Enthalpy and Momentum Fluxes during Hurricane Earl Relative to Underlying Ocean Features

BENJAMIN JAIMES AND LYNN K. SHAY

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

ERIC W. UHLHORN

Hurricane Research Division, NOAA/Atlantic Oceanographic and Meteorological Laboratory, Miami, Florida

(Manuscript received 30 August 2013, in final form 6 August 2014)

ABSTRACT

Using dropsondes from 27 aircraft flights, in situ observations, and satellite data acquired during Tropical Cyclone Earl (category 4 hurricane), bulk air–sea fluxes of enthalpy and momentum are investigated in relation to intensity change and underlying upper-ocean thermal structure. During Earl’s rapid intensification (RI) period, ocean heat content (OHC) variability relative to the 26°C isotherm exceeded 90 kJ cm⁻², and sea surface cooling was less than 0.5°C. Enthalpy fluxes of ~1.1 kW m⁻² were estimated for Earl’s peak intensity. Daily sea surface heat losses of 2.6 ± 0.7, 2.7 ± 0.8, and 1.2 ± 0.3 kJ cm⁻² were estimated for RI, mature, and weakening stages, respectively. A ratio $C_H/C_D$ of the exchange coefficients of enthalpy ($C_H$) and momentum ($C_D$) between 0.54 and 0.7 produced reliable estimates for the fluxes relative to OHC changes, even during RI; a ratio $C_H/C_D = 1$ overestimated the fluxes.

The most important result is that bulk enthalpy fluxes were controlled by the thermodynamic disequilibrium between the sea surface and the near-surface air, independently of wind speed. This disequilibrium was strongly influenced by underlying warm oceanic features; localized maxima in enthalpy fluxes developed over tight horizontal gradients of moisture disequilibrium over these eddy features. These regions of local buoyant forcing preferentially developed during RI. The overall magnitude of the moisture disequilibrium ($\Delta q = q_s - q_a$) was determined by the saturation specific humidity at sea surface temperature ($q_s$) rather than by the specific humidity of the atmospheric environment ($q_a$). These results support the hypothesis that intense local buoyant forcing by the ocean could be an important intensification mechanism in tropical cyclones over warm oceanic features.

1. Introduction

Simulation of tropical cyclone (TC) intensity and structure change with coupled models promises to be a key forecast tool to support neighborhood-level, probabilistic warnings of damaging weather elements. This level of precision is essential to efficiently prepare everyone from the general public to government agencies or corporate managers for landfalling hurricanes (Marks and Shay 1998). Over the past decade, it has become increasingly clear that the oceanic component of these models needs to include realistic initial conditions to correctly simulate both oceanic responses to hurricane forcing (e.g., Shay 2010; Halliwell et al. 2011; Jaimes et al. 2011) and atmospheric responses to oceanic forcing (e.g., Emanuel 1999; Schade and Emanuel 1999; Hong et al. 2000). The main reason is because upper-ocean warm thermal structures, the energy source for TCs, are seldom uniform (Jaimes et al. 2011), and TCs often experience rapid intensification (RI) over warm oceanic mesoscale features (e.g., Emanuel 1999; Shay et al. 2000; Lin et al. 2005, 2009; Wada and Chan 2008; Mainelli et al. 2008; Jaimes and Shay 2009, 2010) and decay over cold waters where the energy source vanishes.
Thus, RI in low-shear environments over warm, deep oceanic mixed layers, and weakening before landfall, defines a vital intensity forecast challenge.

In the classical paradigm (Emanuel 1986), intensity is a function of the balance among the surface moist enthalpy flux, energy lost to frictional dissipation, and heat exhausted to the surroundings in the upper troposphere. Within this framework, this balance is the primary determinant of maximum potential intensity, and the mutual dependence between the air–sea heat and momentum transfers is critical. Intensity is thus sensitive to the ratio of enthalpy to drag coefficients \( (C_K/C_D) \), which must lie within a limited range (Emanuel 1995). Within this context, the intensification pathway invokes two key elements: 1) a positive “evaporation–wind” feedback mechanism known as wind-induced surface heat exchange, which depends on wind speed, and 2) a “transfer of heat above and beyond that associated with isothermal expansion,” which is supported by sea-to-air heat fluxes. In an emerging intensification paradigm (Montgomery et al. 2002, 2009; Van Sang et al. 2008), intensity and structural change occur through stochastic deep convective vortex structures that are highly sensitive to the boundary layer moisture distribution, as they obtain their local buoyancy from sea-to-air fluxes of moisture. Under this new paradigm, intensification develops even under low-wind conditions, that is, independent of wind speed.

The key role of the transfer of heat from the ocean to the atmosphere—the most fundamental process of a TC (Malkus and Riehl 1960)—in the intensification paradigms emphasizes the need for better understanding of the air–sea interface and physical interactions between the two systems. Limited observations of this complex air–sea interaction under high-wind conditions have prevented significant improvements in the parameterization of momentum and heat transfers, defining robust limits for the ratio \( C_K/C_D \), and elucidating RI events. Better understanding of these key aspects is critical to improving intensity forecasts. This problem is particularly relevant during RI over warm oceanic structures.

One important caveat of the foregoing intensification paradigms is that they are based on the assumption of constant and homogeneous values of sea surface temperature (SST). Whereas this assumption simplifies theoretical developments, and facilitates our understanding of the intensity and structure change in controlled numerical experiments, it prevents evaluating the role of upper-ocean thermal structure variability on sea-to-air heat fluxes and intensity change. This upper-ocean variability has been observed to significantly enhance sea-to-air enthalpy fluxes such that the ocean forces the atmosphere over mesoscale regimes (e.g., Small et al. 2008, and references therein).

Observations obtained during the life cycle of TC Earl (a rapid intensifier to a category 4 hurricane), as part of the NOAA Intensity Forecasting Experiment (IFEX) (Rogers et al. 2013), from the National Aeronautics and Space Administration (NASA) Genesis and Rapid Intensification Processes (GRIP) experiment (Braun et al. 2013), from the National Science Foundation’s (NSF) Pre-Depression Investigation of Cloud-Systems in the Tropics (PREDICT) experiment (Montgomery et al. 2012), and from U.S. Air Force Reserve (USAFR) reconnaissance flights, provide a valuable opportunity to learn more about intensity change in relation to the underlying ocean and associated sea-to-air enthalpy fluxes. Within this context, data acquired during these experiments along with ocean observations and satellite data (section 2) are used in this research to describe Earl’s evolution in relation to underlying upper-ocean thermal structures (section 3), and to delineate the variability in the air–sea fluxes of enthalpy and momentum along the storm’s track based on the “bulk formulae” (section 4). These estimated enthalpy fluxes are compared in relation to changes in ocean heat content (OHC) relative to the 26°C isotherm derived from satellite products (section 5). The impact of the ratio of enthalpy to momentum exchange coefficients on the estimates for the bulk heat fluxes reported here is discussed in relation to the observed OHC response (section 6). Finally, concluding remarks are presented in section 7. The goal is to show how the spatial variability in oceanic warm and cold thermal structures (including the SST) encountered by Earl impacts the air–sea fluxes of enthalpy and momentum and the ensuing intensity fluctuations.

2. Data and methodology

a. Atmospheric data

Heat and moisture fluxes are central to TC intensity (e.g., Malkus and Riehl 1960; Emanuel 1986). In this research, atmospheric data from global positioning system (GPS) dropwindsondes (Hock and Franklin 1999; hereafter referred as dropsondes) deployed in Earl during 27 research and reconnaissance flights (Table 1) are combined with in situ SST data (section 2b) to estimate momentum, heat, and moisture surface fluxes based on “bulk formulae”:

\[
\tau = \rho_A C_D |U_{10}|^2, \quad \tag{1a}
\]

\[
Q_s = \rho_s c_p C_h |U_{10}| (\text{SST} - T_a), \quad \tag{1b}
\]

\[
Q_l = \rho_a L_{vap} C_q |U_{10}| (q_s - q_a), \quad \tag{1c}
\]
where $\tau$, $Q_s$, and $Q_l$ are the momentum, sensible, and latent heat surface fluxes, respectively; $\rho_a$ is the atmospheric density; $C_p$, $C_p$, and $C_q$ are exchange coefficients of momentum, sensible, and latent heat, respectively; $|U_{10}|$ is the 10-m wind speed ($U_{10}$ hereafter); $c_p = 1004 \text{ J kg}^{-1} \text{ K}^{-1}$ is the specific heat of air at constant pressure; $L_{\text{vap}} = 2.5 \times 10^6 \text{ J kg}^{-1}$ is the latent heat of vaporization; $T_S$ is the 10-m air temperature; $q_s$ is the saturated specific humidity at the SST [assumed to be at 98% saturation at the SST; Buck (1981)]; and $q_a$ is the 10-m atmospheric specific humidity. As will be described in section 2d, $U_{10}$, $T_p$, and $q_a$ are estimated from dropsonde data. Enthalpy fluxes are given by $Q_h = Q_s + Q_l$, and moisture and thermal differences between the ocean and atmosphere are $\Delta q = q_s - q_a$ and $\Delta T = SST - T_a$.

Data from WP-3D, Gulfstream-IV (G-IV), and WC-130J aircraft were postprocessed using the National Center for Atmospheric Research (NCAR) Atmospheric Sounding Processing Environment (ASPEN) software. GRIP NASA DC-8 quality controlled dropsonde data were provided by the NCAR Earth Observing Laboratory under sponsorship of NSF (http://data.eol.ucar.edu/cddsa/id=126.016). Dropsonde instrumentation and data accuracies are described in Hock and Franklin (1999). A total of 557 quality-controlled dropsonde profiles were obtained from the 27 flights listed in Table 1. A list of the relevant parameters (and their units) to be discussed in this study is presented in Table 2.

b. SST data

Using reliable SST data is critical to obtaining good estimates for the enthalpy fluxes based on bulk formulas (1b) and (1c). Near-real-time, in situ SST observations from the U.S. Global Ocean Data Assimilation Experiment (USGODAE) (http://www.usgodae.org/) were used in this research. This dataset includes observations in TC Earl from fixed surface weather buoys and surface drifting buoys. Several of these observations were obtained under the storm’s inner-core region (Fig. 1). For each one of the 27 flights used in this study, in situ SSTs were retrieved from the USGODAE dataset for the 2 h prior to launching the first dropsonde in that particular flight (from $t = -2$ to 0 h; the latter time is the launching time of the first dropsonde), and for the geographic area covered by the flight. Because these scattered SST data points were not spatially collocated with dropsonde points, they were objectively analyzed (OA) for the geographic area of the flight [all OA experiments conducted in this study are based on Mariano and Brown (1992)]. These OA in situ SST structures were used to retrieve SST data at dropsonde points; the latter data are denoted as SST$_{in0}$ and they are collocated in space and time (within a 2-h time window) with dropsonde points. Because in situ SST$_{in0}$ data are from in-storm conditions, they were used in combination with the dropsonde observations in estimating bulk heat fluxes with (1b) and (1c). A second OA in situ SST structure was obtained for each flight, for the same geographic area as SST$_{in0}$ but from $t = 10$ to 12 h following the launching time of the first dropsonde in the flight. This second group of in-storm SSTs is called SST$_{in12}$. The in-storm cooling response to TC Earl is computed as SST$_{in12}$ minus SST$_{in0}$ (section 3b).

To evaluate the effect of the mapping error associated with the OA SSTs, retrievals from the OA SST structures at the 864 available in situ data points (Fig. 1) were compared with the actual data. The overall temperature bias between actual and OA SSTs is $-0.03^\circ \text{C}$ with a standard deviation (STD) of 0.29$^\circ \text{C}$, and the correlation coefficient is 0.99 (Fig. 2). Thus, the mapping error associated with the OA structure is negligible.

c. OHC data

The SST response to TC wind forcing is a key process for storm intensity because it directly affects the most

---

**Table 1. Number of research and reconnaissance flights in TC Earl considered in this study. Only flights where GPS dropsondes were deployed in storm conditions are considered (one NOAA flight conducted in the wake of the storm, and six USAFR reconnaissance flights with no dropsonde data, are not listed). The first four characters in the flight ID are for MMDD, where MM is month and DD is day.**

<table>
<thead>
<tr>
<th>Start date</th>
<th>NOAA WP-3D</th>
<th>NOAA G-IV</th>
<th>USAFR WC-130J</th>
<th>NASA DC-8</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>28 Aug</td>
<td>0828H1</td>
<td>0828N2</td>
<td>0829A1</td>
<td>0829D1</td>
<td>2</td>
</tr>
<tr>
<td>29 Aug</td>
<td>0829H1</td>
<td>0829N1</td>
<td>0830A1</td>
<td>0830D1</td>
<td>4</td>
</tr>
<tr>
<td>30 Aug</td>
<td>0830H1</td>
<td>0830N1</td>
<td>0830A1</td>
<td>0830D1</td>
<td>5</td>
</tr>
<tr>
<td>31 Aug</td>
<td>0831N1</td>
<td>0831A1</td>
<td>0901A1</td>
<td>0901D1</td>
<td>2</td>
</tr>
<tr>
<td>1 Sep</td>
<td>0901H1</td>
<td>0901N1</td>
<td>0901A1</td>
<td>0901D1</td>
<td>5</td>
</tr>
<tr>
<td>2 Sep</td>
<td>0902H1</td>
<td>0902N1</td>
<td>0902A1</td>
<td>0902D1</td>
<td>5</td>
</tr>
<tr>
<td>3 Sep</td>
<td>0903H1</td>
<td>0903N1</td>
<td>0903A1</td>
<td>0903D1</td>
<td>4</td>
</tr>
<tr>
<td>Total</td>
<td>10</td>
<td>7</td>
<td>6</td>
<td>4</td>
<td>27</td>
</tr>
</tbody>
</table>
fundamental process of a TC, the sea-to-air transfer of heat (Malkus and Riehl 1960). This SST response is essentially a problem of internal ocean dynamics where the temperature structure over the oceanic mixed layer (OML) and thermocline plays a critical role (e.g., Price 1981; Shay et al. 1992, 1998, 2000; Hong et al. 2000; Jacob et al. 2000; Jaimes et al. 2011). A useful parameter for estimating the upper-ocean thermal energy available for TCs is the hurricane heat potential (Leipper and Volgenau 1972), or ocean heat content (OHC) relative to the 26°C isotherm depth (h26):

\[
\text{OHC} = \rho_1 c_p \int_{z=h_{26}}^{z=\eta} [T(z) - 26] \, dz, \tag{2}
\]

where \( \rho_1 = 1026 \text{ kg m}^{-3} \) is the reference seawater density, \( c_p \) is the specific heat at constant pressure (4.2 kJ kg\(^{-1}\) K\(^{-1}\)), \( T(z) \) is the upper-ocean temperature structure that includes SST, and \( \eta \) is the sea surface height (in meters). Note that we use the 26°C isotherm here since it is the temperature assumed for tropical cyclogenesis (Palmen 1948). Shay and Brewer (2010) describe in detail the computation of OHC from space-based measurements [sea surface height anomaly and Tropical Rainfall Measuring Mission (TRMM) Microwave Imager (TMI) SST], in combination with climatological values of the OML depth as well as the 20° and 26°C isotherm depths. These climatological values are from the Systematically Merged Atlantic Regional Temperature and Salinity (SMARTS) climatology (Meyers 2011; Meyers et al. 2014).

SMARTS was developed to produce satellite-based daily images of OHC for the North Atlantic Ocean basin. Over a 10-yr period, altimeter data were used to obtain daily values based on a centered average that considers information from day -5 to day +5 when at least three satellite altimeters were available. These satellite-derived values were compared to more than 60,000 thermal profiles from XBT transects, Argo profiling floats, Prediction and Research Moored Array in the Tropical Atlantic (PIRATA) moorings, and aircraft expendables including those from the Deepwater Horizon flights (Shay et al. 2011; Meyers et al. 2014). SMARTS is used operationally at the National Hurricane Center (NHC) and the National Environmental Satellite, Data, and Information Service (NESDIS) (Shay et al. 2012). A similar effort is ongoing in the North Pacific except the database has more than 267,000 in situ thermal profiles.

A different approach to represent the upper-ocean thermal energy available for TCs considers sea temperature vertically averaged over the upper 80 and 100 m, \( T_{80} \) and \( T_{100} \), respectively (Price 2009; Lin et al. 2013; Pun et al. 2014). While \( T_{80} \) and \( T_{100} \) look promising for estimating the potential intensity index for TCs, we use the 26°C isotherm here since it is the temperature assumed for tropical cyclogenesis. Therefore, in this research SMARTS OHC is used to characterize the thermal energy available for TC Earl. All available in situ measurements that provided subsurface temperature structure for the time and geographic region of TC Earl are considered in SMARTS (Meyers et al. 2014), including the USGODAE SST data used in estimating SST\(_{in0} \) and SST\(_{in12} \). Thus, SMARTS OHC is consistent with these SST products.

Daily SMARTS OHC structures were used to retrieve OHC values individually for each flight, for the day and dropsonde points related to the flight. These in-storm OHC structures are named OHC\(_{in} \). For each flight, a second set of OHC data was retrieved at the same data points as OHC\(_{in} \) (i.e., dropsonde points), but for the day before the flight; these data are named OHC\(_{pre} \). Evaluated pre- and in-storm OHC fields are used here to describe Earl’s intensity change in relation to the spatial variability in the upper-ocean thermal structure (section 3), and to evaluate the sea surface heat loss [or enthalpy fluxes estimated with (1)] compared with changes in OHC (sections 5 and 6).

### Cluster analysis

The composite analysis method of dropsonde data has been used to investigate several characteristics...
of TCs, including inner-core structure (Rogers et al. 2012), wind speed vertical profile structure (Powell et al. 2003), and sea surface inflow angle (Zhang and Uhlhorn 2012). By combining data from several similar storms into a single realization, this method increases the statistical significance of the analysis, helps fill data gaps, and provides a general characterization of TC fields. This method, however, tends to smooth characteristics of individual TCs from a large number of storms that may not necessarily be similar (Zhang and Uhlhorn 2012), and the assumption of a steady-state inherent to this method prevents investigation of the time-dependent nature of TCs.

Since the goal of this research is examining the variability in air–sea fluxes and storm intensity in Earl in relation to the along-track variability in upper-ocean thermal structure, the present treatment defines a modified version of the composite analysis method to produce several realizations of the near-surface storm structure over the storm track. The goal is to define dropsonde clusters within a constrained window in time, space, and similarity in storm strength or intensity tendency (as feasible). Each cluster must resolve the storm’s four quadrants and provide a good depiction of its radial structure, including the inner-core region. Within this context, dropsonde data from the 27 flights were organized into seven clusters along the storm track, where data outside a radius of 12 times the radius of maximum winds ($R_{\text{max}}$) reported in NHC best track files are ignored in the definition of the cluster (Fig. 3 and Table 3). Concurrence of the three field experiments and USAFR reconnaissance flights facilitates this approach, as most clusters consider data from more than three flights. Whereas this approach addresses the time-dependent nature of the problem, it reduces the statistical significance of analyses, particularly in clusters 1 and 4, because of undersampling.

Four of the clusters consider a single storm category (Figs. 4a,b): C1 (TS), C3 (H4), C4 (H3), and C5 (H4); however, there is intensity fluctuation in C4 and C5. The other three clusters consider two categories and a single intensity trend: C2 (H1 to H2), C6 (H3 to H2), and C7 (H1 to TS). Partitioning the latter three clusters (and
C4 and C5) into smaller clusters for individual storm category (or single intensity trend) is not convenient because an incomplete coverage of the horizontal structure of the storm is obtained. One of the objectives of individual in-storm flights is to sample the four quadrants of the storm (the total observational period is typically 8 h). Rapidly intensifying or weakening storms often change intensity from one category to another during a single flight from which dropsondes are launched. Under these circumstances, assigning a single storm category to dropsonde clusters is difficult. To address this issue, the dominant intensity tendency is used in describing intensity changes in the clusters (Table 4).

Note that the dominant intensity tendency in C2 satisfies the criterion of RI (Fig. 4c), conventionally defined as a 30 kt (15.43 m s\(^{-2}\)) intensification in 24 h (Kaplan and DeMaria 2003). Assuming there is a symmetric counterpart for rapid weakening (RW), C6 satisfies this criterion and C7 nearly satisfies it. While Earl was in a nearly steady state during most of C3 (Fig. 4a), the slightly positive intensity trend in C5 results because 78% of the dropsondes were launched during the intensification phase in this cluster (first 15 h of the cluster interval). Some dropsonde points along the storm track in Fig. 3 appear to be associated with a different storm category than that actually related to the cluster (Fig. 4b) because they seem to be intruding on the geographic area of the contiguous clusters. This is a visual artifact because some dropsondes in the cluster were launched in outer storm’s regions, that is, over regions ahead (future storm’s center position) or behind (past storm’s center position) the current storm’s center that was approximately located at the cluster’s core. Deployment time (Fig. 4b), rather than geographic position in relation to the storm track, facilitates relating dropsondes to the actual category the storm had on the Saffir–Simpson scale when they were launched. Thus, Fig. 4b should be used in identifying the storm category in the clusters, rather than color of underlying storm track segments.

For each data cluster, dropsonde geographic points are referenced in a storm coordinate system (cylindrical coordinates) that removes storm motion, where \(r\) is the radial distance from the storm center and \(\lambda\) is the azimuthal angle based on the best track of the storm reported by NHC; \(r\) is normalized by \(R_{\text{max}}\). Binned azimuthal means (averaged from \(\lambda = 0^\circ\) to \(360^\circ\)) of observed 10-m values of wind speed, air temperature, and relative humidity are computed from the dropsonde data, where the bin size in the radial direction is \(R_{\text{max}}/4\). These mean profiles (a function of \(r\)) are used to get interpolated values, that is \(U_{10}(r)\), \(T_s(r)\), and \(q_a(r)\), at actual dropsonde points referenced in the storm coordinate system; these interpolated values

TABLE 3. Number of dropsondes and research flights per cluster. A total of 392 dropsondes that were deployed within a radius of \(12R_{\text{max}}\) from the storm’s center are considered in this analysis. Note that dropsondes from two flights (0901A1, 0902A1) are split into two contiguous clusters. The flights are ordered sequentially in accord the starting time of the sampling (drop time of the first dropsonde in the flight).

<table>
<thead>
<tr>
<th>Cluster</th>
<th>C1</th>
<th>C2</th>
<th>C3</th>
<th>C4</th>
<th>C5</th>
<th>C6</th>
<th>C7</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dropsodes</td>
<td>39</td>
<td>69</td>
<td>78</td>
<td>29</td>
<td>60</td>
<td>55</td>
<td>62</td>
</tr>
<tr>
<td>Flights</td>
<td>2</td>
<td>4</td>
<td>6</td>
<td>3</td>
<td>7</td>
<td>3</td>
<td>4</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Time</th>
<th>0828N2</th>
<th>0829D1</th>
<th>0830H1</th>
<th>0831N1</th>
<th>0901A1</th>
<th>0902A1</th>
<th>0903H1</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>0828H1</td>
<td>0829N1</td>
<td>0830D1</td>
<td>0901H1</td>
<td>0901D1</td>
<td>0902D1</td>
<td>0903A1</td>
</tr>
<tr>
<td></td>
<td>0829A1</td>
<td>0830A1</td>
<td>0901A1</td>
<td>0901N1</td>
<td>0902H1</td>
<td>0902H1</td>
<td>0903N1</td>
</tr>
<tr>
<td></td>
<td>0829I1</td>
<td>0830N1</td>
<td>0901I1</td>
<td>0901I1</td>
<td>0902H1</td>
<td>0902A1</td>
<td>0902A1</td>
</tr>
</tbody>
</table>

FIG. 3. As in Fig. 1, but for the along-track geographic distribution of dropsonde data clusters; labels C1–C7 identify the clusters. The 392 dropsonde points presented (out of 557 total data points, or 70.4%) are situated within a distance of \(12R_{\text{max}}\) from the storm center; this distance defines the cluster’s outer boundary. The numbers of dropsondes and flights considered in each cluster are presented in Table 3.
values are used in (1). In addition to $U_{10}$, $T_a$, and $q_a$, all other atmospheric and oceanic variables (SST and OHC) used in this research are referenced in the storm coordinate system. As described in sections 2b and 2c above, SST and OHC data are collocated with dropsonde points for each flight. The one-to-one relationship between geographic coordinates of the ocean data and dropsonde points is used to project the related SST and OHC data points into the storm coordinate system for each cluster.

Mean profiles of $U_{10}$ in general extend within the envelope of the overall STD for this parameter (Fig. 5). This is also the case for mean profiles of $T_a$ and $q_a$ (not shown). The overall mean profile (red line) substantially smooths $U_{10}$. Using this single mean profile for the entire along-track distance would impact the estimates for the fluxes (fluxes would be overestimated and underestimated for low and high wind regimes, respectively), thus preventing an investigation of the spatially evolving nature of the problem. This is precisely why data clusters over the storm’s track are used herein. Small-scales structures in Fig. 5 could be associated with secondary eyewall formation (outer region). These signals are slightly smoothed when data are referenced in regular grids via OA (section 4).

3. Hurricane Earl

a. Hurricane Earl and variability over the North Atlantic subtropical gyre

Earl developed from a strong tropical wave that originated off the west coast of Africa on 23 August.

\[
\begin{array}{cccccc}
\text{Cluster} & \text{C1} & \text{C2} & \text{C3} & \text{C4} & \text{C5} & \text{C6} & \text{C7} \\
\Delta T_c (h) & 8.0 & 17.5 & 25.8 & 15.7 & 22.6 & 13.9 & 13.5 \\
\Delta V_{max} (m \cdot s^{-1}) & 2.5 & 15.0 & 2.2 & -3.6 & 7.6 & -11.1 & -7.5 \\
\end{array}
\]

TABLE 4. Dominant intensity tendency ($\Delta V_{max}$) during the time frame ($\Delta T_c$) of the dropsonde clusters. The time interval between the drop time of the first and last dropsondes in the cluster is $\Delta T_c$; it is determined from information related to the actual drop time from dropsonde files. The difference between minimum and maximum values in $V_{max}$ during the time period of the clusters is defined as $\Delta V_{max}$; it is estimated from NHC 6-h best track data.
The system strengthened into a tropical storm by 1200 UTC 25 August, over waters to the west-southwest of the Cape Verde Islands. The tropical storm moved westward to west-northwestward over the southern branch of the North Atlantic subtropical gyre, in an environment of weak wind shear that facilitated its gradual intensification over SSTs of 28–29°C (Cangialosi 2011).

Earl continued moving over the southern branch of the North Atlantic subtropical gyre, in an environment of weak wind shear that facilitated its gradual intensification over SSTs of 28–29°C (Cangialosi 2011). Earl continued moving over the southern branch of the North Atlantic subtropical gyre where the oceanic thermal structure progressively deepened as the storm moved westward (OHC structure, Fig. 6b). In a persistent environment of weak wind shear over this region, Earl became a hurricane by 1200 UTC 29 August as it approached the Leeward Islands (Cangialosi 2011). Subsequently, Earl experienced rapid intensification becoming a category 4 hurricane by 1800 UTC 30 August over warm waters of the Antilles Current (local component of the North Atlantic subtropical gyre located north of the Caribbean Islands) where OHC levels were high (>90 kJ cm⁻², Fig. 6b) and SST values were relatively uniform at ~30°C (Fig. 6d).

Earl’s intensification was temporarily halted by increased wind shear and an eyewall replacement cycle just north of Hispaniola (Cangialosi 2011), where the storm weakened to H3 by 0000 UTC 1 September (Fig. 6b). Earl subsequently experienced a second intensification to H4 by 1800 UTC 1 September, attaining its maximum intensity of 64 m s⁻¹ by 0600 UTC 2 September over a region where OHC levels were above 50 kJ cm⁻². By 1200 UTC 2 September, Earl started to rapidly weaken during another concentric eyewall replacement cycle as it moved in a less favorable atmospheric environment [increased wind shear and dry air, Cangialosi (2011)] and

![Diagram](image-url)
over cooler shelf waters along the western flank of the Gulf Stream where OHC vanished (Fig. 6b).

b. In-storm cool wake

The in-storm SST response to TC Earl is presented in Fig. 6f. During the intensification sequence from 29 to 30 August, including RI, sea surface cooling was of $O(0.5^\circ C)$; these low cooling levels occurred over a region where OHC values remained above 90 kJ cm$^{-2}$ (Figs. 6a,b). Over this region of small cooling, in-storm SSTs $>29^\circ C$ prevailed during RI (Figs. 6d,e). Similar RI events over regions of small sea surface cooling where OHC levels are above 80 kJ cm$^{-2}$ were observed during Hurricanes Opal (Shay et al. 2000) and Katrina and Rita (Scharroo et al. 2005; Mainelli et al. 2008; Jaimes and Shay 2009, 2010).

In-storm sea surface cooling of less than 1$^\circ C$ persisted during the 30th hour that Earl was H4, where SSTs prevailed above 29$^\circ C$. On 1 September, Earl weakened to H3 during an eyewall replacement cycle (Cangialosi 2011) as it moved over a region of OHC levels of about 70–90 kJ cm$^{-2}$, where sea surface cooling was approximately 1$^\circ C$. Over this region, in-storm SSTs were above 28$^\circ C$ during this weakening event (Figs. 6d,e). On 2 September, Earl experienced a second intensification and became H4 again. This new intensification stage occurred over a region where OHC levels were between 50 and 70 kJ cm$^{-2}$, and the associated sea surface cooling was less than 1.5$^\circ C$ (Fig. 6f). In-storm SSTs of more than 28$^\circ C$ still extended under the storm during this period of second intensification (Figs. 6d,e).

By 1800 UTC 2 September, Earl started undergoing rapid weakening as it moved over strong horizontal gradients in the OHC structure where heat content levels were decreasing below 50 kJ cm$^{-2}$ in the direction of storm propagation. Over these gradients, in-storm sea surface cooling increased to a maximum value of 2$^\circ C$ on the right side of the storm track (Fig. 6f). Over this region of nearly zero OHC levels and RW, in-storm SSTs sharply decreased to below 26$^\circ C$ (Figs. 6d,e). That is, thermal energy available over this region of RW was not enough to maintain the storm.

4. Air–sea fluxes

Heat and moisture fluxes are one of the most fundamental processes of TCs and are critical to intensity change. Even modest SST differences are believed to alter heat and moisture fluxes in extreme winds (Cione and Uhlhorn 2003). Because it is difficult to get direct measurements of these turbulent fluxes in TCs (e.g., Drennan et al. 2007; Zhang et al. 2008), especially in high-wind conditions, they are usually determined from imperfect “bulk formulae” [(1)] that utilize near-surface atmospheric and upper-ocean temperature data (e.g., Shay et al. 2000; Jacob and Shay 2003; Shay and Uhlhorn 2008; Lin et al. 2013). Key terms in the estimates for the fluxes are temperature and specific humidity differences ($\Delta T$, $\Delta q$) and the bulk transfer coefficients ($C_D$, $C_K$). Recent studies indicate a leveling off of the surface drag coefficient between 28 and 33 m s$^{-1}$ (Powell et al. 2003; Donelan et al. 2004; Shay and Jacob 2006; French et al. 2007; Jarosz et al. 2007; Sanford et al. 2007), and $C_K/C_D$ becomes independent of wind speed (Black et al. 2007; Drennan et al. 2007; Zhang et al. 2008). Based on the ONR CBLAST results, the dependence of the exchange coefficients with wind speed is used here to estimate air–sea fluxes with (1).

a. Algorithm

The algorithm used in this research is based on recent observational studies in hurricanes (Powell et al. 2003; Black et al. 2007; Zhang et al. 2008), and has been successfully tested in numerical experiments (Gopalakrishnan et al. 2013). Within this context, the exchange coefficient of momentum and enthalpy are given, respectively, by

\[
C_D = \begin{cases} 
(4 - 0.6U_{10}) \times 10^{-3} & \text{for } U_{10} < 5 \text{ m s}^{-1} \\
(0.7375 + 0.0525U_{10}) \times 10^{-3} & \text{for } 5 \leq U_{10} < 25 \text{ m s}^{-1} \\
2.05 \times 10^{-3} & \text{for } U_{10} \geq 25 \text{ m s}^{-1}
\end{cases}
\]

and

\[
C_K = \begin{cases} 
(1.5 - 0.25U_{10}) \times 10^{-3} & \text{for } U_{10} < 2 \text{ m s}^{-1} \\
(0.975 + 0.0125U_{10}) \times 10^{-3} & \text{for } 2 \leq U_{10} < 10 \text{ m s}^{-1} \\
1.1 \times 10^{-3} & \text{for } U_{10} \geq 10 \text{ m s}^{-1},
\end{cases}
\]
where \( C_h = C_q = C_K \). Note that (3) is based on direct measurements where the maximum wind speed was 30 m s\(^{-1}\) (Black et al. 2007; Zhang et al. 2008). Dropsonde wind speeds in TC Earl extend well beyond this value, and the ratio \( C_K/C_D \) is considered constant at ~0.54 for \( U_{10} > 25 \) m s\(^{-1}\) (Fig. 7). Based on the CBLAST data, the mean value of the ratio \( C_K/C_D \) is between 0.63 and 0.7. However, the error bars extend from about 0.5 (and even less at 25 m s\(^{-1}\)) to about 0.75 [Fig. 4 in Zhang et al. (2008)]. Thus, (3) considers the observed lower limit for the estimates of the fluxes. Exchange coefficients estimated via absolute angular momentum and total energy budgets based on dropsonde profiler data deployed in major hurricanes indicate that the ratio \( C_K/C_D \) does not significantly increase for wind speeds greater than 50 m s\(^{-1}\) (Bell et al. 2012). Comparisons of direct upper-ocean volume transport induced by Hurricane Frances (2004), and wind-driven volume transport from a numerical model that considers \( C_D \) values from Powell et al. (2003) to calculate wind stresses, indicate that these values of \( C_D \) reproduce reasonable results for wind speeds greater than 50 m s\(^{-1}\), as the modeled volume transport was overestimated by roughly 20% relative to the observations (Sanford et al. 2007). Therefore, in the present treatment, it is assumed that (3) is still valid at the high wind speeds observed during Earl. The uncertainty in the estimates for the fluxes reported here is evaluated in section 6 by comparing the sea surface heat loss, or enthalpy fluxes estimated with (1) and (3), against the OHC response [time evolution of (2)].

Estimates for the fluxes based on (3) and for \( C_K/C_D = 0.7 \) and \( C_K/C_D = 1 \) are also compared in section 6 in relation to this OHC change.

**b. Air–sea fluxes in relation to oceanic variability, and intensity change**

Heat fluxes were estimated with (1) and (3) at dropsonde points using SST\(_{inb}\). Atmospheric data and the corresponding SST and OHC structures, as well as estimates for the ensuing fluxes, were referenced in the storm coordinate system and objectively analyzed into regular grids, independently for each data cluster. These fields are presented in Fig. 8, where the interest is on structures within a radius of \( 4R_{max} \) (hereafter referred to as storm’s inner structure). The OHC\(_{pre} \) structures are used in this discussion because in-storm OHC structures, in addition to sea surface cooling by air–sea fluxes, are affected by thermal responses at the OML base (section 6a).

During RI (C2, see Fig. 4c), OHC levels under the storm’s inner core were on average 10 kJ cm\(^{-2}\) higher compared with those observed at TS conditions (C1), reaching maximum values of 120 kJ cm\(^{-2}\) (Figs. 8.1a, 8.2a). Enthalpy fluxes increased in average from 600 W m\(^{-2}\) in C1 to 1000 W m\(^{-2}\) in C2 (Figs. 8.1d, 8.2d), whereas the SST structure was essentially flat (Figs. 8.1b, 8.2b). Enthalpy fluxes during RI were clearly dominated by moisture fluxes (Figs. 8.1d–f; note the difference in color scale); sensible heat fluxes only represent between 25% and 28% of the total enthalpy fluxes as found elsewhere (Shay and Uhlhorn 2008; Lin et al. 2009; Uhlhorn and Shay 2012).

High OHC levels of more than 110 kJ cm\(^{-2}\) were observed during the first time the storm was H4 (C3, Fig. 8.3a), and maximum enthalpy fluxes were at 1000 W m\(^{-2}\) (Fig. 8.3d). A salient aspect of this nearly steady-state stage (see Fig. 4) is that the region of maximum enthalpy fluxes was larger than the area of maximum momentum fluxes (Figs. 8.3c,d). Note that enthalpy fluxes of more than 800 W m\(^{-2}\) occurred far away from the eyewall, over most of the \( 4R_{max} \) region. When Earl weakened to H3 (C4), OHC levels were between 70 and 90 kJ cm\(^{-2}\) (Fig. 8.4a). Here, enthalpy fluxes reached their overall maximum strength at ~1100 W m\(^{-2}\) (Fig. 8.4d). The underlying OHC levels presumably decreased from C3 to C4 because the storm moved over an ocean environment where stratification and horizontal OHC (and SST and OML) gradients tightened in the direction of storm motion (see Fig. 6). Increased air–sea fluxes have been observed in less severe atmospheric conditions over horizontal thermal gradients in mesoscale eddies and oceanic fronts (Small et al. 2008). A second maximum peak in enthalpy fluxes extended beyond a distance from (3 to 4)\( R_{max} \) in the direction of storm motion (Fig. 8.4d), and was presumably associated with an eyewall replacement.
cycle that could have contributed to the weakening of the storm from C3 to C4 (Cangialosi 2011). The horizontal extension of the region of maximum enthalpy fluxes was much larger than the one for momentum fluxes (Figs. 8.4c,d). The hypothesis is that the underlying OHC structure, from the actual and past storm positions, was energetic enough to maintain two eyewalls via enhanced enthalpy fluxes over a relatively large area.

Earl reached its peak intensity the second time it was H4 (C5, Fig. 8.5c). Note that this period of overall maximum wind intensity followed the eyewall replacement cycle that occurred in C4. A pause in intensification is typically observed in intense TCs during eyewall replacement cycles (Willoughby et al. 1982). In spite of stronger winds, enthalpy fluxes were lower than the first time the storm was H4 (Figs. 8.3d, 8.5d). This decrease in enthalpy fluxes was clearly related to a reduction in OHC

---

**FIG. 8.** Along-track variability of the oceanic thermal structure [(1a)–(7a) OHC pre; (1b)–(7b) SST in]; (1c)–(7c) momentum, (1d)–(7d) enthalpy, as well as (1e)–(7e) latent and (1f)–(7f) sensible fluxes (note the difference in scale) during TC Earl, for clusters (top row) 1 to (bottom row) 7. Estimates for fluxes are based on (3). Circles are for radial distance at 1R max intervals (1–4), where R max is from NHC 6-h best track data. Regions with mapping error larger than 40% are blanked out in these OA fields. Vertical lines are for the along-track direction, where storm motion is in the upward direction. See Table 4 for details on the dominant intensity tendency in each cluster.
levels of ~50% from C3 to C5 (120 and 60 kJ cm\(^{-2}\) respectively; Figs. 8.3a, 8.5a). The hypothesis is that, once the eyewall replacement cycle (pause in intensification) was completed in C5, the storm attained its peak intensity despite low OHC levels and reduced enthalpy fluxes. From this point, OHC levels were not energetic enough, enthalpy fluxes were shut down, and the storm experienced a RW (C6, Figs. 4c and 8.6d). Earl continued its weakening stage in C7 (nearly RW, Fig. 4c), where enthalpy fluxes reversed direction and exerted buoyancy forcing onto the underlying ocean (Fig. 8.7d).

In summary, moisture fluxes dominated the enthalpy fluxes, and these fluxes were affected by underlying ocean structures through the SST, suggesting strong buoyancy forcing of the hurricane boundary layer by warm, deep oceanic features that were not significantly cooled during RI and the time Earl remained as a major hurricane (Fig. 6f), as shown in previous studies (e.g., Shay et al. 2000; Jaimes and Shay 2009, 2010; Lin et al. 2009, 2013). In addition, the mismatch between momentum and enthalpy flux structures indicates that wind-induced surface heat exchange may not be the leading intensification mechanism during Earl, in agreement with recent numerical experiments where local buoyancy generated by sea-to-air vapor fluxes (latent instability) supported generation of deep localized convection (near-surface convergence) that stretched the preexisting vertical vorticity, creating localized cores of intense cyclonic vorticity and diabatic heating (Montgomery et al. 2009). Results presented in Fig. 8 are consistent with these numerical results, where warm oceanic features could play a critical role in delineating the buoyancy structure that forces the hurricane boundary layer.

c. Thermodynamic disequilibrium and near-surface enthalpy

Horizontal structures in \(\Delta q\) and \(\Delta T\) (thermodynamic disequilibrium) were largely impacted by the upper-ocean thermal structure rather than SST structures (cf. columns a and b in Fig. 8 with columns b and c in Fig. 9). The visual correlation between the variability in OHC and \(\Delta q\) is particularly striking (Fig. 9, columns a and b). Compared with conditions at TS intensity level (C1), RI in C2 was associated with increased thermodynamic disequilibrium over warm oceanic features. This RI event was related to the overall peak in \(\Delta q\) as well as to a tightened radial gradient in this parameter (Fig. 9.2b), suggestive of strong buoyancy forcing from the ocean. During the intensification sequence from C1 to C2 to C3, the lowest 100-mb (1 mb = 1 hPa) layer became progressively saturated, attaining values of relative humidity above 95% over most of the storm’s inner-structure region (Figs. 9.1d, 9.2d, 9.3d). During nearly steady state at H4 (C3), the horizontal structure in moisture disequilibrium became nearly axisymmetric (Fig. 9.3b), and \(\Delta T\) reached its overall peak (Fig. 9.3c). The thermodynamic disequilibrium was gradually reduced during C4, C5, and C6 (RW cluster), and eventually became negative over regions where OHC vanished (C7) (e.g., SSTs decreased below 26°C). Within this context of reduced thermodynamic disequilibrium, intensification in C5 appears to be related to high levels in \(\theta_{em}\) (equivalent potential temperature \(\theta_e\), vertically averaged over the lower 100 mb) of more than 360 K that uniformly extended outside \(R_{max}\) (values at the center were of more than 370 K). This nearly axisymmetric pattern in \(\theta_{em}\) was only observable in C5 (Fig. 9.5f), and is presumably related to enhanced enthalpy fluxes that occurred in C4 during the eyewall replacement cycle (Fig. 8.4d). Note that this large increase in \(\theta_{em}\) can also be related to adiabatic expansion.

In a recent study based on composite analyses from several storms it was found that, on average, TCs in the North Atlantic intensified when the near-surface atmospheric environment was drier (lower values in \(\theta_e\)) than in weakening storms; \(\Delta q\) was mainly controlled by the specific humidity of the atmospheric environment (Cione et al. 2013). Here (section 6c), evidence is presented showing that in the Earl case the saturation specific humidity at the SST mainly controlled \(\Delta q\) during rapid intensification over warm oceanic features (oceanic control via enhanced buoyancy forcing).

5. Upper-ocean heat loss and intensity change

Sea surface heat loss

To gain further insight into the impact of enthalpy fluxes on storm intensity, fluxes are integrated during 8 h for each cluster. For consistency, this time interval was chosen because it is the shortest time frame for the clusters (C1, Table 4). This integration yields estimates for the sea surface heat loss (Shay and Uhlhorn 2008). The sea surface heat loss \(\chi\) as a function of the cross-track distance \(r_x\) is given by

\[
\chi(r_x) = \int_{t_1}^{t_2} Q_h(t) dt, \tag{4}
\]

where \(Q_h = Q_s + Q_l\), \(t = r_x/U_h\) is time, \(U_h\) is the translation speed of the storm, and \(r_x\) is the along-track distance. Note that this time integration corresponds to along-track integration over a distance \(r_x\) traveled by the storm in 8 h at speed \(U_h\). Here \(U_h\) is defined as the storm’s mean translation speed during the time frame of the cluster; its STD is used to define the error bars in (4). The along-track length for integration, \(r_x\), is not necessarily...
the same for all data clusters because $U_h$ changed (it ranges from approximately $(-2 \text{ or } -3)R_{\text{max}}$ to $2 \text{ or } 3$ $R_{\text{max}}$ during individual integrations centered at the inner-core region). The quantity $t = r_j/U_h$ is, however, approximately the same (8 h).

With the exception of C7 (weakening cluster), the maximum 8-h sea surface heat loss occurred under the eyewall region $(1-2)R_{\text{max}}$, Fig. 10). Maximum values of $1.3 \text{kJ cm}^{-2}$ were estimated at TS intensity. During RI (C2), $\chi$ attained maximum values of $\sim 2.2 \text{kJ cm}^{-2}$ at the eyewall, or $\sim 1.7$ times the heat loss at TS intensity. Values of more than $1.7 \text{kJ cm}^{-2}$ extended to a radius of $\sim 5R_{\text{max}}$ under the influence of deep OHC structures (see Fig. 8.2a). An 8-h sea surface heat loss of $2.6 \text{kJ cm}^{-2}$ ($2 \text{ times the value of } \chi$ at TS intensity) was required to maintain the storm at a nearly steady-state H4 (C3); this value represents a total heat loss of $\sim 9.75 \text{kJ cm}^{-2}$ for the 30 h the storm maintained this status. A secondary eyewall is clearly seen in the $\chi$ structure for C4 at $8R_{\text{max}}$ (Fig. 10b), where 8-h surface
heat loss values of 2 kJ cm$^{-2}$ were estimated. This supports the idea that an eyewall replacement cycle contributed to storm weakening from C3 to C4 (Cangialosi 2011). The $\chi$ structure flattened during RW (C6), where maximum values of ~1.7 kJ cm$^{-2}$ were estimated. During weakening to TS status (C7), the sea surface gained heat from Earl by about 0.8 kJ cm$^{-2}$ in 8 h.

6. Discussion

a. Sensitivity to the $C_K/C_D$ ratio

Given the known uncertainty in the values of exchange coefficients of enthalpy and momentum and its ratio ($C_K/C_D$), it is of interest to evaluate the estimates for the fluxes within the context of different published values for this ratio. Thus, two other sets of estimates for the fluxes were obtained using (3a) for the drag coefficient, considering $C_K/C_D = 0.7$ (green line, Fig. 11) and $C_K/C_D = 1$ (blue line, Fig. 11).

Because OHC is a better predictor than just SST when evaluating the relevant upper-ocean cooling response (since it includes the depth of the 26°C isotherm as shown above), it is of interest to compare these estimates for the fluxes in relation to the rate of change in OHC during the forced stage (when the storm is overhead), assumed here as the difference between OHC$_{\text{pre}}$ and OHC$_{\text{in}}$ (section 2c). The rate of change in OHC is defined as $\Delta\text{OHC}/\Delta t = (\text{OHC}_{\text{in}} - \text{OHC}_{\text{pre}})/\Delta t$. Note that in addition to the sea surface heat loss by enthalpy fluxes $\Delta\text{OHC}/\Delta t$ can be affected by the horizontal and vertical advection of oceanic thermal structure and by irreversible upper-ocean vertical mixing. Thus, in-storm $\Delta\text{OHC}/\Delta t$ is assumed to represent the upper bound for the estimates of the fluxes. For future work, it might be instructive to derive this expression [i.e., differentiate (2)] and examine the individual contributions using in situ data.

The prestorm OHC horizontal structure indicates that Earl moved over three ocean regimes (Fig. 6a): a weakly stratified, warm deep regime (from 28 August to 1 September); a transition regime where OHC gradually decreased in the direction of storm propagation, indicative of increasing temperature stratification (from 1 to 3 September); and a cool, highly stratified regime where OHC vanished (beyond 4 September). Consistent with these stratification regimes, the rate of change in OHC was smaller over the warm regime, indicative of a small cooling response (because of the deep thermal structure); it attained maximum values over the transition regime (because of the tighter horizontal and vertical gradients in the thermal structure); and it nearly vanished over the cool regime where the ocean became thermodynamically forced by downward enthalpy fluxes (Fig. 11).

Within the context of this OHC response, fluxes that consider $C_K/C_D = 1$ are overestimated (Fig. 11). With exception of the peaks on 1 and 2 September, the sea surface heat loss estimated under this approach is larger than $\Delta\text{OHC}/\Delta t$ over the warm regime, and nearly equal over the transition regime. Fluxes estimated with (3) and with $C_K/C_D = 0.7$ are within the envelope of $\Delta\text{OHC}/\Delta t$ over the warm regime, are smaller than $\Delta\text{OHC}/\Delta t$ over the transition regime, and are nearly equal to $\Delta\text{OHC}/\Delta t$ over the cold regime.

Some important conjectures can be made within the context of these results. 1) The heat loss to TCs over warm ocean regimes can be comparable to the total heat loss over the upper ocean during the forced stage. This is not a surprising result because it is well known that upper-ocean cooling by vertical mixing over cooler
thermocline waters is very small in warm, deep ocean features (e.g., Shay et al. 2000; Jaimes and Shay 2009, 2010; Jaimes et al. 2011; Lin et al. 2013). 2) The heat loss to the storm over the ocean transition regime is between 15% and 40% of the total in-storm OHC response. These numbers are comparable to those usually reported in the literature (e.g., Price 1981; Jacob et al. 2000; Cione and Uhlhorn 2003). 3) This variability in upper-ocean cooling responses underscores the need for including realistic ocean structures in coupled numerical models for hurricane intensity forecasting. 4) Values of 0.54 (Fig. 7) and 0.7 for \( \frac{C_K}{C_D} \) reproduced reasonable estimates for the sea surface heat loss in relation to \( \frac{\Delta \text{OHC}}{\Delta t} \). 5) Using (3) for wind speeds greater than 30 m s\(^{-1}\) seems to be a reasonable approach, and it also supports recent findings that indicate the ratio \( \frac{C_K}{C_D} \) does not significantly increase for wind speeds greater than 50 m s\(^{-1}\) (Bell et al. 2012). 6) Field experiments are needed to obtain direct measurements of the upper-ocean temperature and current responses in TCs. Data from these measurements can be used to evaluate or challenge the hypothesis proposed here that in-storm \( \frac{\Delta \text{OHC}}{\Delta t} \) represents the upper bound for the fluxes.

Based on numerical experiments, it was recently proposed that the uncertainty in the drag coefficient might be irrelevant to the intensity problem, as allowing random variations in \( C_D \) within an envelope of 60% with respect to its mean profile (computed from CBLAST data) for wind speeds above the capping limit, did not impact TC intensity (Thomsen et al. 2012). A caveat of these experiments is that they used a constant value of SST = 27°C. That is, upper-ocean cooling was prevented in the numerical experiments. This cooling directly affects the most fundamental process of TCs, the sea-to-air transport of enthalpy. As shown by Halliwell et al. (2011), the OML velocity and temperature responses to TC wind forcing are quite sensitive to the choice of \( C_D \) in numerical models. Thus, given the inherent difficulties of acquiring near-surface data in the lower 10–20 m of the hurricane boundary layer, it can be argued that a better way to infer the surface drag coefficient is to use the ocean’s momentum response to TC forcing, as suggested herein and in other studies (Shay and Jacob 2006; Jarosz et al. 2007; Sanford et al. 2007).

b. Impact of wind speed and thermodynamic disequilibrium on enthalpy fluxes

Within the context of the wind-induced surface heat exchange intensification paradigm, enthalpy fluxes are predicted to increase with wind speed. However, this is not necessarily the case during Earl (Fig. 8). To further illustrate this point, consider the enthalpy fluxes and \( U_{10} \) referenced in geographic coordinates (Fig. 12). Regions of maximum \( Q_h \) are not necessarily collocated with regions of maximum \( U_{10} \), in particular, over areas of

![FIG. 11. Sensitivity of the sea surface heat loss (by enthalpy fluxes, \( Q_h \)) to the ratio \( \frac{C_K}{C_D} \), based on (3) (red line; also see Fig. 7); \( \frac{C_K}{C_D} = 0.7 \) (green line); and \( \frac{C_K}{C_D} = 1 \) (blue line). This along-track analysis considers data clusters referenced in geographic coordinates. Data from these clusters are merged into a single structure along the track via OA. Positive values in \( \Delta \text{OHC}/\Delta t \) indicate a decrease in OHC; OHC\(_{\text{in}} \) and OHC\(_{\text{pre}} \) are from Fig. 6. Values for \( Q_h \) and \( \Delta \text{OHC}/\Delta t \) are averaged in the cross-track direction from \( -9 R_{\text{max}} \) to \( 9 R_{\text{max}} \); bars and gray shading are used for the STD. The outer dark gray shade in the STD of \( \Delta \text{OHC}/\Delta t \) accounts for an additional 15% error based on rms differences between direct and satellite measurements of OHC (Shay and Brewster 2010). The bottom colored line indicates storm category on the Saffir–Simpson scale.

![FIG. 12. Along-track variability of standard 10-m wind speed (\( U_{10} \), color) and enthalpy fluxes (\( Q_h \), contours). Regions with mapping errors larger than 40% in \( U_{10} \), and beyond a distance greater than 9\( R_{\text{max}} \) from the storm’s center, are blanked out in these OA fields. Contours for \( Q_h \) are at 500, 600 (boldface line), and 800 W m\(^{-2}\).]
localized maxima that developed during the intensification sequence from H1 to H4.

To gain further insight into the particular processes that are impacting the horizontal structures in enthalpy fluxes, these fluxes are projected onto horizontal structures of $\Delta T$ (thermal disequilibrium), $\Delta q$ (moisture disequilibrium), relative humidity, and OHC in geographic coordinates (Fig. 13). One of the most striking aspects of the enthalpy fluxes is their along-track variability. These fluctuations were impacted by the meandering nature of the Antilles Current (Fig. 13d). The similarity between $\Delta T$ and $Q_h$ is remarkable (Fig. 13a), where $\Delta T$ attains maximum values over regions of maximum OHC levels. Because Earl’s low-level layer is nearly isothermal within a radius of $4R_{\text{max}}$ (Fig. 9, column e), $\Delta T$ is mostly controlled by oceanic structures. Smith and Montgomery (2013) found nearly isothermal low-level layers in Earl’s inner core after analyzing two of the GRIP flights used in the present study.

Because moisture fluxes were the major contributor to the enthalpy fluxes (Fig. 8, columns d–f), the contribution to the elevation of $\theta_e$ by moisture fluxes is substantially larger than the contribution from sensible heat fluxes [also see Smith and Montgomery (2013)]. The moisture disequilibrium was also impacted by OHC structures (Fig. 13b), as the larger values in $\Delta q$ occurred on the left side of the storm track over regions where OHC exceeded 100 kJ cm$^{-2}$ (the blanked-out regions on the left side of the storm track had OHC values larger than this value).

The physical processes observed in Earl are summarized in Fig. 14. Enthalpy fluxes forced intensity fluctuations. This buoyancy forcing grew faster, peaked earlier, and decayed earlier than the maximum wind speed. This result suggests that, rather than controlling heat intake, the wind field adjusted to horizontal pressure gradients created by buoyancy forcing. The increased wind field could in turn enhance wind-induced surface heat exchange as found elsewhere (e.g., Lin et al. 2013). Thus, the latter mechanism was not necessarily the main process controlling enthalpy fluxes into Earl.
Rather than being controlled by just SSTs, the enthalpy fluxes were influenced by underlying OHC structures. Maximum values of the heat fluxes were attained over warm, deep structures where sea surface cooling was small. In contrast, the enthalpy fluxes vanished as the OHC was reduced (SSTs less than 26°C). A steep increase in enthalpy fluxes was observed during rapid intensification (RI) (C2), in comparison to fluxes that occurred during the previous storm stage.

Cluster-averaged enthalpy fluxes are about 18% smaller for $C_K/C_D = 0.54$ [(3)] than for $C_K/C_D = 0.7$ at peak values (C4, Fig. 14). The average difference between these two sets of estimates is $-73.1$ W m$^{-2}$ with an STD of 46.6 W m$^{-2}$. In contrast, the average difference in estimates calculated with satellite-based prestorm TMI SST (not shown) and SST$_{pre}$ is $-83.8$ W m$^{-2}$ with an STD of 99.7 W m$^{-2}$. These results indicate that the uncertainty in the value of $C_K/C_D$ can have a comparable impact to using satellite-based prestorm SSTs rather than in situ SSTs in the estimates for the fluxes.

c. Relative contribution of oceanic and atmospheric saturation specific humidity

Given the key role of moisture disequilibrium in enhancing latent heat fluxes and storm intensity, especially during RI, it is important to evaluate the relative contributions of the oceanic and atmospheric environments to this disequilibrium. As shown in Fig. 15, the overall magnitude of $\Delta q$ was mainly controlled by the saturation specific humidity at the SST rather than by the specific humidity of the atmospheric environment. The increase in $\Delta q$ during RI was clearly impacted by the underlying deep oceanic structure that facilitated maintaining relatively high SSTs during storm passage.

7. Concluding remarks

As observed during the rapid deepening of other major TCs (e.g., Shay et al. 2000; Shay and Uhlhorn 2008; Mainelli et al. 2008; Jaimes and Shay 2009, 2010; Shay 2010; Lin et al. 2009, 2013), Earl’s RI occurred over warm, deep thermal structures where in-storm sea surface cooling was $O(0.5^°C)$, and SSTs above 29°C prevailed during storm passage. Enthalpy fluxes of about 1100 W m$^{-2}$ (approximately 6 times the climatological value over the region of study) were estimated during peak storm intensity. An increase in enthalpy fluxes of 3 times the climatological value was estimated during the RI period of Typhoon Nargis, where maximum values during peak storm intensity were at 900 W m$^{-2}$ (Lin et al. 2009).

By integrating 8-h values (Fig. 10), daily sea surface heat losses of $-6.5 \pm 0.8$ and $-7.8 \pm 1.1$ kJ cm$^{-2}$ were estimated for RI (C2) and the mature stage of the storm (C3), respectively. By contrast, daily sea surface heat losses of $2.3 \pm 0.7$ kJ cm$^{-2}$ from the TC to the ocean during Earl’s weakening from H1 to TS were observed. Momentum and enthalpy fluxes integrated over the full storm wake (or during 7 days; see Table 5 for equations and units) represent approximately power from 2.2 to
where an annual mean value of 200 W m\(^{-2}\) at the top of OHC during the forced stage. This result indicates that is markedly consistent with the rate of change in 2008) reproduce a sea surface heat loss by enthalpy fluxes of the CBLAST results (Black et al. 2007; Zhang et al. 2008) suggest that estimated solar radiative flux. Studies based on numerical experiments suggest that baroclinic instability below the oceanic mixed layer base can reduce the time of restoration of SSTs in the wake of storms by about 50% (Mei and Pasquero 2012). These results would put our estimates at ~8 days; e-folding times of 5–20 days have been observed for the SST cool anomaly to disappear in the wake of moving hurricanes (Price et al. 2008). Note that Uhlhorn and Shay (2012) found significant near-surface warming/restratification only a few days after Lili’s passage (see their Fig. 9), and the average warming was consistent with estimated solar radiative flux.

One of the key results here is that surface exchange coefficients of momentum and enthalpy defined in terms of the CBLAST results (Black et al. 2007; Zhang et al. 2008) reproduce a sea surface heat loss by enthalpy fluxes that is markedly consistent with the rate of change in OHC during the forced stage. This result indicates that a ratio of \(C_K/C_D\) between 0.54 and 0.7 produces reasonable flux estimates, even during RI of major storms. By contrast, a ratio \(C_K/C_D = 1\) or larger, as proposed by Emanuel (1995) within the context of the theoretical maximum potential intensity, will likely overestimate the fluxes in relation to the rate of change in OHC. There is consistency between estimates for the fluxes that consider a ratio of \(C_K/C_D\) between 0.54 and 0.7 for 10-m wind speeds larger than 30 m s\(^{-1}\), and the change in OHC. This result suggests that these ratios can be used in high-wind speed conditions, in agreement with recent results based on budgets of angular momentum and total energy (Bell et al. 2012). The difference in flux estimates calculated with these two ratios indicates that the uncertainty in the value of \(C_K/C_D\) can have a comparable impact to the case when using satellite-based prestorm SSTs rather than in situ SSTs in the estimates. Investigation of more cases is needed to generalize these results, including in situ ocean measurements juxtaposed with atmospheric measurements (e.g., D’Asaro et al. 2011, 2014). Such data provide a methodology for relating the change in OHC (including the OML and oceanic vertical shear estimates) to the inferred heat and moisture fluxes for observational and numerical studies (Uhlhorn and Shay 2012, 2013).

The enthalpy fluxes in Earl were largely controlled by the moisture disequilibrium (latent instability) between the sea surface and near-surface air, rather than by wind speed. This disequilibrium was strongly influenced by underlying upper-ocean warm thermal structures. Localized maxima in enthalpy fluxes developed over tight horizontal gradients of moisture disequilibrium colocated with warm oceanic features; these regions of local buoyant forcing preferentially developed during RI. These results support the hypothesis that intense local buoyant forcing by the ocean is an important intensification mechanism (e.g., Van Sang et al. 2008; Montgomery et al. 2009). Rather than being controlled by stochastic bursts (Van Sang et al. 2008), structure and intensity fluctuations in TCs may be associated with bursts in enthalpy fluxes caused by thermodynamic disequilibrium over localized warm oceanic structures. That the ocean forces the atmosphere via air–sea differences in temperature and humidity over oceanic fronts and mesoscale eddies has been documented in less severe atmospheric conditions (e.g., Small et al. 2008).

During the Impacts of Typhoons on the Ocean in the Pacific (ITOP) field campaign, it was found that in the case of Typhoon Megi \(\Delta T\) and \(\Delta q\) remained nearly constant and enthalpy fluxes increased with increasing wind speed; in the case of Typhoons Fanapi and Malakas \(\Delta T, \Delta q\), and enthalpy fluxes decreased with increasing wind speed (Lin et al. 2013). In the Earl case reported here, enthalpy fluxes grew faster, peaked earlier, and decayed faster than the maximum wind speed (Fig. 14). The results from the Earl case suggest, that rather than controlling the enthalpy fluxes, the wind field adjusted to the pressure gradient created by the buoyancy forcing over warm deep oceanic regimes. The different methods used in analyzing the Earl case (cluster analysis) and ITOP data complicate comparing results from these experiments. Further research is guaranteed to generalize conclusions.
In conclusion, upper-ocean thermal structures modulate SST cooling responses to TC wind forcing, which has been well documented elsewhere. These cooling responses directly control a fundamental process of TCs: the sea-to-air transfer of enthalpy. Realistic upper-ocean thermal structures must be incorporated into ocean–atmosphere coupled TC models, for them to reproduce realistic SST cooling patterns, enthalpy fluxes into the storm, and ensuing intensity fluctuations. This is a first-order issue in the intensity problem, in particular during RI that is still not well documented in coupled models. The large cross-track variability in the OHC changes (gray shade in Fig. 11) implies that this is critical to accurately reproduce the storm track in numerical models in relation to realistic upper-ocean thermal structures.

Acknowledgments. The research team gratefully acknowledges support from the NASA Hurricane Science Program (NASA Award NNX09AC47G), Deep-C (Grant SA1212GoMRI008), and NOAA/NESDIS. The project continues to be grateful to the NOAA Aircraft Operations Center (Dr. Jim McFadden), who made it possible to acquire high-quality data during hurricanes through the Hurricane Forecast Improvement Project (HFIP) and the collaborative ties with NOAA’s Hurricane Research Division directed by Dr. Frank Marks at AOML. Kathryn Sellwood (NOAA/HRD) provided valuable advice during the processing of the raw dropsonde data with ASPEN. Dr. George Halliwell (NOAA/AOML) provided the USGODAE datasets. Comments and suggestions by three anonymous reviewers contributed to significantly improving this paper. NOAA/HRD provided raw IFEX and USAFR dropsonde data. GRIP NASA DC-8 quality controlled dropsonde data were provided by NCAR/EOL under sponsorship of the National Science Foundation (http://data.eol.ucar.edu). Processed altimeter data are from the U.S. Navy’s Altimetry Data Fusion Center (ADFC) at Stennis Space Center; the derived product suite (and its evaluation) is available online (either at http://www.rsmas.miami.edu/groups/upper-ocean-dynamics/research/ocean-heat-content/ or http://www.ospo.noaa.gov/Products/ocean/ocean_heat.html).

REFERENCES


