more than 2°C in the Gulf Common Water, where the maximum estimated surface enthalpy flux was 1.4 kW m⁻², associated with peak surface winds of 51 m s⁻¹. Because of Lili’s asymmetric structure and rapid translational speed of 7 m s⁻¹, the maximum surface heat loss resulting from the surface enthalpy flux was less than 5 kJ cm⁻².

In both hurricanes, the weak ocean thermal response in the LC was primarily due to the lack of energetic near-inertial current shears that develop across the thin OML observed in quiescent regimes. Bulk Richardson numbers remained well above criticality because of the strength of the upper-ocean horizontal pressure gradient that forces northward current and thermal advection of warm water distributed over deep layers. As these oceanic regimes are resistant to shear-induced mixing, hurricanes experience a more sustained surface enthalpy flux compared to storms moving over shallow quiescent mixed layers. Because ocean cooling levels induced by hurricane force winds depend on the underlying oceanic regimes, features must be accurately initialized in coupled forecast models.

1. Introduction

Coupled models that predict hurricane intensity and structure change are being used to issue forecasts to the public, who will increasingly rely on the most advanced weather forecasting systems to prepare for landfalling systems (Marks and Shay 1998; Bender and Ginis 2000). For such models, it has become increasingly clear over the past decade that the oceanic component will have to include realistic initial conditions to simulate not only the oceanic response to hurricane forcing (Price 1981; Sanford et al. 1987; Shay et al. 1992; Price et al. 1994; D’Asaro 2003) but also the atmospheric response to oceanic forcing (Shay et al. 2000; Hong et al. 2000; Lin et al. 2005; Walker et al. 2005; Wu et al. 2007; Shay 2008).
An important example of this latter effect was observed during Hurricane Opal’s passage in 1995, when atmospheric conditions were conducive for Opal’s rapid deepening over 14 h over the Gulf of Mexico (GOM; Bosart et al. 2000). During this deepening process, Opal passed over a warm core ring (WCR) shed earlier by the Loop Current (LC) as detected by radar altimeter measurements of the surface height anomaly (SHA) fields from the National Aeronautics and Space Administration’s (NASA’s) Ocean Topography Experiment (TOPEX)/Poseidon mission (Shay et al. 2000). Although satellite-derived images revealed that sea surface temperatures (SSTs) were 29.5° to 30°C, there was little evidence of this warm ocean feature’s signature compared to the surrounding Gulf Common Water (GCW). Using a coupled model, Hong et al. (2000) performed a series of sensitivity tests with and without this observed WCR. They found that Opal deepened an additional 14 mb over the WCR compared to numerical experiments without it. Walker et al. (2005) found that cold core rings located on the periphery of the larger WCR helped to weaken Hurricane Ivan (2004) just prior to landfall. More recently, Shay (2008) showed that the LC and WCR did not significantly cool during the passage of Hurricanes Katrina (2005) and Rita when these hurricanes rapidly deepened to Category 5 status. These studies emphasize the importance of initializing models with realistic ocean features to couple to hurricane forecasting models (Jacob and Shay 2003; Falkovich et al. 2005; Halliwell et al. 2008).

The upper ocean’s transport from the northwest Caribbean Sea and through the Yucatan Straits significantly influences the GOM circulation patterns. These transports, ~24 Sv (1 Sv = 10⁶ m³ s⁻¹) through the straits, force LC variability and modulate WCR shedding events (Maul 1977; Sturges and Leben 2000; Leben 2005). The LC transports warm subtropical water with a markedly different temperature and salinity structure into the GOM compared to the GCW (Shay et al. 1998). As the LC intrudes north of 25°N, WCRs with diameters of 100–200 km separate from the LC at an average interval of 6–11 months, based on radar altimeter-derived SHA fields (Sturges and Leben 2000). In contrast, when the LC retracts south of 25°N, the time envelope for WCR shedding events increases to an average of more than 17 months (Leben 2005). Regardless of the northward LC penetration, these anticyclonically rotating WCRs propagate westward at speeds of 3–5 km day⁻¹ (Elliot 1982). Note that both the LC and WCR features contain upper-ocean currents of up to 1.7 m s⁻¹ (Forristall et al. 1992; Oey et al. 2005). At any given time, the GOM may have two or three WCRs embedded within its circulation pattern, with smaller-scale cold core rings located along their periphery. The anticyclonic circulation around the LC exits the GOM through the Florida Straits between the United States and Cuba to form the Florida Current and, eventually, the Gulf Stream. These ribbons of deeper and warmer ocean current features transport heat poleward, representing an integral part of the gyre circulation (Gill 1982).

Investigating a central question about upper-ocean heat, Leipper and Volgenau (1972) developed a relationship to estimate the hurricane heat potential or ocean heat content (OHC), namely,

$$ Q = c_p \int_{h_{26}}^{\eta} \rho(z) [T(z) - 26] \, dz, \quad (1) $$

where $c_p$ is specific heat at constant pressure (4.2 kJ kg⁻¹ K⁻¹), $\rho(z)$ is the density structure, the observed temperature is $T(z)$, and integration limits stretch from the depth of the 26°C isotherm ($h_{26}$) to the surface ($\eta$). In subtropical regimes such as the LC, OHC values exceed 100 kJ cm⁻² (Leipper and Volgenau 1972). That is, the 20° and 26°C isotherm depths are located at ~300- and 150-m depths in this subtropical water mass, compared with ~100- and 50-m depths, respectively, in the GCW.

To improve our understanding of the LC response to the passage of a mature hurricane, a series of experiments was conducted from National Oceanic and Atmospheric Administration (NOAA) WP-3D research flights (N42RF, N43RF) and the NOAA Gulfstream-IV (N49RF) aircraft during the passage of Hurricanes Isidore and Lili. The experimental sampling strategy was designed to deploy global positioning system (GPS) sondes (Hock and Franklin 1999), airborne expendable current profilers (AXCPs), airborne expendable conductivity, temperature, and depth profilers (AXCTDs), and airborne expendable bathythermographs (AXBTs) prior to, during, and subsequent to hurricane passage. This experimental effort in Hurricanes Isidore and Lili improved upon a previous Hurricane Gilbert experiment (Shay et al. 1992) by measuring prestorm, in-storm, and poststorm currents, temperatures, and salinities along with detailed atmospheric temperature, humidity, and wind soundings (Table 1). The objective of this paper is to document the evolving thermal and momentum ocean response to the heat, moisture, and momentum fluxes across the air-sea interface based on the combination of GPS sonde profiler data (Hock and Franklin 1999) and remotely sensed surface winds from the Stepped Frequency Microwave Radiometer (SFMR; Uhlhorn et al.
To accomplish this objective, this paper is organized as follows: a description of the profiler data, including chronologies of Hurricanes Isidore and Lili, is in section 2; the experimental approach, including the surface wind forcing and air–sea parameters, is described in section 3; the observational analysis and air–sea fluxes are discussed in section 4; section 5 describes the forced LC response with respect to ocean cooling and mixing; and the results are summarized, together with concluding remarks, in section 6.

2. Hurricane chronology

a. Hurricane Isidore

The tropical wave from which Isidore developed originated over Cape Verde, where several periods of fluctuating storm intensity occurred until this wave crossed the 50°W meridian (Pasch et al. 2004). However, only when the tropical depression moved into the western Caribbean Sea was it named Tropical Storm Isidore on 17 September 2002. Isidore then moved slowly along a northwest track, and at 1800 UTC 19 September, Isidore was upgraded to a hurricane with a central pressure of 983 mb. Isidore’s winds exceeded 40 m s⁻¹ on 20 September as it approached the Isle of Youth and made landfall along the tip of western Cuba. The storm remained over Cuba for about 12 h, re-emerging over the Yucatan Straits midday on 21 September. This slow-moving hurricane (~4 m s⁻¹) traveled westward (Fig. 2), intensifying to a Category 3 storm just north of the Yucatan Straits over the LC (see Fig. 1). Isidore moved over the Yucatan Shelf on 22 September and made landfall on the Yucatan Peninsula for the next 36 h, weakening to a minimal tropical storm (TS) embedded within a broad atmospheric circulation pattern. Isidore subsequently moved northward and created a cool wake of 28.5°C SSTs across the central GOM (Fig. 1b), making landfall on 26 September just west of Grande Isle, Louisiana.

b. Hurricane Lili

Lili was also a tropical wave of Cape Verdean origin, starting on 16 September 2002 (Pasch et al. 2004). This wave became a tropical depression on 21 September, and as the system moved just west of north at ~10 m s⁻¹, initial intensification to TS status occurred on 23 September. The TS subsequently weakened to an open tropical wave on 26 September, but as the wave slowed it redeveloped into a TS late on 27 September, with a minimum central pressure of 994 mb. Lili intensified to hurricane status at 1200 UTC 30 September while passing over the Cayman Islands. As Lili tracked along a north-northwest trajectory after emerging off the Cuban coast (Fig. 2), the hurricane intensified to Category 3 status (~51 m s⁻¹) over the LC and to a Category 4 storm (61 m s⁻¹) in the south-central GOM just north of the boundary between the LC and the GCW. During this period, Lili’s radius of maximum winds ($R_{max}$) decreased from 25 km over the LC to 18 km along the northern boundary between the LC and the GCW when the storm system was moving at 7 m s⁻¹. Lili rapidly weakened to Category 1 status owing to a combination of enhanced atmospheric shear, the intrusion of dry air along the western edge (Pasch et al. 2004), and interactions with the shelf water cooled by Isidore. Hurricane Lili made landfall at 1300 UTC 3 October near Intracoastal City, Louisiana.

3. Experimental approach

During the 2002 NOAA Hurricane Research Division’s (HRD’s) hurricane field program, a joint National Science Foundation (NSF) and NOAA experiment measured both the kinematic and thermodynamic upper-ocean response to a propagating mature tropical
cyclone over the LC. Motivated by the Hurricane Opal case, the experimental objective was to measure the levels of upper-ocean cooling and shear-induced mixing in the LC circulation system. To achieve this objective, the experiment used 16 research flights, each deploying oceanic and atmospheric expendable probes in the same location before, during, and after the passage of Hurricanes Isidore and Lili (Table 1). A set of prestorm flights was conducted during 18–23 September 2002; in-storm flights occurred on 21 September and 2 October; and poststorm surveys were acquired on 22 and 23 September and 4 October. In addition, there were also two landfall experiments on 22 September and 3 October off the Yucatan Peninsula and along the Louisiana Coast for Isidore and Lili, respectively. Success rates for the oceanic profilers, defined here as receiving radio frequency (RF) signals on the aircraft, were greater than 85%. GPS sondes were also concurrently deployed from the aircraft, including the G-IV used to map the regional-scale atmospheric structure over the GOM from flight level to the surface in the storm and the surrounding environment.

The comprehensive set of measurements included both in situ and remotely sensed data. The data acquired included current and temperature profiles from AXCPs, temperature and salinity profiles from

![Fig. 1. (a) LC (red contour) position relative to the best tracks of Isidore (black track, originating in bottom center) and Lili (black track, originating in bottom-right corner) in September–October 2002 with intensities (see legend) and the location of NDBC 42001 (blue dot). (b) Observed SSTs (°C) at NDBC 42001 with gray shading indicating lifetimes of Isidore and Lili.](image-url)
FIG. 2. HRD H*Wind surface wind analysis of Hurricanes (a) Isidore at 2130 UTC 21 Sep 2002 and (b) Lili at 0700 UTC 02 Oct 2002. Isotachs are contoured every 5 m s$^{-1}$. Data used to generate this analysis include observations from the SFMR, GPS dropwindsondes, QuikSCAT scatterometer, and available hourly buoy reports. The storm track is indicated by the solid line; the dark box shows the ocean data analysis region considered for this research.
AXCTDs, temperature profiles from AXBTs, and surface directional wave spectra from the NASA scanning radar altimeter (Wright et al. 2001). Note that although the AXCPs, AXCTDs, and AXBTs all measure temperature, the temperature measurements only penetrate to 350 m for the AXBTs compared to 1000 and 1500 m for AXCTDs and AXCPs, respectively. A second difference is that temperatures from AXBTs and AXCPs are measured with a thermistor with an accuracy of 0.2°C; an AXCTD’s thermistor accuracy is 0.12°C. In this study, SSTs from all expendable profilers are defined to be near-surface temperatures within the first few meters of the sea surface. Also, deeper profiles allow estimates of geostrophically balanced currents relative to 750 m for assessing initialization schemes used in ocean models (Halliwell et al. 2008). Finally, the surface wind field was measured from observations by GPS sondes (Hock and Franklin 1999) and SFMR (Uhlhorn et al. 2007). These in situ data are cast within parameters used with the Hurricane Gilbert dataset.

a. Isidore

Given uncertainties in Isidore’s track, two grids of prestorm ocean profilers were deployed from research aircraft in the northwest Caribbean Sea and in the south-central part of the GOM on 18 and 19 September 2002 (Figs. 3a,b). Success rates in acquiring profiles were ~75% from oceanic probes. In some cases, the compound surrounding the thermistor was compromised below 100 m because of pressure effects that caused bad temperature profiles from AXCPs. A few AXCTDs also had incorrect calibration coefficients and were not used in the analyses below. On each of the prestorm flights, there were eight probes with no RF signals. During the dual-aircraft Isidore flight on 21 September (Fig. 3c), there were eight failures out of 63 deployed profilers, and 19 additional AXBTs deployed from N42RF with 5 RF failures. Several AXCPs malfunctioned along the western boundary of the Yucatan Straits, where large currents and current shears caused the thin wire connecting the probe to the surface unit to break. Given the dual-aircraft mission, this still produced unprecedented coverage of the upper ocean as a hurricane intensified to Category 3 status. On 22 September, there were 29 additional AXBTs deployed over the same regime where the in-storm flights deployed profilers, including over the Yucatan Shelf. During the poststorm experiment on 23 September (~2 days later), 64 probes were deployed along transects similar to those of the prestorm and in-storm flights. Seven of these profilers did not transmit data back to the aircraft.

b. Lili

One week later, on 28 September 2002, TS Lili moved to the northwest toward the Yucatan Straits following a track similar to Isidore. During this time, a prestorm flight was conducted in front of the projected TS Lili (and potential hurricane) track on 29 September (not shown) by deploying AXBTs. On 30 September, additional profilers were deployed from an in-storm flight centered on Lili at 20°N, 81°W, including six AXCPs with four probes providing profiler data (not shown). As Lili continued along this northwest trajectory, moving over the western tip of Cuba as a Category 1 storm, the in-storm flight on 2 October (Fig. 3f) was centered fortuitously on the pre-Isidore grid of 19 September (Fig. 3b). On this research flight, 63 profilers were deployed, with 12 of them not providing any RF signals to the aircraft. It was during this early flight on N43RF and the later N42RF flight on 2 October [in support of a NOAA/National Environmental Satellite, Data, and Information Service (NESDIS) ocean winds experiment deploying AXBTs] that Lili deepened to a Category 4 storm just northwest of the boundary between the LC and the GCW in the central GOM basin. A post-Lili experiment was then conducted on 4 October by deploying the same number of profilers, with only five RF failures. Thus, these oceanic and atmospheric measurements were acquired when two hurricanes, separated by 10 days, were intensifying over the same oceanographic regime.

c. Air-sea parameters

Air–sea parameters and scaling arguments, defined in Table 2, are used to place the observations into a nondimensional framework based on Price (1983). The wavelength of the oceanic response induced by a moving tropical cyclone is proportional to the product of the storm translation speed \( U_h \) and the local inertial period (IP; Geisler 1970). Based on a 4 m s\(^{-1}\) translation speed and an IP of 1.3 days (~31 h), the predicted wavelength \( \Lambda \) for Isidore is 450 km (Table 2). For Lili, this wavelength is ~770 km because of faster translation speeds of ~7 m s\(^{-1}\) and IPs ranging from 1.3 days (31 h) to 1.16 days (28 h) over the LC and GCW, respectively. The \( R_{\text{max}} \) for these storms over the LC
FIG. 3. SHA field (colors indicate heights in cm) from radar altimetry and oceanic profile deployments for (a) pre-Isidore in the northwest Caribbean Sea, (b) pre-Isidore in the south central GOM, (c) Isidore in-storm across the Yucatan Straits, (d) post-Isidore (1 day), (e) post-Isidore (2 days), (f) Lili in-storm over the grid in (b), and (g) post-Lili on 4 October relative to the Isidore and Lili tracks with hurricane category levels as in Fig. 1. AXBT data are indicated by triangles; AXCP, by boxes; AXCTD, by circles. Black symbols indicate good data; white symbols indicate probe failure.
The first baroclinic mode wave phase speed in the LC is approximately 1.5 m s\(^{-1}\); in the GCW, \(R_{\text{max}}\) decreased to 18 km during a deepening cycle to a Category 4 storm. In this context, along- and cross-track directions are nondimensionalized in terms of \(\Lambda\) and \(R_{\text{max}}\) to examine the forced ocean response relative to the observed storm structure (Shay et al. 1998).

The first baroclinic mode wave phase speed in the LC regime compared, with 0.3 IPs in the GCW. This more rapid time scale is due to a decrease in \(R_{\text{max}}\) coupled with a faster phase speed when Lili moved over the GCW. Note that this time scale was \(O(2 \text{ IPs})\) during Gilbert due to an \(R_{\text{max}}\) that was 3 times larger than that observed in either Isidore or Lili (Shay et al. 1992).

Isotherm displacements (\(\eta\)) scale as \(\tau_{\text{max}}/(\rho_o f U_h)\), about 11 m (6 m) in the expected Isidore (Lili) ocean response. The geostrophic velocity response \(V_{\text{gr}}\) proportional to \(g' \eta/(f R_{\text{max}})\), is predicted to be weak, with values of 1 to 3 cm s\(^{-1}\) in the LC compared with 5 cm s\(^{-1}\) in the GCW, consistent with the Gilbert dataset (Shay et al. 1992). Wind-driven ocean velo-
ity \( V_{ml} \) scales as \( \tau_{max} R_{max} / (\rho \cdot h U_i) \) representing a velocity of 24–38 cm s\(^{-1}\), with similar values in the thermocline based on the expression \( V_{ml} h / b \) (where \( h \) is the initial OML depth and \( b \) is the thermocline thickness). Weak upper-ocean stratification distributed over deeper layers will cause the LC response to be weaker than in the more stratified GCW (Price 1981; Shay and Elsberry 1987; Shay et al. 1992). In the GCW, however, the predicted wind-forced thermocline current is 12 cm s\(^{-1}\) because of stronger stratification and shallower OML depths. The magnitude of the ocean response depends not only on the characteristics of the hurricane forcing but also, crucially, on the background initial ocean conditions based on these scaling arguments. Price (1983) defined a mixed-layer Burger number as

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

where \( S \) is the nondimensional storm speed \( U_i / (2R_{max} f) \) and \( h \) is the OML depth. For the LC, \( M \) ranged between 0.04 and 0.08 because of large values of \( S \) for Lili (0.8) and for Isidore (0.54). Over the GCW, Lili’s acceleration toward the northwest coupled with a smaller \( R_{max} \) caused \( S \) to increase to 1.15. The thermocline Burger numbers \( T = h / b M \) are nearly equal to \( M \) in the LC because the ratio of thermocline and OML thicknesses is \( O(1) \), as noted above. In the GCW, however, \( T \) is 0.2 (i.e., a magnitude larger than in the LC), representing more dynamical coupling between the OML and thermocline layer. The resultant blue shift in the mixed-layer inertial frequency is proportional to \( M/2 \) (Table 2), equating to frequencies shifted above \( f \) by 2% to 4%, consistent with previous results (Shay et al. 1998).

4. Analysis approach

a. Orbital velocities

To examine the current and shear measurements from the AXCPs, the surface wave–induced orbital velocity signals must be removed from the current profiles observed from the in-storm flights. When averaged over a cycle, the resolved low-frequency surface waves do not contribute significantly to the mean current (with the exception of a small Stokes drift due to non-linear wave structure). Observed current profiles are fit to the Sanford et al. (1987) three-layer model with an assumed monochromatic, linear, deep-water surface gravity wave superimposed; thus,

\[
u(z) = \left[ C \cos \left( \frac{\omega z}{W} \right) + S \sin \left( \frac{\omega z}{W} \right) \right] e^{i z} + U_i + S_z \left[ z - \frac{(Z_{i-1} - Z_i)}{2} \right], \text{ for } (i = 1, 3),
\]

where \( u(z) \) is the modeled east–west current profile [a similar expression holds for the north–south component \( v(z) \)], \( C \) and \( S \) are the amplitudes of the orbital velocities associated with surface waves, \( \omega = \sqrt{\frac{\rho g}{\rho}} \) is the wave frequency for wavenumber \( k \) following from linear theory, \( W = -4.5 \text{ m s}^{-1} \) is the AXCP fall rate, and \( U_i \) and \( S_z \) are the mean current and shear in layer \( i \), respectively. Observed current profiles are fit to (4) using a standard Levenberg–Marquardt nonlinear least squares regression (Marquardt 1963) by minimizing the error over a range of trial wave periods \( T = 7–14 \text{ s} \) (\( \omega = 2 \pi / T \)). An additional constraint is that the current profiles are continuous across layer interfaces, thereby reducing the number of free parameters from eight to six.

Model fits to the Isidore (21, 23 September) and Lili (2 October) current profiles are listed in Tables 3, 4, and 5, respectively. In-storm orbital velocity amplitudes were typically 1 m s\(^{-1}\), in accord with previous experimental efforts (Sanford et al. 1987). As shown in Fig. 4, the RMS residual currents, estimated by calculating differences between the model (4) and the observed profile to a depth where \( k z = -2 \) (i.e., where wave amplitudes are reduced to about 13% of their surface velocity amplitudes based on the fits), were generally less than 10 cm s\(^{-1}\), indicative of a reasonably good fit for strongly forced mixed layer conditions. However, there were a few exceptions with relatively large residuals. Vertical shears of the horizontal velocity components in the upper two layers (\( Z_1, Z_2 \)) were \( O(10^{-2} \text{ s}^{-1}) \), whereas the shears in the lower layer (\( Z_3 \)) decreased by an order of magnitude. The orbital velocity amplitudes have been removed from the observed current profiles to examine the observed upper-ocean response (Shay et al. 1992; Price et al. 1994).

b. Objective analyses

Objectively analyzed fields of observed variables are produced using a statistical interpolation methodology. The analysis procedure used here is the OAXS package (developed at Canada Bedford Institute of Oceanography), which is based on the algorithm presented in Bretherton et al. (1976). This approach uses a linear optimal interpolation technique to estimate values at
grid points based on observations at the nearest neighbors. A covariance model of the form (Freeland and Gould 1976)

$$\rho(r) = e^{-r^2 / (1 + r^2 / 3)}$$  \(5\)

is used, where \(r\) is the weighted nondimensional distance between an observation and a grid point. In the absence of detailed climatological information available about LC structural and temporal variability over the scales of interest exposed to severe forcing, a quantitative development of the optimal parameter scales is difficult (e.g., Baker et al. 1987). Therefore, scales are chosen through trial and error to adequately resolve the mesoscale ocean structure. Because the largest variability is observed near the surface during hurricane passage, spatial scales here are smaller, and they increase with depth (Table 6). The horizontal covariance model \(\rho(r)\) is plotted for each layer as a function of dimen-

### Table 3. Coefficients from fits with the Sanford et al. (1987) model and the Isidore storm AXCP profiles in the upper 200 m where \(Z_0\) is the start depth of the good data used in the fit; \(T\) is the period of the surface wave with coefficients of \(C\) and \(S\); \(Z_{1,2,3}\), \(V_{1,2,3}\), and \(S_{1,2,3} \times 10^{-2}\) represent layer depth, layer-averaged currents, and current gradients in each layer, respectively; and \(R\) is the residual current not explained by the model to a depth of \(e^{-2}\).  

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The Isidore (Lili) grids are rotated 270° longitude, respectively. The vertical grid contains 151 uses an 3 x 3 grid structure used for all Lili (Isidore) analyses is 
nominal measurement errors of surface temperatures, 0.2 °C (not shown) over the remainder of the domain.
Assuming properly chosen scales for the interpolation, uncertainty estimates for each gridded field are available based upon the input measurement noise. For nominal measurement errors of surface temperatures, maximum mapping errors are 0.4°C in the northwest part of the pre-Lili grid because of data sparsity in that region compared to temperature mapping errors of 0.2°C (not shown) over the remainder of the domain. Similarly, mapping errors for 26°C isotherm depths range from 1 to 3 m, whereas those associated with OHC values have a maximum of 12 kJ cm⁻². Finally,
current mapping errors are typically in the realm of 0.1–0.2 m s⁻¹, with the larger values located in the northwest corner of the pre-Lili domain. In all of the mapped fields, these mapping errors are much less than those in the observed oceanic signals (e.g., high signal-to-noise ratios).

c. Air–sea fluxes

Sea surface forcing is described by the fluxes of momentum, heat, and moisture. These fluxes are estimated from bulk formulas utilizing near-surface (10 m) atmospheric thermodynamic and wind measurements and upper-ocean thermal data. Atmospheric data are measured from the large number of GPS sondes deployed within the storm from both aircraft (Table 1), including sondes deployed from Air Force Reserve reconnaissance and NOAA G-IV synoptic flights. Surface winds are estimated from SFMR measurements by sensing brightness temperatures at multiple frequencies, as thoroughly described in Uhlhorn et al. (2007).

From each GPS dropwindsonde (Hock and Franklin 1999), 10-m values of temperature and specific humidity were acquired and objectively analyzed using the Bretherton et al. (1976) OI method projected onto a storm-relative grid aligned with the direction of storm motion. SFMR wind observations are objectively analyzed using HRD’s H²Wind system (Powell and Houston 1996) and then interpolated bilinearly to a storm-

| Time UTC | Variable | T | C | S | Z₀ | V₁ | S₁ | Z₁ | V₂ | S₂ | Z₂ | V₃ | S₃ | Z₃ | R |
|----------|----------|---|---|---|----|----|----|----|----|----|----|----|----|----|
| 1755     | u        | 8.0| 211| 44| -9 | 22 | 1.46| -35| 7  | -0.16| -90 | 10 | 0.02| -200| 7.6|
| 1807     | u        | 8.5| 64 | 41| -9 | -6 | 0.91| -35| -13| -0.17| -90 | -7 | -0.03| -200| 3.2|
| 1820     | u        | 11.0| 18 | 4 | -8 | 114| -0.02| -60| 79 | 1.17| -120| 40 | 0.09| -200| 3.2|
| 1846     | u        | 9.0| -39| 6 | -8 | -28| 0.48| -60| -16| -0.83| -120| 1  | 0.20| -200| 9.2|
| 1929     | u        | 8.5| -23| 31 | -15| -6 | -1.51| -50| 11 | 0.27| -120| -4 | 0.14| -200| 3.5|
| 1946     | u        | 7.5| 37 | 143| -10| 53 | 3.01| -25| 14 | 1.28| -50 | -1 | -0.03| -200| 7.2|
| 2003     | u        | 8.0| 14 | -17| -10| 29 | -0.13| -70| 22 | 0.72| -100| 8  | 0.06| -200| 5.7|
| 2047     | u        | 7.0| 4  | 2 | -10| 92 | -0.09| -70| 78 | 1.10| -100| 46 | 0.30| -200| 4.7|
| 2055     | u        | 13.0| -17| 5 | -10| -20| -0.23| -80| -6 | -0.33| -120| 10 | -0.21| -200| 3.5|
| 2108     | u        | 9.0| -20| 19 | -15| 34 | 0.02| -70| 24 | 0.26| -120| 15 | 0.00| -200| 2.3|
| 2117     | u        | 13.0| 50 | 66 | -15| 35 | 1.11| -70| 15 | -0.30| -140| 25 | 0.01| -200| 4.7|
| 2125     | u        | 8.5| 32 | 6 | -15| 65 | 0.50| -70| 48 | 0.17| -100| 34 | 0.37| -160| 2.5|
| 2233     | u        | 9.0| 3  | 3 | -15| 67 | 0.62| -70| 33 | 1.11| -100| 6  | 0.36| -160| 2.6|
| 2318     | u        | 10.0| 6 | -27| -15| -5 | 5.02| -25| -25| -0.38| -50 | -22 | 0.19| -65 | 2.2|
| 2333     | u        | 9.0| -56| 18 | -15| 15 | -1.94| -25| 6  | 1.44| -50 | -18 | 0.87| -65 | 1.2|
| 0007     | u        | 10.0| -9 | -6| -22| 1  | -0.10| -50| 0  | 0.10| -110| -4 | 0.02| -200| 5.1|
| 0020     | u        | 12.0| 223| -42| -22| 36 | 6.45| -50| -24| -1.02| -110| 13 | -0.14| -200| 11.1|
| 2007     | u        | 9.0| 8  | -8| -22| 0.23| -45| -10| -0.92| -80 | -5 | 0.19| -200| 10.2|
| 2135     | u        | 8.0| 21 | 17 | -8| 14 | -1.14| -45| 33 | 0.12| -80 | 16 | 0.26| -200| 23.7|
| 2233     | u        | 8.0| 92 | 36 | -10| -42| -1.65| -60| -4 | 0.19| -100| 0  | -0.15| -200| 4.9|
| 2318     | u        | 10.0| 58| -17| 87 | 0.15| -60| 89 | -0.30| -100| 62 | 0.67| -200| 10.9|
| 2333     | u        | 11.0| 12| 0 | -12| 47 | 0.24| -80| 30 | 0.44| -120| 22 | -0.03| -200| 3.5|
| 0007     | u        | 10.0| 3 | 15 | -12| 21 | 0.35| -80| -29| -0.23| -120| -38 | 0.35| -200| 3.3|
| 0020     | u        | 8.0| 18 | 64 | -10| -23| 1.10| -60| -21| -1.48| -100| 0  | 0.17| -200| 5.9|
| 0007     | u        | 7.0| -72| -65| -10| 60 | -0.40| -60| 75 | -0.28| -100| 46 | 0.71| -200| 6.7|
| 0020     | u        | 15.0| -25| 5 | -10| -48| 1.45| -25| -37| -0.59| -100| 30 | -0.09| -200| 8.7|
| 0007     | u        | 9.0| -26| 84 | -10| -51| 0.30| -25| -25| -0.75| -100| -3 | 0.14| -200| 5.4|
| 0020     | u        | 8.0| -37| -13| -10| -12| -0.68| -40| 5  | 0.60| -60 | 6  | 0.07| -200| 3.4|
| 0007     | u        | 8.0| 48 | 8 | -10| -38| 2.06| -40| -35| -3.29| -60 | -7 | 0.06| -200| 4.3|
| 0020     | u        | 9.0| 30 | 10 | -8| -52| -0.20| -40| -22| -1.35| -80 | -5 | 0.16| -200| 5.5|
| 0020     | u        | 9.0| -7 | 1 | -8| 79 | 0.18| -40| 60 | 0.82| -80 | 37 | 0.12| -200| 9.1|
| 0020     | u        | 8.5| -48| 20 | -10| -3 | -0.27| -100| 14 | -0.28| -130| 17 | 0.02| -200| 4.8|
| 0020     | u        | 8.0| -8 | 6 | -10| 5  | 0.35| -100| 7  | -1.18| -130| 3  | 0.62| -200| 5.1|

Table 5. Same as Table 3, but for Lili post-storm AXCP profiles.
relative grid as for the GPS atmospheric thermodynamic variables (See Fig. 2). Finally, sea surface temperature observations are also optimally interpolated to the same storm-relative grid (as noted above), resulting in a set of variables at a common location from which the spatial distribution of the bulk surface momentum, sensible, and latent heat fluxes are estimated; that is,

\[
|\tau| = \rho_a C_d |\mathbf{U}_{10}|^2, \quad (6)
\]

\[
Q_s = \rho_a c_p C_h |\mathbf{U}_{10}| \Delta T, \quad \text{and} \quad (7)
\]

\[
Q_l = \rho_a L_v C_q |\mathbf{U}_{10}| \Delta q, \quad (8)
\]

where \(\rho_a\) is the atmospheric density; \(C_d, C_h,\) and \(C_q\) are exchange coefficients of momentum, sensible and latent heat, respectively; \(\mathbf{U}_{10}\) is the 10-m wind speed; \(c_p\) is specific heat of air at constant pressure; \(L_v\) is the latent heat of vaporization; and \(\Delta T (T_s - T_a)\) is the difference between SST and 10-m air temperature. SSTs are defined to be near-surface temperatures within the first few meters of the sea surface from the expendables,\(^1\)

\(1\) The temperature of a bulk OML, defined as the depth where the temperature decreases by more than 0.2°C, is given by a vertical average from near-surface to this depth. During periods of strong wind forcing, the upper ocean is well mixed in temperature, which represents a bulk OML temperature as observed during hurricanes (Sanford et al. 1987; Shay et al. 1992, 2000).

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**Table 6.** Horizontal and vertical correlation scales used in the objective analyses.

<table>
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<tr>
<th>Depth range (m)</th>
<th>Horizontal scale (L (°))</th>
<th>Vertical scale (m)</th>
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<tr>
<td>600–750</td>
<td>5.0</td>
<td>200</td>
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**Fig. 4.** Examples of model fits (solid) using a three-layer approach of Sanford et al. (1987) compared to observed profiles (dots) for (a)–(c) Isidore and (d)–(f) Lili for the (a), (d) \(u\) and (b), (e) \(v\) components (m s\(^{-1}\)). (c), (f) Differences between observed and model profiles for both velocity components normalized by the surface wave amplitudes (from Tables 3 and 4). Note that the RMS differences are smaller in Lili, but because the surface wave is less energetic, the (c) normalized differences for Lili are larger than (f) for the Isidore profiles.
and $\Delta q (q_s - q_a)$ is the difference between the saturated specific humidity at the SST and unsaturated 10-m atmospheric specific humidity. The surface drag coefficient $C_d$ is computed from the Large and Pond (1981) relationship but is capped at a maximum value of $2.5 \times 10^{-3}$, based on recent results indicating a threshold or saturation value of $C_d$ at $28$–$33 \text{ m s}^{-1}$ wind speeds (Powell et al. 2003; Donelan et al. 2004; Shay and Jacob 2006; Jarosz et al. 2007). Heat exchange coefficients ($C_h, C_q$) are set equal to $C_d$, which is conservative compared to the theory proposed by Emanuel (1995) (i.e., Emanuel’s theoretical results suggest that this enthalpy and drag coefficient ratio lies between 1.2 and 1.5 in severe hurricanes). An additional ocean forcing mechanism results from the surface precipitation flux (rain rate). Freshwater input by rain can alter the ocean’s response both by direct cooling (caused by rain at a lower temperature than the SST) and by reducing the salinity, which stabilizes the OML and reduces the rate of vertical mixing (Jacob and Kolinsky 2007). Tropical Rainfall Measuring Mission (TRMM) Microwave Imager (TMI)-derived rain rates (Fig. 6) were used to estimate surface precipitation fluxes for Isidore and Lili (with maximum rain rates of 35 and 20 mm hr$^{-1}$, respectively).

Surface flux distributions during the Isidore in-storm flight (21 September) are shown in Fig. 7. Peak enthalpy flux is found in the right rear quadrant of the storm to be $\approx 1.8 \text{ kW m}^{-2}$ as a result of high SSTs that show negligible decrease from prestorm conditions. The maximum momentum flux ($\approx 7 \text{ N m}^{-2}$) is located in the right front quadrant and is associated with a highly symmetric storm because of its fairly slow $4 \text{ m s}^{-1}$ translational speed. Similarly, the estimated surface fluxes in Lili are shown in Fig. 8. Lili’s surface wind field (2 October) indicates marked asymmetry resulting from the more rapid storm motion toward the NW ($\approx 7 \text{ m s}^{-1}$), and correspondingly, surface fluxes are enhanced on the right side of the track. Compared to Isidore, the maximum surface enthalpy flux in Lili is weaker ($1.4 \text{ kW m}^{-2}$) despite peak surface winds of 51 m s$^{-1}$. This lower flux results primarily from SSTs observed in the LC regime that are approximately 1°C lower than those during Isidore. This is an important point that highlights how modest surface temperatures differences can effectively alter surface heat fluxes under hurricane wind conditions (Cione and Uhlhorn 2003).

To improve our understanding of how these estimated fluxes relate to sea–air heat exchange, enthalpy (heat and moisture) fluxes are integrated in the along-track direction to obtain a cross-track (radial) distribution of the ocean heat loss through the sea surface. An along-track spatial coordinate is used to convert to

![Fig. 5. Covariance model weighting function $\rho(r)$ as a function of dimensional distance $r \times L$ ($^\circ$) applied to observations for the objective analyses as per Table 6 for differing length scales $L$.](image)

![Fig. 6. Rain rates (mm h$^{-1}$) based on TRMM data during (a) Isidore and (b) Lili.](image)
time, assuming a steadily moving storm based on the observed storm speed (Table 2). Estimated surface heat losses (kJ cm$^{-2}$) for Isidore and Lili are shown in Fig. 9. At the eyewall, surface heat loss in Isidore is 9.5 kJ cm$^{-2}$, compared to 4.5 kJ cm$^{-2}$ during Lili. These differences result from higher enthalpy fluxes (1.8 versus 1.4 kW m$^{-2}$) and slower storm speeds (4 versus 7 m s$^{-1}$) in Hurricane Isidore.

Based on the TRMM data (see Fig. 6), net freshwater input (precipitation minus evaporation, hereafter $P - E$ rate) is estimated for both storms by integrating these data in the along-track direction (Fig. 9b). As suggested by the fluxes, the $P - E$ rate in Isidore was three times larger than in Lili (between $\pm 4R_{\text{max}}$). At levels of 300 mm within Isidore’s core, rain impacted the OML balance through the $P - E$ rate. In contrast, the $P - E$ rate in Lili’s core was 115 mm because of a faster translation speed. Such rain events induce changes in the OML salinity balance of 0.2 to 0.4 practical salinity units (psu), as documented by conductivity, temperature, and depth (CTD) measurements acquired in typhoon wakes in the western Pacific Ocean (Pudov and Petrichenko 2000). Thus, the OML salinity balance and the surface buoyancy flux must be accounted for in ocean response models for light and strong winds (Price et al. 1986; Jacob and Koblinsky 2007).

**d. Temperature and velocity profiles**

Current and temperature profiles from the B’–B transect (see Fig. 3g) along 1.5 to 2 $R_{\text{max}}$ to the right of the track are shown in Fig. 10, 2 IPs ($\pm 63$ h) following Lili’s passage over the domain. Current profiles along the northern part of transect (B’) in the GCW indicated an anticyclonic rotation with depth suggestive of verti-
cal energy propagation out of the wind-forced OML (Leaman 1976). During Gilbert’s passage, this observed anticyclonic rotation with depth was found to 4 times more energetic than the cyclonically rotating component (Shay and Jacob 2006). This current vector rotation forced strong current shears beneath the OML, inducing cooling by entrainment mixing processes. This effect lowers Richardson numbers to below criticality for a deepening and cooling OML (Pollard et al. 1973; Price 1981; Jacob et al. 2000). In the center of the B’–B transect, OML currents approach 1 m s$^{-1}$ flowing toward the east. Notice that the warmer thermal structure approaches 100 m depth where the currents remain relatively constant with depth in the LC. This baroclinic current structure tends to be in geostrophic balance with a current reversal at 500 m and with weaker currents extending to a depth of 1000 m. By contrast, the current structure at point B is shallower, with maximum OML currents of 0.35 m s$^{-1}$.

In the GCW, located in the northwest corner of the analysis domain, OML cooling and deepening are typical signatures of a stronger ocean response that often provide a negative feedback during hurricane passage (Chang and Anthes 1978; Bender and Ginis 2000). However, in the LC, this negative feedback to the atmosphere is minimized compared to that observed over the GCW. Because the assumption of horizontal homogeneity is violated in the LC and WCR regimes, Halliwell et al. (2008) stress the importance of initializing three-dimensional oceanic models to accurately predict intensity from coupled forecasting models. In fact, Falkovich et al. (2005) introduced a numerical approach for feature-based ocean modeling that involves cross-frontal sharpening of the background temperature and salinity and, hence, the density fields. These studies underscore the need for three-dimensional experimental data to improve oceanic model initialization schemes.
e. Temperature and velocity sections

Background ocean flows are set up by horizontal pressure gradients resulting from nonzero temperature and salinity gradients and may play a significant role in affecting the development of strong wind-driven current shears within the LC and WCR regimes (Fig. 11). Pre- and post-Isidore measurements across the Yucatan Straits indicate strong density and pressure gradients associated with the LC. Pre-Isidore measurements suggest a northward-flowing LC of 1 m s$^{-1}$ skewed toward the western boundary of the Yucatan Straits. This is the region where the horizontal density and pressure gradients sharpen because of a strong bottom slope. The initial 26°C isotherm depth was a maximum of ~150-m depth before Lili; the isotherm depth then decreased by about 10 m on 4 October, due in part to upwelling along the track. Spatially averaged baroclinic currents increased to 0.7 m s$^{-1}$ in response to Lili. As in the oceanic response to Isidore, there was little evidence of significant upper-ocean cooling between snapshots. Although the time envelope between the pre- and post-Lili measurements was nearly two weeks, there was not much evidence that this Category 3 hurricane impacted the LC even just 2 days following passage. By contrast, forced near-inertial current shears deepened and cooled the OML by about 4°C in the GCW during Hurricane Gilbert’s passage (Shay et al. 1992).

5. Forced response in the Loop Current

An important question emerging from recent studies is the upper-ocean cooling levels during hurricane passage. Although early studies focused on the concept of negative feedback during hurricane passage (i.e., Chang and Anthes 1978), observational evidence has suggested that the STW mass associated with the LC and WCR does not significantly cool compared to the GCW. Based on profiler measurements before and after Rita, the oceanic cooling was minimized, suggesting negative feedback even though Rita was a Category 5 hurricane over the LC (Shay 2008). Historical records suggest that once a hurricane enters the GOM basin it will likely intensify prior to making a landfall (Marks and Shay 1998). Contrasting the details of the oceanic response differences of the LC and WCR versus the GCW water masses has implications with respect to negative feedback to the atmosphere. The focus here is on examining the observed oceanic response to Hurricanes Isidore and Lili.

a. SST

Pre- and post-Isidore and Lili SST fields are shown in Fig. 13. Pre-Isidore SSTs ranged from 28.5°C to 29.5°C over the experimental domain. During the post-Isidore experiment on 23 September, SST changes were ob-
FIG. 10. Section B–B’ current vector stick plot (cm s\(^{-1}\)) and temperature (°C) at 1.5 \(R_{\text{max}}\) from Lili’s track on 4 Oct (poststorm). Time is scaled in terms of inertial period. Black dots represent the OML depth in the upper panel extending from the surface to 150-m depth.

FIG. 11. (top) Pre- and (bottom) post-Isidore along-track section of temperature (°C, color) and geostrophic velocity (m s\(^{-1}\), dashed contours) across the Yucatan Straits. The heavy dashed black line depicts the depth of the 26°C isotherm.
FIG. 12. Same as Fig. 11, but for (top) pre- and (bottom) post-Lili.

FIG. 13. (a) Pre- and (b) post-Isidore SSTs; (c) ΔSST for Isidore; (d) pre- and (e) post-Lili SSTs; and (f) ΔSST for Lili (all in °C) relative to Isidore’s or Lili’s track (black line) across the Yucatan Straits or the southeast GOM relative to bottom topography (dotted lines) for the 200- and 1000-m-depth contours. Storm motion is indicated by arrows in (a) for Isidore and (d) for Lili, respectively. Coordinate system (cross-track: xR_max; along-track: yL-1) dimensions are relative to the mean storm locations from in-storm flights (for which the R_max and L values are given in Table 2).
served along these bottom topographical gradients, with the largest SST changes of 4.5°C occurring over the Yucatan Shelf (Fig. 13c). That is, SSTs decreased to less than 25°C, because along-shelf wind stress driving a net surface offshore flow resulted in upwelling of a shallow seasonal thermocline. Although significant SST cooling and upward isotherm displacements occurred over the shelf just prior to landfall, only small thermal structure and isotherm depth changes were observed across the Straits to the western tip of Cuba. Thus, SSTs remained above 28.5°C in the straits a day after Isidore, suggestive of less negative feedback to the storm. Given the 10-day interval between Isidore and Lili, pre-Lili SSTs were 29°C to 29.5°C over most of the experimental domain. After Lili’s passage, SSTs decreased to only 28.5°C in the LC; however, along the northwest part of the measurement domain, SSTs cooled to less than 27°C in the GCW. Consistent with the National Data Buoy Center (NDBC) 42001 measurements (see Fig. 1), this observed SST change equated to more than 2°C cooling (Fig. 13f) as Lili intensified to a Category 4 storm in the south-central GOM (Pasch et al. 2004).

b. 26°C isotherm depths

Prior to Isidore, a strong horizontal gradient of the 26°C isotherm depth was observed, such that depths were found to be ranging from more than 150 m in the Yucatan Straits to less than 30 m over the Yucatan Shelf (Fig. 14). These horizontal differences were constrained by strong cross-stream topographical gradients separating the Yucatan Straits from the Yucatan Shelf. Consistent with these large SST changes over the shelf, isotherm depths decreased along these bottom topographical gradients after Isidore. However, in the center and eastern part of the deeper Yucatan Straits, isotherm depth increased by 20 m (Fig. 14c), which might be a manifestation of the downwelling cycle. This alternating upwelling cycle along the steep bottom slope and over the shelf and downwelling cycle in the straits is an integral part of the oceanic response to hurricane forcing (Geisler 1970). These processes result in a tightening of the isotherm depths (and hence the thermocline) and their gradients across the abrupt topographical changes.

The corresponding 26°C isotherm depths for pre-Lili conditions were located at more than 150-m depth in the southeast part of the LC (Fig. 14d) and decreased to 50 m along the northwest periphery in the GCW. Compared to a monthly climatology (Teague et al. 1990), these isotherm depths seem to be anomalously deep, but they were consistent with satellite-derived isotherm depths derived from satellite altimetry based on a seasonal climatology (Halliwell et al. 2008). Given Lili’s rapid translation speed, poststorm isotherm depths changed little compared to the pre-Lili values in the LC (Fig. 14e); that is, isotherm depths ranged between 90–140 m in the LC compared to about 100–150 m before the storm. The isotherm displacements induced by Lili were about 10 m, which is consistent with scaling argu-
ments using both the maximum wind stress and translation speed values in Table 2.

c. OHC variability

Pre- and poststorm OHC distributions (Fig. 15) reflect these SSTs and 26°C isotherm depths. Pre-Isidore OHC in the northwest Caribbean basin and through the eastern part of the Yucatan Straits exceeded 160 kJ cm\(^{-2}\), in accord with satellite-derived OHC values derived from radar altimetry (Halliwell et al. 2008). Over the Yucatan Shelf, pre-Isidore OHC values were about 40 kJ cm\(^{-2}\) (Fig. 15a), suggestive of a shallow seasonal thermocline maintained by the trade winds (Gill 1982). In the post-Isidore distributions, the OHC values were less than those observed during prestorm conditions, although by less than 20 kJ cm\(^{-2}\) along the western parts of the straits; along the eastern part, the OHC increased by about 20 kJ cm\(^{-2}\), consistent with a downwelling signal (i.e., deepening of the 26°C isotherm; Gill 1982). In the post-Isidore distributions, the OHC values were less than those observed during prestorm conditions, although by less than 20 kJ cm\(^{-2}\) along the western parts of the straits; along the eastern part, the OHC increased by about 20 kJ cm\(^{-2}\), consistent with a downwelling signal (i.e., deepening of the 26°C isotherm; Gill 1982). Over the shelf, however, the OHC losses were more than 40 kJ cm\(^{-2}\) where SSTs cooled by 4.5°C. In the center of the straits, there was essentially no OHC change, which was presumably due to northward heat transport by the LC. Evidently, these large spatial gradients in SSTs and OHC across the Yucatan Straits impacted the enthalpy fluxes that affected Isidore’s intensification to Category 3 status. Because isotherm depths decreased by about 20 m in the LC with a 1°C SST cooling, the LC essentially maintained OHC levels of more than 100 kJ cm\(^{-2}\) prior to Lili (Fig. 15d). Maximum OHC levels from profiler measurements exceeded 140 kJ cm\(^{-2}\) compared to satellite-inferred values of ~130 kJ cm\(^{-2}\) (not shown). By contrast, in the GCW, OHC values after Lili decreased by more than 35 kJ cm\(^{-2}\), compared to less than 15 kJ cm\(^{-2}\) in the LC regime. This larger OHC loss, associated with an SST decrease of more than 2°C, may have been caused by enhanced vertical shears because the surface enthalpy fluxes only accounted for about 4–5 kJ cm\(^{-2}\) of heat loss through the surface. Given that the post-Lili survey was conducted 1.5 IPs (48 h) after passage (Fig. 15e), the OHC loss may have been greater than this value because a major contributor to the oceanic heat budget is associated with the northward advection of heat by the LC. Differencing pre- and post-Lili OHC fields indicates that average fluxes were significantly less than 17 kJ cm\(^{-2}\) d\(^{-1}\) (~2.0 kW m\(^{-2}\)) because of the hurricane passage. Along the northwest part of the Lili domain (i.e., the GCW), thin OMLs cool and deepen quickly during hurricane passage where the SST decreases typically range from 3 to 6°C (Price 1981; Shay et al. 1992; Jacob et al. 2000) and induce a negative feedback (Chang and Anthes 1978; Bender and Ginis 2000). Price et al. (1994) argued that the North Atlantic subtropical front over which Hurricanes Josephine (1984) and Gloria (1985) passed is inconsequential to the simulated ocean response. In the three-dimensional LC regime, however, the horizontal pressure gradients
and balanced currents are considerably stronger than in the North Atlantic Ocean subtropical front. Thus, to simulate the response, ocean models must be initialized with these basic state flows (Marks and Shay 1998; Halliwell et al. 2008).

d. Vertical current shear

The most effective process for OML cooling and decreasing SST is by entrainment mixing (downward heat flux) across the base of the OML associated with vigorous near-inertial current shears (e.g., Price 1981; Schade and Emanuel 1999). This process is parameterized in numerical models and has been shown to produce widely varying results depending on the chosen mixing parameterization (Jacob et al. 2000; Jacob and Shay 2003). The results presented here suggest that strong prestorm current regimes may limit the development of wind-forced near-inertial currents and their vertical shears.

Current profiles are used to assess the vertical shear using profiles with a vertical resolution of 2 m after removal of the orbital velocities (see section 4a). Based on (4), current shears at the OML base are estimated from the model-fitted shear coefficients in layer 2 ($S_2$) from Tables 3, 4, and 5. The means and standard deviations of $S_2$ are compared to Norbert and Gilbert shear measurements in Table 7. Within measurement error (Gregg et al. 1986), weaker shears were observed in the current profiles acquired during two severe hurricanes. These estimated shear values in layer 2 are objectively analyzed for both in- and poststorm fields only (Fig. 16). Evident in these fields is the weaker shear observed in the LC regime compared to measured shears in the GCW. Within the LC, for example, current shears ranged from $0.5–1.5 \times 10^{-2}$ s$^{-1}$, but outside the LC, the shears were $2.0–5.0 \times 10^{-2}$ s$^{-1}$, or 2 to 4 times larger. This effect is obvious in the in-storm Isidore and poststorm Lili current fields (Figs. 16a,d). The lagged current shear response in Lili may also result from rapid storm motion and its asymmetric wind stress distribution compared to the symmetric and slower-moving Isidore. Given these wide-ranging results between two distinct water masses, considerably more current and shear measurements are needed to fully assess model parameterizations.

e. Bulk Richardson number

The bulk Richardson number is estimated from the following expression:

$$Ri_b = \frac{g \Delta T}{\delta V^2},$$  \hspace{1cm} (9)

where $\alpha$ is the coefficient of thermal expansion ($2 \times 10^{-4}$ °C$^{-1}$; Kraus and Businger 1994), $h$ is the OML depth, $\Delta T$ is the temperature difference between the bulk OML and averaged temperature in layer 2, and the magnitude of the bulk current shears ($|\delta V|$) are determined from the model-fitted mean current differences between layers 1 and 2 as per Tables 3, 4, and 5. Pollard et al. (1973) used a value of unity for the bulk Richardson number as a condition for the onset of mixing processes at the OML base, whereas Ellison and Turner (1959) found that more appropriate values ranged between 0.4 and 0.8 for the initiation of vertical mixing. Price (1981) used a $Ri_b$ of 0.8 as the critical value for mean current shear-induced mixing based on a scaled entrainment law from experimental laboratory results.

Small values of $Ri_b$ indicate regions where shear-induced mixing is likely to overwhelm the damping effect of stratification during the cooling process (Fig. 17). However, in-storm measurements in Isidore indicate $Ri_b > 1$ over most of the domain. These values in the LC are consistent with the observed 1°C cooling and the OHC change of 20 kJ cm$^{-2}$. Even where the maximum SST cooling of 4.5°C was observed, bulk Richardson numbers were above critical values (Figs. 17a,b) suggesting that upwelling of colder thermocline water (induced by the wind stress curl) was the dominant mechanism over the shelf. The results for Lili are similar (Figs. 17c,d), although they are more evident in the poststorm analysis of current shear measurements. Both of these cases point to a reduction in shear-induced mixing events in regions of strong background currents; that is, the presence of deep warm layers coupled with strong background flows of 1 m s$^{-1}$ (caused by horizontal density and pressure gradients) precludes the generation of strong shears observed in the near-inertial wave wake (Shay et al. 1998). These results point to a physical mechanism that must be well understood in coupled models that predict hurricane intensity.

6. Summary and concluding remarks

The goal of the aircraft-based experiment was to measure the three-dimensional current, temperature,
and salinity responses excited by hurricane passage with a focus on assessing the responses in and over the LC. The aircraft-based sampling strategy resulted in several snapshots of upper-ocean current, temperature, and salinity structure, required to examine the response to the passage of two Category 3 hurricanes moving over the same domain over a 10-day period. Given the inherent uncertainties of storm-track prediction for the pre-Isidore flights, this experimental objective was achieved with a high degree of success for Isidore and Lili. This is one of only a few datasets in which currents and shears were directly measured during hurricane passages (Sanford et al. 1987; Shay et al. 1992; Price et al. 1994; Sanford et al. 2005). In this case, the GPS sondes and SFMR data complement these in situ ocean measurements to improve our understanding of their mutual response.

The ocean response to the passage of Hurricanes Isidore and Lili was investigated using in situ observations from oceanic and atmospheric sondes and remotely sensed ocean surface winds (Uhlhorn et al. 2007). These hurricanes intensified to major hurricane status (Category 3 and above) over the LC, and Lili’s deepening cycle continued into the central GOM basin as the storm reached Category 4 status just northwest of the LC boundary. Atmospheric conditions were conducive for deepening, as noted by Pasch et al. (2004), over an ocean where the extent of the cooling was $O(1^\circ C)$ and the upper-ocean heat loss was less than $20 \text{ kJ cm}^{-2}$. Even after both hurricanes, the upper ocean was still

![Image of current shears](image-url)
warm, with SSTs of 28.5° to 29°C and OHC levels exceeding 100 kJ cm⁻². These results are similar to observations acquired prior and subsequent to Hurricane Rita, which followed a track close to Katrina over the LC and WCR (Shay 2008). Based on these profiler data, lessons learned from Isidore and Lili included the following:

1) the three-dimensional LC precludes the development of strong vertical current shears to force mixing and deepening of the OML despite applied wind stresses of 7 N m⁻² due to the strength of the current and the depth of the warm isotherms;

2) cooling of 4.5°C over the Yucatan Shelf to Isidore was due to wind-forced upwelling, and more than 2°C cooling during Lili was due to shear-induced mixing events in the GCW; and,

3) maximum surface heat loss from the ocean was less than 10 kJ cm⁻² at $R_{\text{max}}$, where enthalpy fluxes ranged from 1.4 to 1.8 kW m⁻² for Lili and Isidore, respectively.

The first point is important with regard to initializing ocean models with realistic background conditions (Halliwell et al. 2008; Mainelli et al. 2008). Emanuel (2001) argues that the upper ocean can be treated as a series of one-dimensional column models for coupled hurricane forecasting. The results reported, and other recent studies such as Wu et al. (2007), imply that three-dimensional advective tendencies must be accounted for in such models. Because the ocean response is weakest in strong frontal regimes, the negative feedback to the atmosphere is much less than over quiescent ocean regimes. The third conclusion has direct rel-

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**FIG. 17.** Same as Fig. 16, but for $R_i$.
evidence to the threshold value for sustaining a hurricane's intensity. Leipper and Volgenau (1972) suggested a value of 17 kJ cm$^{-2}$ day$^{-1}$, but this threshold must be revisited to understand how much oceanic heat loss is related to surface enthalpy fluxes versus entrainment heat loss to the thermocline. In this context, Cione and Uhlhorn (2003) argue that only inner-core SSTs are necessary for intensity forecasting, but their results are inconclusive because the OHC held constant even though SSTs were changing. With temperature profiles, OHC changes relative to the 26°C isotherm can be estimated at finer scales than currently available from coarse altimeter tracks (Cheney et al. 1994).

The scientific issue is not just the OHC magnitude; the depth of the 26°C isotherms in the LC and WCR regime are of equal importance in determining the value of the OHC as per Eq. (1). The deeper this layer of warm water, the more turbulent mixing is required to overturn and cool the OML. Significant internal wave shears associated with near-inertial motions were not observed in the LC during Isidore and Lili because of the strength of the horizontal pressure gradients that force the background currents. If shear-induced mixing is arrested, significant OML cooling and deepening will not occur. In contrast, when current shears are large, they lower the bulk Richardson numbers to below critical values and cool the SST (a proxy for OML temperatures under high winds) as the OML deepens (Price 1983; Shay et al. 1992). Entrainment heat fluxes into the thermocline will result in lower surface enthalpy fluxes that feed the hurricane and impact hurricane intensity (Chang and Anthes 1978; Schade and Emanuel 1999, Bender and Ginis 2000). For coupled model forecasting, currents and shears (momentum responses) are as important as thermal profiles for simulating OML cooling and deepening processes. Given the number of vertical mixing models using bulk and turbulence closure schemes (Jacob and Shay 2003), exploring parameter space under differing ocean conditions (water masses) will require more current (and shear) measurements than have been previously acquired under hurricane conditions. Once obtained, these could be integrated into the NOAA Intensity Forecasting Experiments (Rogers et al. 2006). The ocean’s momentum response to hurricane forcing has been used to determine the behavior of the surface drag coefficient (Shay and Jacob 2006; Jarosz et al. 2007). These data are needed to test and evaluate innovative forecasting schemes involving both air–sea and vertical mixing parameterizations. This is a crucial test for coupled models designed for operational hurricane intensity forecasts (Marks and Shay 1998).

Acknowledgments. LKS acknowledges the support of the National Science Foundation for both the experiment and data analysis through Grants ATM-01-08218 and ATM 04-44525; flight hours were provided by NOAA’s Hurricane Research Division (HRD) and the National Hurricane Center. E. Uhlhorn was supported by both NOAA and ONR-CBLAST under the leadership of Dr. Peter Black. We gratefully acknowledge Dr. Jim McFadden of the Office of Aircraft Operations (OAO) and Drs. Hugh Willoughby and Frank Marks (NOAA HRD) in orchestrating these aircraft experiments. Dr. Paul Chang of NOAA/NESDIS directed the Ocean Winds flights on N42RF and provided support for both the Isidore and Lili flights. Mr. Michael Black (HRD field director) directed all field support for these flights. We appreciate the extraordinary efforts of the OAO pilots, engineers, and technicians during these experimental efforts. Tom Cook and Scott Guhin assisted in the aircraft-based experiments, and Jodi Brewster provided graphical support. Reviews from three anonymous reviewers significantly improved the manuscript.

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