1.3.1: Introduction

The objective of this report is to document progress since IWTC-V in this thematic area from experimental, observational, empirical, theoretical, and numerical perspectives. The report begins by describing progress in upper-ocean processes that include the oceanic mixed layer (OML) and the thermocline. This section is followed by a discussion of the air-sea interface that includes surface winds and waves, and the communication to the atmospheric boundary layer through the momentum and enthalpy fluxes across the interface. These findings are summarized within a global context with specific recommendations on these important science issues to the WMO Commission on Atmospheric Science.

1.3.2: Upper-Ocean Processes

Coupled oceanic and atmospheric models to predict hurricane intensity and structure change will eventually be used to issue forecasts to the public who increasingly rely on the most advanced weather forecasting systems to prepare for landfall (Marks and Shay 1998). For such models, it has become increasingly clear over the past decade that ocean models will have to include realistic initial conditions to simulate not only the oceanic response to hurricane forcing (Price 1981, 1994; Sanford et al. 1987; Shay 2001; D'Asaro 2003, Jacob and Shay 2003), but also to simulate the atmospheric response to oceanic forcing (Shay et al. 2000, 2006; Hong et al. 2000; Walker et al. 2005; Lin et al. 2005; Wu et al. 2006).

An important example of this later effect was observed during the passages of hurricanes Katrina, Rita, and Wilma during the 2005 Atlantic Ocean hurricane season. Favorable atmospheric conditions prevailed in the Northwest Caribbean Sea and Gulf of Mexico (GOM) as the Loop Current (LC) extended several hundred kilometers north of the Yucatan Strait. As these storms moved over the deep warm pools, all three hurricanes...
explosively deepened and were more closely correlated with the ocean heat content (OHC) variations (and deep isotherms) than with the sea-surface temperatures (SST) distributions, which were essentially flat and exceeded 30°C over most of the region with slight warming along the northern GOM shelf. By contrast, the OHC and 26°C isotherm depths indicated the LC and its deep warm layers as it was in the process of shedding a mesoscale warm core ring (WCR) in August and September 2005. A cold core ring (CCR) that advected cyclonically around the shed WCR may have helped weaken Rita before landfall. Walker et al. (2005) also found that Hurricane Ivan may have encountered a CCR prior to landfall. These studies point to the importance of initializing coupled ocean-hurricane models with realistic warm and cold ocean features.

As shown in Figure 1.3.1b, upper-ocean mixing and cooling are a strong function of forced near-inertial current shears that reduce the Richardson numbers below criticality, which induces entrainment mixing (Price 1981; Schade and Emanuel 1999; Shay 2001; D’Asaro 2003; Jacob and Shay 2003). The contributions to the heat and mass budgets by shear-driven entrainment heat fluxes across the ocean mixed layer (OML) base are 60 to 85%, surface heat fluxes are between 5 to 15%, and horizontal advection by ocean currents are 5 to 15% (Price et al. 1994; Jacob et al. 2000). All of these processes impact the SST and OHC variability. In addition, wind-driven upwelling of the isotherms due to net upper ocean transport away from the storm modulate the shear-induced (S) ocean mixing events by an upward transport of cooler water from the thermocline. This transport increases the buoyancy frequency (N), which tends to keep the Richardson number above criticality. In the LC and WCR regimes with deep, warm thermal layers, cooling induced by these physical processes (Fig. 1) is considerably less as much greater turbulent-induced mixing by the current shear is required to cool and deepen the OML (Shay et al. 2000, Uhlhorn...
and Shay 2006). Quantifying the effects of forced current (and shear) on the OHC and SST distributions is central to accurately forecasting hurricane intensity and structure change.

Figure 1.3.2: OHC map and inset showing NRL mooring locations (red) and SRA wave measurements (black) relative to Ivan’s storm track and intensity. The OHC pattern shows the WCR encountered by Ivan prior to landfall. The cooler shelf water (OHC < 20 KJ cm\(^{-2}\)) resulted from the passage of Frances two weeks earlier.

1.3.2.1 Oceanic Response

Hurricane Ivan (2-24 September 2004) moved over the NW Caribbean Sea with a radius of maximum wind (R\(_{\text{max}}\)) of ~36 km (Fig. 1.3.2). Favorable environmental conditions of high OHC water (>150 KJ cm\(^{-2}\)) plus outflow enhanced by upper-atmospheric flow ahead of an approaching trough helped Ivan maintain Cat-5 strength over 24-30 h (about one inertial period). Upon entering the GOM as a Cat-5 storm, Ivan weakened to a Cat-3 storm due to a combination of lower OHC, vertical shear in the atmosphere associated with an upper-level trough, and dry air being drawn into its circulation. Within 12 hours of landfall, Ivan encountered a WCR and surface pressure decreased by ~10 mb. However, cooler GOM shelf water forced by Hurricane Frances (10 days earlier) along with increasing shear both acted to oppose intensification during an eyewall replacement cycle. Thus, Ivan was an example of the impact of alternating positive and negative oceanic feedbacks on hurricane intensity.

As shown in the Fig. 1.3.2 inset, the Naval Research Laboratory (NRL) had previously deployed several moorings about 180 km south of Mobile, Alabama as part of the Slope to Shelf Energetics and Exchange Dynamics (SEED) project from early May through early November 2004. Hurricane Ivan passed directly over these Acoustic Doppler Current Profiler (ADCP) moorings (Teague et al. 2005), which provided the temporal evolution of the 3-D current (and shear) structure at 2 to 4 m vertical resolution (Fig. 1.3.3). The current response at one of the Ivan moorings is shown for the near-inertial wave band (band-pass filter of the detided current signals) where the local inertial period is ~24 h. Profiler data starting from 50 m and extending to 492 m (4-m bins) were filtered between 22 to 28
hours. Second, the profiler data that were vertically averaged to estimate the depth-averaged current response (upper panels of Fig. 1.3.3) have similar amplitudes of 5 to 8 cm s\(^{-1}\) as those found in the limited vertical sampling from the Hurricane Frederic arrays (Shay and Chang 1997). These barotropic signals arrive before the storm owing to the large phase speeds of these oscillations.

![Figure 1.3.3: Band-pass filtered cross-track (u) and along-track (v) time series at mooring 8 with t = 0 the time of hurricane passage and the time axis normalized by the inertial period. Shown top to bottom are depth averaged cross-track, baroclinic cross-track after the depth-average is removed (contoured), along-track depth-average, and along-track baroclinic (contoured). Time interval spans 15-29 September (Data courtesy NRL-Stennis).](image)

After the depth-averaged flows are removed, the band-pass filtered signals (lower panels in Fig.1.3.3) illustrate the ocean baroclinic response to hurricane forcing. Notice the downward propagation of the baroclinic energy from the wind-forced OML into the thermocline, which is consistent with modal separation (Gill 1984). The baroclinic motions have a characteristic time scale for the phase of each mode to separate from the wind-forced OML when the wind stress scale (2R\(_{\text{max}}\)) exceeds the deformation radius associated with the first baroclinic mode (~ 30 to 40 km). This time scale increases with mode number due to decreasing phase speeds (Shay 2001). This vertical propagation is primarily associated with the predominance of the clockwise rotating energy (Shay and Jacob 2006). The vertical velocity signal in the upper ocean induced by Ivan was about 1.5 cm s\(^{-1}\).

As shown in Fig. 1.3.4, similar strong near-inertial current response was observed by ElectroMagnetic Autonomous Profiling Explorer (EM-APEX) floats deployed in front of Hurricane Frances (2004) by Sanford et al. (2006). These Lagrangian profiling floats have provided a new view of near-inertial, internal wave radiation in unprecedented detail that includes not only the temperature and salinity (and thus density), but also the horizontal velocity structure. The phase propagation of the forced near-inertial currents is upward and
is associated with downward energy propagation from the wind-forced OML. The shears (S) associated with these wind-forced currents lead to mixing events that significantly contribute to the observed ocean cooling of 2 to 2.5°C (and deepening) of the OML under the hurricane. Since the profilers also measure the density structure (and buoyancy frequency N), these Lagrangian floats provide the evolution of the Richardson numbers (N² S⁻²) for the first time under strongly forced hurricane conditions. Thus, the EM-APEX floats represent a new tool to improve our understanding of upper ocean processes and variability for a spectrum of conditions. For these near-inertial motions, the currents rotate clockwise with depth as wind-driven energy propagates downward into the thermocline while the phase propagates upward (Leaman 1976), which appears to be the case for the first few IP in the Ivan and Frances measurements. Rates of vertical energy flux forced by hurricanes have an average value of ~2 ergs cm⁻² s⁻¹ (Shay and Jacob 2006).

Figure 1.3.4: Current (U,V in m s⁻¹), salinity (psu) and density or sigma-t (kg m⁻³) at Rmax during the passage of hurricane Frances (2004) as measured by the EM-APEX) floats developed by Webb and APL-University of Washington. Floats were deployed from a WC-130 by the USAF Reserve unit (from Sanford et al. 2006).

1.3.2.2 Ocean Heat Content Variability

If the upper-ocean warm layer is thick and has a large total OHC, the SST will decrease slowly during TC passage, the negative feedback mechanism (or “brake”) will be weak, and the ocean will tend to promote TC intensification. Shay et al. (2000, 2006) showed that the OHC relative to the depth of the 26°C isotherm takes this into account and is a better
indicator than just SST of the potential for TC intensification. Leipper and Volgenau (1972) estimated OHC as:

$$\text{OHC} = c_p \int_0^{D_{26}} \rho [T(z) - 26] \, dz,$$

where $c_p$ is specific heat at constant pressure, $D_{26}$ is the 26°C isotherm depth, and OHC is zero wherever water above 26°C is not present. OHC (and $D_{26}$) is monitored using satellite remote sensing techniques (see http://iwave.rsmas.miami.edu/~nick/heat) using measurements from satellite-based radar altimeter estimates of the surface height anomaly (SHA) field from NASA TOPEX, Jason-1, U.S. Navy Geosat Follow-On-Mission (GFO) (Cheney et al. 1994; Scharroo et al. 2005) and SST cast within a two-layer model (e.g., Goni et al. 1996) and a seasonal climatology (Mainelli-Huber 2000). Since ocean features only move a few km day$^{-1}$, altimeter-derived SHA locates warm (cold) mesoscale features that are usually identified as positive (negative) values as observed during Opal (Shay et al. 2000) and Ivan (Walker et al. 2005). These daily estimates are used in the Statistical Hurricane Intensity Prediction Scheme (SHIPS) to forecast intensity at the Tropical Prediction Center (DeMaria et al. 2005; Mainelli et al. 2006). Of relevance to the prediction problem, regions with thick (100-200 m) warm layers such as the Caribbean Sea, the LC, and WCR have high OHC (i.e., deep warm layers) and provide more sustained heat to the atmosphere during hurricane passage. Other regions such as the interior GOM, where the Gulf Common Water (GCW) has a smaller OHC distributed over a thinner (40 m) warm layer, tend to be less favorable for significant intensification (Jacob and Shay 2003).

As shown in Fig. 1.3.5, both Katrina and Rita deepened to a Cat-5 hurricane over a lobe-like structure along the LC’s western flank as they moved at a speed of 5 to 6 m s$^{-1}$. Notice the one-to-one correlation between hurricane intensity and OHC values of $\sim$120 kJ cm$^{-2}$ in the LC. The SSTs of more than 30°C were nearly uniformly distributed in this regime, and did not reveal the complex LC/WCR structure (Sun et al. 2006; Shay et al. 2006). Thus, the surface fluxes increased over the LC after Katrina emerged over the Gulf of Mexico where OHC values exceeded 100 kJ cm$^{-2}$. This OHC level is more than five times the threshold of 16 kJ cm$^{-2}$ day$^{-1}$ integrated over the storm scale as suggested by (Leipper and Volgenau 1972). To compare along-track pressure fluctuations to along-track OHC and SST variations, longitude, latitude, and pressure (stars on pressure curve in Fig. 1.3.5b) from the best track 6-h data were used to estimate storm position and pressure at 2-hour intervals using linear interpolation. For both storms, the SST curve represents the along-track surface temperatures. Notice that normalized OHC values vary inversely to pressure changes (surface pressure decreases, OHC increases), whereas the normalized SST values are essentially flat during Katrina’s and Rita’s passage.

Lin et al. (2005) analyzed remote sensing imagery in the western Pacific Ocean prior and subsequent to the passage of Maemi in 2003 (Fig. 1.3.6). They found that Typhoon Maemi’s intensity increased by 36 m s$^{-1}$ over a eddy-rich regime. Using results in a Coupled Hurricane Intensity Prediction System (Emanuel 2003), the WCRs acted as an insulator between typhoons and the deeper, cooler thermocline water (Wu et al. 2006). That is, the SST response is significantly less as the OML is much deeper in these regimes.
similar to findings in the western Atlantic Ocean basin. Without initializing the model with a warm ocean feature, the simulated typhoon intensity was one category below the observed intensity. Lin et al. (2005) point out the importance of the eddy-rich regime associated with the western boundary current or Kuroshio during the passage of typhoons over these oceanic features. In the western Pacific Ocean, the Kuroshio plays the same role as the Gulf Stream in the western Atlantic Ocean basin (i.e., poleward advection of warm tropical water).

Figure 1.3.5: Left panels (a,c) represent pre-storm OHC (kJ cm\(^{-2}\):color) and 26°C isotherm depth (m: black contour) based on a hurricane season climatology, Reynolds SSTs, Jason-1 and GFO radar altimetry measurements relative to the track and intensity of Hurricane a) Katrina and c) Rita. Right panels (b,d) represent time series of surface pressure (thin black) versus along-track SST (dashed) and OHC (thick black) variations normalized by 30°C and 60 kJ cm\(^{-2}\), respectively. OHC uncertainty limits are based on 6-hourly values averaged in the cross-track direction between +/- 0.5° lat. from the track (from Shay et al. 2006).

On 15 and 26 September 2005, oceanic current, temperature and salinity measurements were acquired from Airborne eXpendable Current Profilers (AXCP), Airborne eXpendable Conductivity Temperature and Depth (AXCTDs) profilers, and Airborne eXpendable
Bathythermographs (AXBTs) in a pattern centered on the LC/WCR (Rogers et al. 2006; Shay et al. 2006). The 15 September flight was originally conceived as a post-Katrina experiment in an area where the hurricane rapidly intensified over the LC/WCR complex.

To obtain in-situ measurements of OHC within the same WCR, two thermistor chain drifters were deployed from the USAF WC-130 aircraft in the path of Hurricane Rita on 21 September 2005. As shown in Fig. 1.3.7, upper ocean T(z) time series were obtained from two drifters that were closest by Rita (distance in upper panels) while circulating around the WCR. Notice that these two drifters were about 80 to 100 km from the WCR center with OHC values of 120 kJ cm$^{-2}$ where the 26$^\circ$C isotherm depths were between 90 to 100 m. Altimeter-derived OHC values in the WCR ranged from 105-120 kJ cm$^{-2}$ so these altimeter values are consistent with the low-pass filtered values in the right panels prior to Rita. The estimated enthalpy fluxes exceeded 2000 W m$^{-2}$ from one of the drifters (not shown) that passed closest to Rita while it was at Category 4 strength with R$_{max}$ < 20 km. Rita then clipped the northeastern part of the WCR as the storm was weakening prior to landfall on the Texas-Louisiana border as a Cat-3 hurricane. As during Hurricane Opal (Shay et al. 2000; Hong et al. 2000), these in situ measurements and inferred surface
enthalpy fluxes are invaluable to help the TC communities understand the atmospheric response to ocean forcing.

Figure 1.3.7: Thermal structure (°C: left panels) and observed and low-pass filtered (solid black) of OHC (kJ cm⁻²: right panels) time series from two drifters deployed prior to Rita (black curve) relative to the WCR center (red) as depicted in the upper panels for the thermal structure in September 2005. The point of closest approach is shown as the dotted line (figure courtesy of Rick Lumpkin of NOAA/AOML).

Using these in situ profiles combined with radar-altimetry fields, isotherm depths and OHC values are compared to assess uncertainties in satellite retrievals for pre- and post-Rita cases (Fig. 1.3.8). Satellite-inferred and in situ structures are well correlated (0.9) for both isotherm depths and OHC variations using Reynolds SSTs (Shay et al. 2006). Regression slope for the OHC is 0.9 with a bias of 1.3 kJ cm⁻² by combining pre and post-Rita data set in the WCR only. For the 26°C isotherm depths, the slope was about 1.1 with a 9.3 m bias where the altimeter-derived value was larger than that from the profiler data. This larger bias was primarily associated with the advection of the CCR between the LC and WCR from the post-Rita data set on 26 Sept. These estimates were also consistent with those derived from drifter-based measurements. While the bias in the depth is quite large, the result suggests this is roughly a 10 to 15% uncertainty in the signals where isotherm depths
ranged from 90 to 105 m in the WCR. Comparisons of several sets of profiler measurements have suggested that the OHC scales as \(0.9\) to \(1\) kJ cm\(^{-2}\) m\(^{-1}\) in the LC and WCR structures, which suggests that if the \(26^\circ\)C isotherm depth is known, the OHC value scales with this empirical result.

![Figure 1.3.8: Regression analysis of a) OHC and b) \(26^\circ\)C isotherm depth (m) for pre-Rita (black circles) and post-Rita (red circles) between in situ (abscissa) and satellite-inferred (ordinate) for the data acquired in the WCR/CCR regime on 15 and 26 September 2005.](image)

1.3.2.3 Background Ocean States

Background oceanic flows that are set up by large horizontal pressure gradients due to \(T(z)\) and \(S(z)\) may play a significant role in altering the development of strong wind-driven current shears within the LC/WCR complex as suggested in Fig. 1.3.9 (Shay and Uhlhorn 2006). Pre- and post-Isidore measurements across the Yucatan Strait indicate strong density and pressure gradients that are associated with the northward-flowing LC at speeds of up to 1 m s\(^{-1}\). In the post-Isidore case, the horizontal gradients were sharpened since the storm cooled the Yucatan shelf waters by more than 4°C compared to less than 1°C across the Yucatan Strait. Here, strong horizontal advection of the thermal and salinity gradients through this regime impacted the oceanic response within the LC. Falkovich et al. (2005) introduced an approach for feature-based ocean modeling that involves cross-frontal “sharpening” of the background temperature and salinity fields according to data obtained in field experiments, which allows specifying the position of the LC in the GOM using available observations.

Briefly, the LC is a highly variable ocean feature in time as it can penetrate \(~500\) km northward of the Yucatan Strait. Recurring WCR shedding events with peak periods from 6 to 11 months (Sturges and Leben 2000) occur when CCRs are located on the LC.
periphery prior to separation. These WCRs, with diameters of ~200 km, propagate west to southwest at phase speeds of 3 to 5 km d\(^{-1}\) (Elliot 1982), and can remain in the Gulf of Mexico for several months. In this LC regime, OHC values relative to the 26°C isotherm exceed 100 kJ cm\(^{-2}\) (Leipper and Volgenau 1972). Such OHC levels are resistive to storm-induced cooling by wind-driven current shears across the OML base. Nof and Pichevin (2001) and Nof (2005) suggest that the LC cycle can be explained in terms of the momentum imbalance paradox theory. When a northward-propagating anomalous density current (Yucatan Current) flows into an open basin (GOM) with a coast on its right (Cuba), the outflow expands near its source to form a clockwise-rotating bulge (LC). The expansion of the current (WCR formation) is a necessary condition to satisfy the momentum flux balance along the northern coast of Cuba. Two-thirds of the outflow mass flux goes into this expanding bulge with the remaining flux forcing the downstream current. The subsequent WCR separation is due either to planetary vorticity gradients or topographic effects (Cherubin et al. 2005), where ~80% of the inflow forces the current and the remaining inflow goes into a WCR (Nof 2005). This theory predicts a mass partition that has been observed during WCR shedding events in North Atlantic Ocean simulations.

Figure 1.3.9: Pre-(upper) and post-Isidore (lower) thermal (°C: color) and northward (into the page) geostrophic velocity (m s\(^{-1}\): dashed) cross-section from expendables deployed on 19 and 23 September 2002 across the Yucatan Straits. Heavy dashed line represents the 26°C isotherm depth (from Shay and Uhlhorn 2006).

Data assimilative ocean nowcasts are an effective method for providing initial and boundary conditions to the ocean component of nested, coupled TC prediction models. The
thermal energy available to intensify and maintain a TC depends on both the temperature and thickness of the upper ocean warm layer. The ocean model must be initialized so that ocean features associated with relatively large or small ocean heat content (OHC) are in the correct locations and T-S (and density) profiles, along with the OHC, are realistic. Ocean nowcast-forecast systems based on HYCOM (Bleck 2002; Chassignet et al. 2003; Halliwell 2004) were evaluated in the northwest Caribbean and eastern Gulf of Mexico for September 2002 prior to Hurricanes Isidore and Lili, and for September 2004 prior to Hurricane Ivan. In this region, the OHC distribution associated with the LC/WCR complex as well as coastal upwelling must be accurately initialized for the ocean model. In this context, measurements are critically important not only for assimilation, but to evaluate initial and boundary ocean fields.

Figure 1.3.10: OHC (kJ cm\(^{-2}\)) in the northwest Caribbean Sea and southeast GOM from an objective analysis of aircraft observations, satellite altimetry, HYCOM NRL-CH nowcast, and HYCOM NRL-MODAS nowcast (four left panels). Temperature (right top) and salinity (right bottom) vertical profiles at a location in the northwest Caribbean Sea, where red lines are climatological profiles (GDEM3 dashed, WOA01 solid), solid blue lines are observed profiles, dashed blue lines are MODAS profiles, and black lines are model nowcasts (HYCOM-NRL dashed and HYCOM-MODAS solid).

An examination of the initial analysis prior to Isidore is from the experimental HYCOM nowcast-forecast system of the NRL experiments in the Atlantic basin at 0.08° resolution. This model product assimilates both satellite altimetry SHAs (Cooper and Haines 1996) and optimally interpolated SSTs. Comparison of OHC maps hindcast by HYCOM to OHC maps objectively analyzed from aircraft measurements and derived from satellite
observations (left panels of Fig. 1.3.10) demonstrate that this HYCOM analysis (labeled HYCOM NRL-CH in the figure) reproduces the LC orientation but underestimates OHC by ~50%. In the NW Caribbean Sea, the T(z) hindcast tends to follow the September climatology but does not reproduce the larger OHC values. In the HYCOM hindcast, the upper ocean is less saline than both climatology and observations above 250 m (Fig. 1.3.10) and less saline than the observations between 250 and 500 m. HYCOM structure was subsequently relaxed to the Navy three-dimensional MODAS (Fox et al. 2002) and T-S analyses were generated from all available in-situ observations. The first HYCOM NRL-CH nowcast was adversely impacted by a poor initialization that could not be corrected by including only the SHA fields. Biases were reduced in this HYCOM NRL-MODAS product compared to observations in both horizontal maps and vertical profiles. Evaluation of the next-generation NRL nowcast-forecast system (Cummins 2003) is being done by performing hindcasts from mid-2003 to the present for Ivan (2004) and Katrina (2005) and Rita (2005). Initial evaluation of pre-Ivan conditions is encouraging because the large cold bias was no longer present, and because the LC/WCR complex (includes the CCR) was well represented (see Figs. 1.3.2 and 1.3.6). Such evaluations of model-generated products are needed prior to coupling with a hurricane model to insure that ocean features are in the correct place and have structural characteristics that are realistic.

### 1.3.2.4 Vertical Mixing Parameterization

One of the significant effects on the upper-ocean heat budget and the fluxes to the atmosphere is the choice of entrainment mixing parameterizations. Jacob et al. (2006) have conducted sensitivity tests using five schemes: K-Profile Parameterization (KPP: Large et al. 1994); Goddard Institute for Space Studies Level-2 closure (GISS: Canuto et al. 2002); Level-2.5 turbulence closure scheme (MY: Mellor and Yamada, 1982); quasi-slab dynamical instability model (PWP: Price et al. 1986); and the turbulent balance model of Gaspar (KT: 1988) that is a modified version of Kraus and Turner (1967). As shown in Fig. 1.3.11 for quiescent ocean initial conditions, the range of fluxes in the directly forced region of Hurricane Gilbert exceeds 500 Wm$^{-2}$ for these five schemes.

For the Hurricane Gilbert case, Jacob and Shay (2003) simulated OML temperatures for realistic initial conditions and compared with profile observations to identify appropriate mixing schemes. The three higher-order turbulent mixing schemes (KPP, MY, GISS) considered will lead to a more accurate ocean response simulation. However, these comparisons are limited by data availability and therefore routine measurements are necessary to evaluate the ocean component of the coupled system. Similar to the post-season track and intensity verifications, more ocean observations must be acquired to evaluate the different schemes to build a larger statistical base. Given the large range in the simulated surface fluxes for different schemes, this is a crucial step toward reducing this uncertainty. The approach of stand-alone ocean simulations using derived realistic atmospheric forcing used here allows us to evaluate the ocean model and associated parameterizations. Since boundary layer forcing structure from the atmospheric component of the coupled model is subject to additional uncertainties, this approach will eventually lead to reduction in uncertainties of the ocean component in the coupled system based on observations.
The OML salinity evolution with and without precipitation forcing highlights the effect of precipitation on the upper ocean salt budget (not shown). For the no precipitation case, OML salinity variability simulated by all the above mixing schemes was similar ahead of the storm center. In the PWP scheme, however, salinity increases significantly in the right-rear quadrant over the first half of an inertial period due to enhanced mixing. While in the other mixing cases, the simulated salinity evolution is similar with minimal changes in the KT model due to less intense mixing (no vertical shear). By including precipitation forcing, the salinity in the OML began to decrease about 0.5 IP before the storm with maximum freshening of the OML observed in the KT case. This freshening process due to enhanced precipitation increases static stability in the mixed layer, leading to a simulated salinity balance for PWP case that is more consistent with the other schemes. OML temperature and salinity evolution in cases with and without precipitation for PWP scheme indicates a mean temperature and salinity differences of 0.5°C and 0.2 psu in the OML layer. An average freshening of 0.2 psu is seen in the wake of the storm in all the five cases when precipitation forcing was used, which is consistent with CTD measurements acquired during the Spectrum 90 expeditions (Pudov and Petrichenko 2000). Precipitation temperatures have minimal effect on the salinity evolution in the OML. Simulated results from the three-higher order schemes did not differ significantly from each other.

Figure 1.3.11: Evolution of OML quantities at a location 2 R_{max} right of the storm track for quiescent ocean initial conditions in the Gilbert case (Jacob and Shay 2003): a) Mixed layer temperature (°C), b) Mixed layer salinity (PSU), and c) Fluxes to the atmosphere. Note the wide range of variability in the OML temperature and in the fluxes to the atmosphere for the five mixing parameterizations (see inset in panel (c)).
1.3.3: Air-Sea Interface:

Because the underlying ocean significantly affects TC intensity, attention has been drawn toward gaining a better understanding of the physical interaction between the atmosphere and ocean during these events. Unfortunately, due to limited observations at the air-sea interface in high-wind conditions, the understanding has not progressed nearly enough to significantly improve the parameterization of momentum and energy transfer. The relationships of the transfer processes to small-scale roughness (Charnock 1955) and surface-layer stability (Monin-Obukhov similarity theory) are fairly well understood under low- to moderate-wind conditions (Large and Pond 1981), but additional phenomena not typically observed such as the maturity of the sea state (Donelan 1990) and sea spray (Fairall et al. 1994; Wang et al. 2001) have also been shown to modulate the heat and momentum exchange. These effects under TC-force winds have been primarily studied in controlled laboratory experiments (Donelan et al. 2004). In a TC environment, both young and mature waves are present and impact the air-sea fluxes and OML and ABL processes.

The TC intensity is maintained in part by the balance between the heat gained by the boundary layer of the storm and the energy lost due to friction. Emanuel (1986) proposed a theory requiring this relative balance to be the primary modulator of intensity. Based on this view, it is assumed that under certain conditions there should be a level of mutual dependence of the air-sea transfer processes of heat and momentum. Indeed, it has been suggested through highly idealized model simulations (Ooyama 1969; Rosenthal 1971; Emanuel 1995; Braun and Tao 2000) that the TC intensity is sensitive to the ratio of enthalpy transfer coefficient to drag coefficient ($C_h C_d^{-1}$). The conclusion that this quantity probably lies within a rather limited range (<1.5), is consistent with the observation that most TCs do not reach their maximum potential intensity (DeMaria and Kaplan 1993).

1.3.3.1 Surface Wave Field

On moored NOAA buoys, wave spectral energy is derived from accelerometers or inclinometers that measure the heave accelerations, or vertical displacements, of moored pitch-roll buoys that use the new Multi-functional Acquisition and Reporting System (MARS) payload system. When Hurricane Emily approached buoy 44014 from the southeast with sustained winds of 28 m s$^{-1}$, the significant wave heights reached 8 m (Fig. 1.3.12a). Maximum wave spectral energies were largely contained in the swell portion of the spectrum (i.e., ~13 s wave) that decayed rapidly after Emily’s passage. These wave spectral energies contained in the lower-frequency intervals associated with the swell began to increase several hours in advance of storm passage, peaked at the point of closest approach, and subsequently decayed over 1-2 inertial periods (IP~20 h) in the wake. In frequency bands between 0.2-0.4 Hz, the modulation of the wave spectral energies continued for an extended period of time (Faber et al. 1997). Wave spectral energies and significant wave heights indicate peaks that occur over IP time scales.

More recently, Hurricane Lili approached buoy 42001 from a south-southwest direction after crossing the western tip of Cuba, just as she reached her maximum intensity (Cat- 4). As Lili passed within $R_{max}$ to the west of the buoy, winds reached 48 m s$^{-1}$, significant
wave heights peaked above 10 m, and wave spectral energies exceeded 220 m² Hz⁻¹ (Fig. 1.3.12b). This is the region of the storm where the maximum ocean response is observed compared to the other quadrants. The largest values of wave spectral energy are again concentrated in the lower frequency band with maximum values at 0.4 IP (~10 h) prior to passage and persist for about 1.5 IP (~40 h) after the closest

Figure 1.3.12 : Time series of a) wind speed (solid) and significant wave height (dotted), and b) contoured wave spectral energy (0.1 to 1 m² Hz⁻¹) from buoys 44014 and 42001 during and subsequent to tropical cyclones Emily (left panels) and Lili (right panels). Time series are scaled by local inertial periods of ~20 hrs for buoy 44014 and ~27 hrs for 42001.

approach of Lili. Smaller amplitudes of the wave spectral energy in the higher frequency (0.2-0.4 Hz) intervals are evident 1.7 IP (~48 h) prior to passage, and persist after Lili’s passage over the buoy. The phase of the oscillations, which is most pronounced between the frequency intervals 0.2-0.4 Hz, is near the local IP (~27 h). Such observations should be used to test coupled ocean-wave models to assess their performance under strong wind and wave conditions.

Wang et al. (2005) documented the wave response to Ivan over the NRL SEED moorings. Wave heights significantly increased with peak values when the radial distance between the mooring and storm center was ~75 km (Fig. 1.3.13c). Hₚ reached maximum values of 16 m to 18 m on moorings 3, 4, and 5 and were larger than those observed at the NDBC
buoy (15.9 m). The maximum wave height was recorded to be 27.7 m at mooring 3, and wave height variations were consistent with the radial variations in the surface wind of Ivan. At $R_{\text{max}}$, the model predicted a maximum wave height of ~21 m (Fig. 1.3.13c). Previous studies have suggested in a hurricane wave field that the maximum wave height approaches $1.9 \times H_s$, which would be consistent with these measurements. The moored

Figure 1.3.13: a) Hurricane Ivan satellite image at 1850 UTC 15 September 2004 in the GOM with the green line representing the track of Ivan at 3-h intervals moving over the SEED moorings (blue). Panel b) represents the evolution of $H_s$ (circles) and $H_{\text{max}}$ (crosses) as a function of normalized distance relative to $R_{\text{max}}$. $H_s$ is from NDBC buoy 42040 (dotted curve) and its radial distance to Ivan’s center is shown by the green squares. Panel c) represents $H_s$ and $H_{\text{max}}$ as a function of normalized distance from the center compared to the exponential distance: digitized values of a segment $15^\circ$ CW from the forward direction of a numerically simulated wave field (black asterisks). Blue curve depicts the line of $H_{\text{max}}=1.9H_s$ where circles and crosses are as in panel b) (from Wang et al. 2005).

measurements sampled only a small part of the domain influenced by Ivan’s broad wind field. SRA measurements (Wright et al. 2001) from the aircraft should be used to extend
the mooring measurements to a scale commensurate with Ivan’s wind field as suggested in Fig. 1.3.2.

1.3.3.2 Surface Winds

Surface winds in hurricanes have been estimated remotely using the Stepped-Frequency Microwave Radiometer (SFMR) from aircraft (Uhlhorn and Black 2003). With six frequencies, the SFMR measures radiative emissions, expressed in terms of brightness temperatures ($T_b$), from the ocean and the atmosphere. The percentage of foam coverage on the sea surface is known to increase monotonically with wind speed. At microwave frequencies, foam acts as a blackbody emitter. As foam increases, the ocean emits microwave energy more readily, and assuming a constant SST, the $T_b$ increases. Given an accurate physical model that relates ocean surface wind speed and rain-rate to measurements of $T_b$ at frequencies, a set of equations may, in theory, be inverted to calculate the surface winds. Uhlhorn et al. (2006) has assessed SFMR measurements on aircraft for operational surface wind speed measurements in the active 2005 hurricane season, which provided ample data to evaluate both instrument performance and surface wind speed retrieval quality up to Cat-5 hurricanes. A new microwave emissivity and wind speed model function based on comparisons with direct measurements of surface winds in hurricanes by GPS dropwindsondes is shown in Fig. 1.3.14. This function eliminates a previously-documented high bias in moderate SFMR-measured wind speeds (10-50 m s$^{-1}$), and additionally corrects an extreme wind speed (>60 m s$^{-1}$) systematic underestimate in the past cases. The model function behaves differently below and above the hurricane wind speed threshold (32 m s$^{-1}$), which may have implications for air-sea momentum and kinetic energy exchange. The change in behavior is at least qualitatively consistent with recent laboratory and field results concerning the drag coefficient ($c_d$) in high wind speed conditions, which show a fairly clear “leveling-off” of $c_d$ with increased wind speed above ~30 m s$^{-1}$ as discussed below.

1.3.3.3 Surface Drag Coefficients

Since the energy source for the TC is the ocean, knowledge of the heat and moisture fluxes across the interface and into the ABL are critical elements. However, the exchange coefficients for heat, moisture, and momentum are not well known for the high wind speed and ocean surface wave conditions. Momentum transfer between the two fluids is characterized by the variations of wind with height and a surface drag coefficient that is a function of wind speed and surface roughness. However, it is difficult to acquire flux measurements for the high wind and wave conditions under the eyewall. Since 1997, GPS sondes have been deployed from aircraft to measure the Lagrangian wind profiles in the atmospheric boundary layer in TCs. Powell et al. (2003) found a logarithmic variation of mean wind speed in the lowest 200 m, a maximum speed at 500 m, and a gradual weakening with height to 3 km. From these estimates, the surface stress, roughness length, and neutral stability drag coefficient determined by the profile method suggest a leveling of the surface momentum flux as winds increase above hurricane-force and a slight decrease of the drag coefficient with increasing winds.
Figure 1.3.14: Excess emissivity from SFMR compared to 10-m (surface) winds measured from GPS dropsondes during the 2005 season. The total number of samples is 160 and the rms difference between the SFMR model function was 0.011 (from Uhlhorn et al. 2006).

Donelan et al. (2004) describe a series of wind-wave tank experiments that included stress measurements from hot-film anemometry and digital particle image velocimetry, and laser/line scan cameras for measuring the water surface elevation. Reynolds stress was measured directly with an x-film anemometer at low and moderate (centerline) wind speeds (0 to 26 m s\(^{-1}\)). The stress determined at the measured elevations was used to correct the values at the surface with the measured horizontal pressure gradient in the tank. At higher winds, surface stress was determined from a momentum budget of sections of the tank. The steady-state surface stress increases the momentum of the wave field with increasing fetch, which drives a downwind current near the surface and maintains a downwind slope of the mean surface (mean surface elevation increasing in the downwind direction). The horizontal pressure gradient drives a return flow (upwind flow) in the bottom of the water column, which causes a drag on the bottom of the tank. Finally, the horizontal pressure gradient in the air that produces the wind adds to the slope of the water surface - the "inverted barometer" effect.

Measurements of the drag coefficient from this laboratory experiment, referenced to the 10-m wind speed, are shown in Fig. 1.3.15. Wind speed was measured at 30 cm height in the tank and extrapolated to the 10-m using the logarithmic dependence on height and was verified between crest height and 30 cm for all but the two highest wind speeds. Three other data sets were obtained in the wind-wave tank using the profile method (in which the vertical gradient of mean horizontal velocity is related to the surface stress), the Reynolds stress method, and the momentum budget or "surface slope" method. The agreement among the various methods validates the momentum budget method which, being insensitive to airborne droplets, allows a measurement of the surface stress at the highest
winds generated. Notice the characteristic behaviour of the drag coefficient as the surface conditions changes from aerodynamically smooth (characterized by a decrease in the drag coefficient at low-wind speeds) to aerodynamically rough (drag coefficient increasing with wind speed) conditions. In rough flow, the drag coefficient is related to height of the “roughness elements” per unit distance downwind or, more precisely, the spatial average of the downwind slopes. Unlike a solid surface, the roughness elements (or waves) are responsive to the wind so that the drag coefficient increases between 3 and 33 m s\(^{-1}\).

Figure 1.3.15: Laboratory measurements of the neutral stability drag coefficient (\(x 10^{-3}\)) by profile, eddy correlation (“Reynolds”), and momentum budget methods. The drag coefficient refers to the wind speed measured at the standard anemometer height of 10 m. The drag coefficient formula of Large and Pond (1981) is also shown along with values from Ocampo-Torres et al (1994) derived from field measurements (from Fig. 2 in Donelan et al. 2004).

In a TC, the wind changes direction and speed over relatively short distances compared to those required to approach full wave development. Consequently, the largest waves in the wind-sea move relatively slowly compared to the wind and often travel in directions differing from the wind. Under such circumstances, these long waves contribute to the aerodynamic roughness of the sea as hypothesized by Kitaigorodskii (1968) and demonstrated by Donelan (1990). Measurements at sea (e.g., Large and Pond, 1981) and in laboratories (e.g., Donelan 1990; Ocampo-Torres et al. 1994) amply demonstrate the increasing aerodynamic roughness with increasing wind speed. A “saturation” of the drag coefficient does appear once the wind speed exceeds 33 m s\(^{-1}\) (Fig. 1.3.15). Beyond this speed, the ocean surface simply does not become any rougher in an aerodynamic sense. At the highest wind speed, the significant height and peak frequency of the waves in the laboratory were 9 cm and 1.4 Hz. In the range of wind speeds of 10 to 26 m s\(^{-1}\), the laboratory measurements parallel the open-ocean measurements (Large and Pond 1981), but are lower. These measurements suggest aerodynamic roughness saturation beyond 10
m height wind speeds of 33 m s$^{-1}$. The saturation level for the drag coefficient is 0.0025, which corresponds to a roughness length of 3.35 mm. Powell et al. (2003) show “saturation” of the drag coefficient at 0.0026 at about 35 m s$^{-1}$. Shay and Jacob (2006) also find a “saturation” at about 30 m s$^{-1}$, but their analytical function, derived by equating internal wave fluxes to the surface stress, revealed a value 0.0034 before this tapering-off trend. The possibility of a limiting state in the aerodynamic roughness of the sea surface is of critical importance in understanding and modelling the development of hurricanes and other intense storms. Donelan et al. suggest a change in flow characteristics leading to saturated aero-dynamic roughness at boundary layer wind speeds in excess of 33 m s$^{-1}$. Obviously, more research is needed to quantify the surface drag coefficient given its importance for calculating the enthalpy fluxes (Emanuel 2003).

1.3.3.4 Wind-Wave Coupling

![Figure 1.3.16: Drag coefficients ($C_d$) from various observation-based values, empirical formulas, and model outputs as a function of $U_{10}$. Symbols represent observations from GPS sonde wind profiles (Powell et al. 2003). Vertical bars represent 95% confidence limits. Solid line is an extrapolation of the Large and Pond (1981) formula. Dash-dot line is the bulk formula used in GFDL hurricane model. Shaded and hatched areas represent ranges between upper and lower bound of $C_d$ obtained by the URI coupled wave-wind model and an internal estimation of WAVEWATCH III, respectively (from Moon et al. 2004).]

In the GFDL model, the momentum flux is parameterized with a constant non-dimensional surface roughness (or Charnock coefficient) and the stability correction is based on the Monin-Obukhov similarity theory, regardless of the wind speed or the sea state. In addition to wind speed, Donelan (1990) found $C_d$ also depends on the sea state represented by the wave age. Moon et al. (2004) investigated the Charnock coefficient under TC conditions using a coupled wind-wave (CWW) model. In the CWW model, the surface wave directional frequency spectrum near the spectral peak is calculated using the WAVEWATCH III (Tolman 2002) model and the high frequency part of the spectrum was parameterized using the theoretical model of Hara and Belcher (2002). The wave spectrum is then introduced to the wave boundary layer model of Hara and Belcher (2004) to estimate the Charnock coefficient at differing wave evolution stages. They found that $C_d$ (Fig. 1.3.16) levels off at high wind speeds, which is consistent with the above studies cited in Fig. 1.3.14. The most important finding of this study is that the relationship
between the Charnock coefficient and the input wave age (wave age determined by the peak frequency of wind energy input) is not unique, but strongly depends on wind speed. The regression lines between the input wave age and the Charnock coefficient have a negative slope at low wind speeds but have a positive slope at high wind speeds. This behavior of the Charnock coefficient provides one explanation why \( C_d \) under a TC, where seas tend to be “young,” is reduced in high wind speeds.

In the presence of surface waves, the momentum flux from the atmosphere may be different from the flux into the ocean if the wave field is spatially/temporally varying. That is, the momentum transfer from the atmosphere consists of two components: momentum transferred directly into the ocean current (by viscosity) and momentum transferred to the waves (through the wave-induced stress). The total momentum transfer into ocean then consists of two components: momentum transferred directly from the atmosphere (by viscosity) and momentum transferred from the waves (through the wave breaking). The difference between the fluxes into and out of the waves may be particularly significant under hurricane conditions since the wave field is far from fully-developed. This difference can reach up to 20% in the vicinity of \( R_{\text{max}} \), which may potentially have an important impact in the coupled hurricane-wave-ocean model.

### 1.3.3.5 Enthalpy Fluxes

Fluxes of heat and moisture are central to the TC intensity question and are usually determined from bulk formulae that utilize near-surface atmospheric thermodynamic and wind observations and upper-ocean temperature data measured by expendable probes. Specifically, the atmospheric variables are estimated from the large number of GPS sondes (Hock and Franklin 1998) deployed within the storm from both NOAA aircraft as well as Air Force Reserve reconnaissance flights. Surface winds are routinely measured by the SFMR (Uhlhorn et al. 2006). The GPS sondes measure profiles of temperature, pressure and humidity from flight level to the surface. For each measurement, 10-m values of these quantities are optimally interpolated to a storm-relative grid aligned with the direction of storm motion. The SFMR wind observations are objectively analyzed (Powell and Houston 1996) and interpolated to the same grid as the GPS thermodynamic data. Finally, the SSTs observed by expendable ocean probes on in-storm flights are interpolated to the same grid as the GPS measurements as in Isidore and Lili (Shay and Uhlhorn 2006). Key items in the estimates for heat and moisture fluxes are temperature and specific humidity differences and the bulk transfer coefficient (i.e., ratio of the enthalpy exchange coefficient to surface drag coefficient (discussed above). As most of the recent studies have indicated a leveling off of the surface drag coefficient at 30 m s\(^{-1}\), Emanuel (2003) suggests that this ratio becomes independent of wind speed and that the ratio should be \( O(1) \), since below this value intense hurricanes cannot be simulated in the numerical model. Heat exchange coefficients are typically set equal to \( C_d \), which is conservative based on these recent theories. An additional ocean forcing mechanism is the surface precipitation flux (rain rate). As noted above, freshwater input by rain can alter the ocean response both by direct cooling due to rain that has a lower temperature than the SST, and by stabilizing the OML by decreasing the salinity and reducing the mixing rate (Price et al. 1986; Jacob et al. 2006).
Estimates of enthalpy fluxes during Isidore and Lili were sensitive to the storm translation speed. In Hurricane Isidore, the peak enthalpy flux $\sim 1.8 \text{ kW m}^{-2}$ is in the right-rear quadrant of the storm due to the high SSTs ($\sim 30^\circ \text{C}$) due to a negligible decrease from pre-storm conditions, especially over the warm LC where ocean cooling was minimal. Although the maximum momentum flux (7 Pa) is in the right-front quadrant, Isidore's wind stress field was highly symmetric as it moved at only 4 m s$^{-1}$. Estimated maximum surface fluxes in Lili were about $1.4 \text{ kW m}^{-2}$ due in part to the marked asymmetry associated with the faster storm translation speed (7 m s$^{-1}$) and smaller SSTs by about 1$^\circ \text{C}$. This result highlights how even modest SST differences can effectively alter the surface heat flux in extreme winds.

Enthalpy fluxes may also be integrated along the track to obtain the cross-track (radial) distributions of net sea surface heat loss. Using the space-to-time conversion in the along-track direction, estimated surface heat losses for Isidore and Lili are shown in Fig. 1.3.17. The surface heat loss in Isidore ($\sim 9 \text{ kJ cm}^{-2}$) is almost a factor of two larger than in Lili ($\sim 4.5 \text{ kJ cm}^{-2}$) due to the enhanced flux, slower storm speed, and larger horizontal SST gradients along the western side of the Yucatan Strait. Fluxes are responding to the changes in the upper ocean mixed layer budget and OHC during both storms that are moving over a strong current system. Cione and Uhlhorn (2003) argue that it is only inner-core SSTs that the storm responds to if the OHC was held constant. However, OHC is not remaining constant underneath a storm if SSTs are changing by 1$^\circ$ to 2$^\circ \text{C}$. SST (or near-surface temperatures from profilers) represents the surface boundary condition for OHC estimates from profilers and satellite-based retrieval algorithms, and the SST decrease is not only a function of surface heat flux, but also the stress- and shear-induced mixing (as illustrated in Fig. 1.3.1)
1.3.3.6 Sea Spray

Over the last decade, Fairall et al. (1994) and Kepert et al. (1999) have been developing a hierarchy of models of the sea spray production at high winds and the subsequent thermodynamic effects of the evaporation of spray on hurricane boundary layers. The three steps in this process are: i) characterization of the size spectrum of droplets produced by the ocean as a function of the forcing (wind speed, stress, wave breaking, etc); ii) computation of the exchanges of heat and moisture between the droplets and an unperturbed near-surface layer structure; and iii) accounting for the ‘subgrid-scale’ distortion of the standard surface layer temperature and relative humidity structure by the droplets (a process referred to as ‘feedback’). The present source strength parameterization is derived from the Fairall-Banner physical sea spray model that predicts the size spectrum of sea spray produced by the ocean in terms of wind speed, surface stress, and wave properties.

By extending Banner et al. (2000) approach, the The Fairall-Banner spectrum has been parameterized into a simple mass flux representation in terms of surface friction velocity. The unperturbed thermodynamic effects are based on integrals of the ratios of thermodynamic and suspension time constants following Andreas. Finally, a diagnostic feedback parameterization has been developed that characterizes how evaporating droplets of various sizes modify the stratification of the air near the surface, which in turn reduces further droplet evaporation but enhances sensible heat flux carried by the droplets. The present form of the parameterization has two tuning coefficients: one that scales the intensity of the source strength, and the other that affects the partitioning of enthalpy flux between sensible and latent heat. As shown in Fig. 1.3.18, the ratios of momentum of enthalpy transfer coefficients are scaled with wind speed for differing choices of the source strength. Recently, this parameterization was implemented in the GFDL model and a version of WRF model. Preliminary tests with Hurricanes Ivan and Isabel showed sensitivity to the sea spray parameterization, but there are dependencies with the non-droplet (direct) transfer specifications in the models. More testing is needed to examine the sensitivity of these results and determine the role of sea spray on the enthalpy fluxes and hence storm intensification.

Figure 1.3.18: The ratio of momentum ($c_d$) to enthalpy ($c_p$) bulk transfer coefficients when the effects of sea spray are included via different specifications of the droplet source strength (figure courtesy of Chris Fairall).

By extending Banner et al. (2000) approach, the The Fairall-Banner spectrum has been parameterized into a simple mass flux representation in terms of surface friction velocity. The unperturbed thermodynamic effects are based on integrals of the ratios of thermodynamic and suspension time constants following Andreas. Finally, a diagnostic feedback parameterization has been developed that characterizes how evaporating droplets of various sizes modify the stratification of the air near the surface, which in turn reduces further droplet evaporation but enhances sensible heat flux carried by the droplets. The present form of the parameterization has two tuning coefficients: one that scales the intensity of the source strength, and the other that affects the partitioning of enthalpy flux between sensible and latent heat. As shown in Fig. 1.3.18, the ratios of momentum of enthalpy transfer coefficients are scaled with wind speed for differing choices of the source strength. Recently, this parameterization was implemented in the GFDL model and a version of WRF model. Preliminary tests with Hurricanes Ivan and Isabel showed sensitivity to the sea spray parameterization, but there are dependencies with the non-droplet (direct) transfer specifications in the models. More testing is needed to examine the sensitivity of these results and determine the role of sea spray on the enthalpy fluxes and hence storm intensification.
1.3.4 Summary

Significant progress has been made in understanding the basic oceanic and atmospheric processes that occur during TC passage. The need is to isolate fundamental physical processes involved in the interactions through detailed process studies using experimental, empirical, theoretical, and numerical approaches. As demonstrated from new measurements, these approaches are needed to improve predictions of tropical cyclone intensity and structure.

Considerable ocean-atmosphere variability occurs over the storm scales that include fundamental length scales such as the radius of maximum winds and another scale that includes the radius to gale-force winds. Here, the fundamental science questions are how the two fluids are coupled through OML and ABL processes, and what are the fundamental time scales of this interaction? These questions are not easily answered as the interactions must be occurring over various time/space scales. For example, one school of thought is that the only important process with respect to the ocean is under the eyewall where ocean cooling has occurred. While it is at the eyewall where the maximum wind and enthalpy fluxes occur, the broad surface circulation over the warm OML also has non-zero fluxes that are contributing thermal energy to the TC. The deeper the OML (and 26°C isotherm depth), more heat (OHC) is available to the storm through the enthalpy fluxes. Notwithstanding, it is not just the magnitude of the OHC, since the depth of the warm water is important to sustaining surface enthalpy fluxes. Process studies need to begin to look at these multiple scale aspects associated with the atmospheric response to ocean forcing.

With regard to the oceanic response to the atmospheric forcing, an important missing ingredient in many studies is the role of the forced and background current fields. In addition to aircraft-based sampling by AXCPs and AXCTDs and new profiling floats such as the EMAPEX and the SOLO (during CBLAST), efforts along the southeastern United States are underway to deploy long-range, HF-radars to map the surface currents to 200 km from the coast as part of an integrated ocean observing system (Shay et al. 2006a). Such measurements would not only be invaluable to map the wind-driven surface currents during high winds, but also in the case of phased arrays, to map the directional wave spectra over the domain. These measurements could then be used to not only examine air-sea interactions, but also assess the relative importance of surface wave-current interactions under strong wind conditions in an Eulerian frame of reference.

The variability of the surface drag coefficient has received considerable attention over the last five years including the Coupled Boundary Layer an Air-Sea Transfer (CBLAST) program sponsored by the Office of Naval Research. Several treatments have come to the conclusion that there is a leveling off or a saturation values at about 30 m s⁻¹ +/- 3 m s⁻¹. The ratio of the enthalpy coefficient and the drag coefficient is central to air sea fluxes impacting the TC boundary layer. In this context, the relationship between the coupled processes such as wave breaking and the generation of sea spray and how this is linked to localized air-sea fluxes remains a fertile research area. A key element of this topic is the
atmospheric response to the oceanic forcing where there seems to be contrasting viewpoints. One argument is that the air-sea interactions are occurring over surface wave (wind-wave) time and space scales and cause significant intensity changes by more than a category due to very large surface drag coefficients. Yet empirical studies suggest the values to be between $2.5 \times 10^{-3}$ compared to recent coupled model studies. While, these sub-mesoscale phenomena may affect the enthalpy fluxes, the first-order balances are primarily between the atmospheric and oceanic mixed layers.

1.3.5 Recommendations

Based on an Air-Sea Interaction Workshop at NCEP during May 2005 (see http://iwave.rsmas.miami.edu/~nick/AS_HWRF_wksp_rev.pdf), the recommendations are:

i.) Given the range of uncertainty in the surface drag (e.g., wave effects), heat fluxes (e.g., sea spray), and initial conditions (e.g., wind field) beyond $30 \text{ m s}^{-1}$ using CBLAST data, assess how these combined uncertainties propagate through the coupled ocean-hurricane model;

ii.) Develop an archive of data sets and model outputs and make these archives publicly available for research and operational purposes. Investigate the potential use of these data sets in assimilation, evaluation, and verification of models (e.g., HYCOM) and parameterization schemes;

iii.) Create an in-situ tropical cyclone ocean-atmosphere observing program for pre-storm, storm, and post-storm environments. Develop optimal observing strategies and observational mixes for spatial evolution of upper ocean, interface, and atmospheric fields (including secondary circulations such as roll vortices in the hurricane boundary layer); and;

iv.) Develop improved ocean model initialization schemes through data assimilation of satellite and in situ measurements, and test mixing parameterizations for a spectrum of ocean, wave and atmospheric conditions including the impact of waves on the surface heat, moisture, and momentum fluxes, and thus on the evolving OML.

A common theme in these discussions was that national and international research and forecasting programs need to build on recent field programs (ONR-CBLAST, NSF-sponsored measurements on Hurricane Isidore and Lili) and acquire in situ data over a broad spectrum of atmospheric and oceanographic forcing regimes. These data are needed to test models and examine the parameter space in these air-sea interaction and vertical mixing schemes. Ultimately, future research initiatives must now have strong experimental, empirical, analytical, and numerical modeling components to further our understanding of these fairly complex coupling processes between these two fluids.

Acknowledgments: L. K. Shay gratefully acknowledges support from the NSF through grants ATM-01-08218, 04-44525 and NOAA Joint Hurricane Testbed program. We are grateful for the efforts of the pilots, technicians, engineers and scientists at the NOAA
Aircraft Operation Center (Dr. Jim McFadden) who make it possible to acquire high
good quality data during hurricanes. Mr. Bill Teague (NRL-Stennis) and Dr. Rick Lumpkin
(NOAA-AOML) kindly shared data and Drs. George Halliwell, S. Daniel Jacob, Tom
Sanford and Chris Fairall shared preprint material used herein. Dr. Russell Elsberry made
editorial comments that were incorporated by Mrs. Penny Jones at the Naval Postgraduate
School. Finally, I appreciate the assistance and guidance of Dr. Hugh Willoughby in
preparing this document.

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