Symmetric and Asymmetric Structures of Hurricane Boundary Layer in Coupled Atmosphere–Wave–Ocean Models and Observations

CHIA-YING LEE AND SHUYI S. CHEN

Rosenstiel School of Marine and Atmospheric Science, University of Miami, Miami, Florida

(Manuscript received 9 February 2012, in final form 5 June 2012)

ABSTRACT

It is widely accepted that air-sea interaction is one of the key factors in controlling tropical cyclone (TC) intensity. However, the physical mechanisms for connecting the upper ocean and air-sea interface with storm structure through the atmospheric boundary layer in TCs are not well understood. This study investigates the air-sea coupling processes using a fully coupled atmosphere-wave-ocean model, especially the couplinginduced asymmetry in surface winds, sea surface temperature, air-sea fluxes, and their impacts on the structure of the hurricane boundary layer (HBL). Numerical experiments of Hurricane Frances (2004) with and without coupling to an ocean model and/or a surface wave model are used to examine the impacts of the ocean and wave coupling, respectively. Model results are compared with the airborne dropsonde and surface wind measurements on board the NOAA WP-3D aircraft. The atmosphere-ocean coupling reduces the mixed-layer depth in the rear-right quadrant due to storm-induced ocean cooling, whereas the wind-wave coupling enhances boundary inflow outside the radius of maximum wind. Storm motion and deep tropospheric inflow create a significant front-to-back asymmetry in the depth of the inflow layer. These results are consistent with the dropsonde observations. The azimuthally averaged inflow layer and the mixed layer, as documented in previous studies, are not representative of the asymmetric HBL. The complex, three-dimensional asymmetric structure in both thermodynamic and dynamic properties of the HBL indicates that it would be difficult to parameterize the effects of air-sea coupling without a fully coupled model.

1. Introduction

The ocean plays an important role in tropical cyclone (TC) intensity as many studies have shown previously (e.g., Emanuel 1986; Rotunno and Emanuel 1987). Most studies have focused on the net effect of the ocean on TC intensity (e.g., Schade and Emanuel 1999; Bender and Ginis 2000; Wu et al. 2007), but they do not explicitly address the impacts of air–sea coupling on storm structure, that is, horizontal and vertical distributions of wind, rain, pressure, and temperature fields that affect the TC intensity. Deep convection in the eyewall and rainbands is connected to the ocean surface through the hurricane boundary layer (HBL), which is difficult to observe in high wind conditions. Recent advancements in high-resolution, fully coupled atmosphere–wave–ocean

DOI: 10.1175/JAS-D-12-046.1

modeling (e.g., Bao et al. 2000; Chen et al. 2007) and coupled observations from the Coupled Boundary Layer Air–Sea Transfer (CBLAST)-Hurricane field program in 2003–04 (e.g., Black et al. 2007) provide an opportunity to address this important problem in TC intensity.

The fundamental concept of the atmospheric boundary layer (ABL) is a layer that is directly influenced by the presence of the earth's surface (Stull 1988). In general, the definition of ABL varies with applications. The definition of the HBL has been a complex issue, and several definitions can be found in the literature. The first commonly used definition is based on the ABL definition in which the height of the HBL is a level at which the virtual potential temperature θ_v is 0.5 K higher than the surface value (Anthes and Chang 1978; Powell 1990). It represents a well-mixed layer and a transition layer aloft in which the θ_v increases quickly with height. The depth of the mixed layer is controlled by the surface heat fluxes, the entrainment from the top of the mixed layer, and the tropospheric subsidence. Observational

Corresponding author address: Dr. Shuyi S. Chen, RSMAS, University of Miami, 4600 Rickenbacker Causeway, Miami, FL 33149.

E-mail: schen@rsmas.miami.edu

studies have shown that the mixed-layer depth increases from the inner core to the outer region of hurricanes (e.g., Zhang et al. 2011). The HBL definition, based on the mixed-layer, is concerned with the thermodynamic property of HBL, which will be referred to as thermodynamic HBL (THBL) in this study.

The second definition is based on an inflow layer associated with the secondary circulation in hurricanes. The boundary inflow is a result of gradient wind imbalance due to the surface friction (Smith 1968), and the top of the inflow layer is defined as where the inflow vanishes (e.g., Smith et al. 2009; Zhang et al. 2011). This definition has an implicit assumption that the inflow in hurricanes is only induced by the surface friction. However, in reality, the inflow is caused by both surface friction and latent heat release from deep convection in the inner core and rainbands of a hurricane. Although the former dominates near the surface, the latter contributes throughout the lower troposphere (e.g., Pendergrass and Willoughby 2009). Therefore, it is ambiguous when using the inflow layer as a definition of HBL. Nevertheless, to distinguish the inflow layer from that of the mixed layer, here we refer to it as dynamic HBL (DHBL). A number of numerical and observational studies have shown the characteristics of the inflow layer. Using a composite of dropsondes from multiple hurricanes, Zhang et al. (2011) shows how the azimuthally averaged height of the DHBL increases outward from a few hundred meters in the inner core to about 1.5 km in the outer region, similar to an idealized hurricanelike vortex study by Montgomery et al. (2001). Just above the DHBL, there is a layer of outflow due to the upward transport of angular momentum, as explained in Kepert and Wang (2001). However, the studies of Kepert (2006a,b) and Schwendike and Kepert (2008) using dropsonde analysis from several hurricanes show a significant asymmetry in winds and the depth of DHBL. Their results indicate a large variability in the depth of inflow around each storm. Many have a much deeper inflow layer in parts of the hurricanes than the composite in Zhang et al. (2011), which raises a question of whether the composite inflow can represent the true structure in hurricanes.

The third definition is related to the Ekman layer. The Ekman layer depth $\sqrt{2K/f}$ is proportional to the square root of turbulent diffusivity K and inversely proportional to the square root of the Coriolis parameter f. However, in hurricanes, the relative vorticity is high and, therefore, the inertial and curvature effects have to be considered in the definition. Rosenthal (1962) was the first to use this definition in hurricanes and found that the Ekman-like layer decreases in depth toward the center of a hurricane. Kepert (2001) and Foster (2009)

show that the scale of the HBL height is given by an Ekman-like scale with the Coriolis parameter replaced by inertial stability I (i.e., $\sqrt{2K/I}$). Here we refer to this definition as IHBL. Maybe because K is usually unknown or has a high uncertainty, there is no direct observational study on IHBL structure. However, Kepert and Wang (2001) showed that the depth of IHBL decreases toward the center of a hurricane due to the increase of inertial stability, which is consistent with the radial variation in HBL height as determined by other observed quantities.

It is obvious that no matter which definition is used, the HBL is always associated with the amount of inward and upward transport of net heat, moisture, and momentum fluxes (Emanuel 1986). The HBL depth is therefore important since the net flux is the integral of flux divergence through the HBL, that is, the flux difference between the top and bottom of the HBL. There may not be a single value or an one-dimensional (1D) radial profile that can adequately describe the threedimensional (3D) HBL structure.

Factors controlling the symmetric and asymmetric structures of HBL are complex and diverse. For instance, air-sea interaction and deep convection in real hurricanes can both contribute to the variability in the HBL. However, they have not been systematically studied in this context. One of the unique features of airsea interaction in TCs is the storm-induced cold wake, which is strongest in the rear-right quadrant (Price 1981; Price et al. 1994). By using an axisymmetric coupled atmosphere-ocean model, Anthes and Chang (1978) first documented the influence of air-sea interaction on the THBL. They showed that the height of the THBL begins to change as soon as the sea surface temperature (SST) in the model is modified, especially at large radii, where the wind-driven mechanical mixing and vertical motion are weak and the height of THBL is mainly controlled by the ocean temperature. However, the effect of the observed asymmetry in hurricane-induced cold wake on the THBL was not possible to address in the axisymmetric model. Powell (1990) examined characteristics of the THBL in rainband regions using aircraft data. He showed that the convective downdraft can modulate the THBL by bringing down the dry air into the THBL, which varies spatially around the storm.

Storm motion can induce asymmetries in the nearsurface winds and DHBL by an enhanced convergence in the front of a hurricane as shown in a depth-averaged slab model by Shapiro (1983). Kepert (2001) and Kepert and Wang (2001) later confirmed this result using both a linear analytical model and a nonlinear multilayer 3D numerical model, respectively. They also found an appreciable asymmetric component at the eyewall and outer regions in TCs. In particular, the storm motion– induced asymmetric component dominates and resulted in a wavenumber-1 asymmetry with the maximum storm-relative inflow in the front-right quadrant and the tangential wind in the front-left quadrant.

Using the GPS dropsondes from Hurricanes Georges (1998), Mitch (1998), Danielle (1998), and Isabel (2003), Kepert (2006a,b) and Schwendike and Kepert (2008) have further documented asymmetric inflow and DHBL structures that vary from storm to storm. There is a deep inflow in the two rear quadrants near the radius of maximum wind (RMW) speed and in the rear-right quadrant in the outer region in Hurricane Georges. Unlike the azimuthally averaged fields shown in Zhang et al. (2011), the low-level outflow layer above DHBL exists only in the front-left quadrant (Kepert 2006a). A similar deep inflow layer was also found in Hurricane Mitch in the two left quadrants (Kepert 2006b). One of the differences is that Hurricane Mitch is a relatively slow-moving storm compared with Hurricane Georges. Schwendike and Kepert (2008) found no outflow in the observed mean profiles from the inner core to the outer region in Hurricane Danielle, whereas there is deeper inflow capped by a layer of outflow near the eyewall in Hurricane Isabel. The depth and strength of the inflow and outflow in hurricanes may also be affected by a moist convective heating-induced imbalance that is not included in the boundary layer models in Kepert and Wang (2001).

Other factors, such as storm-induced surface waves that are highly asymmetric around the storm (e.g., Bao et al. 2000; Wright et al. 2001; Doyle 2002), can also affect the surface winds and boundary layer properties through the modulation of sea surface roughness by the waves (Donelan et al. 2004). This study aims to better understand the effects of air-sea coupling on the HBL using a fully coupled atmosphere-wave-ocean model and observations. Aircraft data are used for both model verification and observation of the HBL. Section 2 provides a detailed description of the coupled model and the data used in this study. Coupled model simulations of Hurricane Frances (2004) are presented in sections 3 and 4. A detailed analysis of the HBL in Hurricane Frances from both the dropsonde data and the model simulations are given in sections 5 and 6, which are followed by some concluding remarks in section 7.

2. Coupled model and data

a. Fully coupled atmosphere-wave-ocean model

The University of Miami Coupled Model (UMCM) can be configured with various atmosphere, wave, and

ocean component models. In this study, UMCM consists of the fifth-generation Pennsylvania State University– National Center for Atmospheric Research Mesoscale Model (MM5) (Grell et al. 1994; Dudhia 1993), the WAVEWATCH III (WW3) (Tolman 1991, 1999; Tolman et al. 2002), and the three-dimensional Price– Weller–Pinkel (3DPWP) upper-ocean model (Price 1981; Price et al. 1994). This configuration is referred to as UMCM-MWP (Chen et al. 2012, manuscript submitted to *J. Atmos. Sci.*, hereafter CZDT). The coupling is done through a coupler that controls the communication between the component models.

The atmospheric component of UMCM-MWP is a multinested MM5 with 45-, 15-, 5-, and 1.67-km grid resolutions. The model domains are 120 \times 120, 121 \times $121, 121 \times 121$, and 151×151 grid points, respectively. The three inner nests are vortex-following moving grids (Tenerelli and Chen 2001). There are 28 vertical levels with 9 of them in the lowest 1 km (approximately at 11, 50, 125, 230, 350, 490, 625, 780, and 950 m). We use the Tao and Simpson (1993) microphysics scheme for all four domains. A slightly modified Kain-Fritsch cumulus parameterization (Kain and Fritsch 1993) is used only on the outer 45- and 15-km domains. The Blackadar boundary layer scheme (Zhang and Anthes 1982) is used here with a modification of the thermal exchange coefficient over the ocean based on Garratt (1992). Garratt's parameterization introduces different roughness scales for temperature z_t and moisture z_q , which is different from the roughness length for momentum z_o . The surface roughness scale is an essential term to calculate the surface exchange fluxes based on bulk formula. For the uncoupled MM5 and coupled MM5-3DPWP applications, the momentum roughness length over the open ocean is calculated from the Charnock's relationship (Charnock 1955). In the fully coupled UMCM-MWP, stress is explicitly computed in vector form from the wave stress using the 2D wave spectra plus the skin drag (CZDT). The turbulent closure is first order and approximated by the K theory.

The 3DPWP is a multilayer upper-ocean circulation model with three-dimensional physical processes including vertical mixing, horizontal advection, vertical advection, and pressure gradient. There are 30 layers in 3DPWP with resolutions varying from 5 m in the mixed layer to 20 m below down to 390-m depth. The model does not have bathymetry. A detailed description can be found in Price et al. (1994).

WW3 version 1.18 is used to simulate ocean surface waves in UMCM-MWP. It was developed by Tolman (1991, 1999) for wind waves in slowly varying, unsteady, and inhomogeneous ocean depths and currents and was evaluated extensively and validated with observations (Tolman et al. 2002). The wind waves are described by the action density wave spectrum $N(k, \theta, x, y, t)$. In this study, we use 25 frequency bands, logarithmically spaced from 0.0418 to 0.41 Hz at intervals of $\Delta f/f = 0.1$ and 48 directional bands (7.5° interval). The WW3 model domain is set to be about the same as the outer domain of MM5. The grid spacing is $\frac{1}{6}$ ° in both the latitudinal and longitudinal directions. The water depth data used in the wave model are the 5-min gridded elevation data from the National Geophysical Data Center.

b. Data

The observational data used in this study include the National Oceanic and Atmospheric Administration (NOAA) WP-3D aircraft GPS dropsondes, the Stepped Frequency Microwave Radiometer (SFMR), and airborne expendable bathythermograph (AXBT) data as well as the Tropical Rainfall Measuring Mission (TRMM) and Advanced Microwave Scanning Radiometer for Earth Observing System (AMSR-E) satellite data. The aircraft data are from NOAA WP-3D aircraft flights from 30 August to 1 September 2004. The storm center locations are from the National Hurricane Center (NHC)'s best-track data, which are used to calculate tangential and radial wind fields from GPS dropsondes. The AXBT data are used for ocean model initialization as well as for calculating surface fluxes. The TRMM Microwave Imager (TMI) and AMSR-E satellite SST data are used in the ocean model initialization.

c. Experiment design and model initialization

Three experiments are conducted using 1) the uncoupled atmospheric model, MM5 (UA); 2) the coupled atmospheric-ocean model, MM5-3DPWP (AO); and 3) the fully coupled atmospheric-waveocean MM5-WW3-3DPWP (AWO). The National Centers for Environmental Prediction (NCEP) global analysis fields (6 hourly and $1^{\circ} \times 1^{\circ}$) are used to initialize the MM5 and provide continuous lateral and lower boundary conditions. The model SST is initialized from NCEP reanalysis SST blended with satellite (TMI/ AMSR-E) SST using the method described in Chen et al. (2001). The NCEP reanalysis and TMI/AMSR-E SST is from 1 day prior to the beginning of the model simulation. The vertical structure of the ocean is initialized using observed and climatological profiles. The temperature profile is blended with a prestorm AXBT observation from the NOAA research aircraft mission and World Ocean Atlas 1994 climatology (Levitus and Boyer 1994) for depths greater than sampled by AXBT observation. The salinity profile is from World Ocean *Atlas 1994* climatology (Levitus et al. 1994), since there is generally no in situ prestorm observation available. Although the 3DPWP in UMCM-MWP can be initialized using the operational Hybrid Coordinate Ocean Model (HYCOM), the HYCOM fields were not used in this case because of a systematic temperature bias compared with the ABXT data during the time period of Hurricane Frances in August 2004.

3. Hurricane Frances (2004)

Frances was a strong category 4 hurricane that reached its peak intensity with the maximum wind speed (MWS) of 63 m s⁻¹ and the minimum sea level pressure of 938 hPa at 1800 UTC 31 August 2004. Its long residence over the open ocean makes it an ideal case for this study (Fig. 1a). Hurricane Frances went through an eyewall replacement cycle on 29-30 August while it slowly weakened. Reintensification began at about 1200 UTC 30 August, and Frances reached its peak intensity as it passed north of the Leeward and Virgin Islands (Beven 2004). The model-simulated storm went through a similar eyewall cycle with a relative minimum intensity at about 0000 UTC 31 August, about 12 h later than the observations (Fig. 1b). The focus here is from 1800 UTC 30 August to 0000 UTC 1 September, when the dropsonde data are available and the storm was away from the landmass. Another advantage of choosing this period is that the model-simulated storm tracks and translation speeds are very close to that of the NHC best-track estimates during this time (Fig. 1a), which is important for comparing the storm structure and HBL properties with observations.

From 0000 UTC 31 August to 0000 UTC 1 September, the three model simulations show a similar trend of intensification as the observations, but they have a significant difference in the model-simulated MWS. Although UA seemingly underpredicts storm intensity in terms of MWS, the mean sea level pressure (MSLP) in UA is much lower than the best-track data (not shown). The inconsistency in the pressure-wind relationship in the uncoupled model is attributed to the lack of coupling to the ocean waves and the unrealistic roughness formulations using the Charnock relationship in UA and AO. Storm-induced SST cooling in AO led to an even weaker MWS than UA as expected. AWO improves the model-simulated surface winds significantly due to the wind-wave coupling that reduces surface stress at high wind speeds (CZDT), The question remains as to how the air-sea coupling affects the atmospheric boundary layer structure in hurricanes, which may be of importance to model-simulated overall storm structure and intensity.



FIG. 1. (a) The NHC best-track (black) and three model-simulated tracks of Hurricane Frances (UA—blue, AO—green, and AWO—red) from 27 Aug to 6 Sep 2004. (b) Observed and simulated MWSs from 1200 UTC 30 Aug to 0000 UTC 1 Sep, which are the airborne observations used in this study.

4. Surface winds and air-sea fluxes

a. Spatial and temporal variations

Although the storm intensity is traditionally estimated by the MWS anywhere in the storm, there is a significant spatial variation of surface winds within each storm at any given time. The spatial distribution of the surface winds is a much better representation of the overall storm structure, which not only affects storm evolution and intensity, but also is the most relevant measure of hurricane impact on the ocean. In this section, all of the surface wind analyses of model simulations and observations are Earth relative. Figure 2 shows the surface wind, enthalpy (sensible + latent heat), and momentum fluxes from the UA, AO, and AWO simulations at 1800 UTC 31 August. The model-simulated surface winds are verified against the SFMR data from the NOAA WP-3D aircraft across the center of the storm from 1650 to 1800 UTC 31 August (Fig. 3). All three simulations are able to capture the asymmetry in the inner core with a relative minimum in the southwestern quadrant (Figs. 2a-c and 3), whereas the model-simulated eyewalls are slightly larger than the SFMR measurement. Overall, AO is relatively weaker than UA and AWO. While the modeled wind profiles are broader than the observation, AWO improves the surface wind speed, especially in the outer region.

The surface enthalpy fluxes are significantly different among the coupled and uncoupled simulations (Figs. 2d–f). The hurricane-induced SST and upperocean cooling reduce the enthalpy flux, especially in the rear-right quadrant in AO and AWO as compared to that of UA. There are substantial differences in enthalpy fluxes between UA and AO, not only in the region of strong cooling, but also near the RMW (\sim 30 km in both simulations), where there is relatively little cooling. This is because the largest reduction of winds in AO occurs at RMW, which has a large effect on the enthalpy flux. It is interesting to note that there is a region of enhanced enthalpy flux downwind of the cold wake in AO, which may be due to a decrease in air temperature upstream when it flows over the cold wake.

The spatial distributions of momentum fluxes are similar to that of surface wind speeds. The momentum flux in UA is stronger in the eyewall region near the RMW than AO and AWO (Figs. 2g–i). The difference between UA and AO is mostly due to the weaker surface winds because of ocean cooling in AO. However, the wind speed is stronger in AWO than AO, which is due to the difference in the stress formulations with and without the wind–wave coupling. Wind–wave coupling reduces stress at higher wind speed in AWO compared to that of the Charnock relationship used in UA and AO (CZDT).

The evolution of surface properties is examined over the period from 1200 UTC 30 August to 0000 UTC 1 September. Figure 4 shows the time series of stormaveraged SST anomaly (i.e., the difference between



FIG. 2. The model-simulated (a)–(c) surface wind speeds, (d)–(f) enthalpy fluxes, and (g)–(i) momentum fluxes from (left) UA, (middle) AO, and (right) AWO, averaged over a 2-h period centered at 1800 UTC 31 Aug. The black arrows indicate the direction of storm motion. The white line in (a)–(c) marks a reconnaissance flight path of the NOAA WP-3D aircraft, where the aircraft measurement will be shown in Fig. 3.

SST and its initial value) and air-sea fluxes averaged over an annular area between the radii of 0.5 and 5.0 times the RMW, and the azimuthally averaged peak wind speed at RMW. On average, about a 0.5° C SST cooling led to more than a 100 W m⁻² reduction in latent heat flux (LH) and about 20 W m⁻² in sensible heat flux (SH) in AO and AWO. The average reduction of LH and SH over the entire 36-h period in AWO (AO) is 10 (15) and 60 (80) W m⁻², respectively, which resulted in a decrease of about 20% in enthalpy flux in AO and about 15% in AWO compared to that in UA.

b. Symmetric and asymmetric structures

Azimuthally averaged fields are used here to represent the symmetric structure. The tangential and radial wind components (V_t and V_r , respectively) are used to describe the mean vortex structure, including the storm size (e.g., the radius of maximum wind) and the secondary circulation (e.g., inflow and outflow), respectively. The azimuthally averaged V_t and V_r at the surface from all three simulations are shown in Fig. 5. The mean V_t is consistent with the surface profiles,



FIG. 3. Surface wind speeds from the SFMR measurement (black), and three model simulations (UA—blue, AO—green, and AWO—red) along the flight path indicated in Fig. 2. The SFMR data are collected during the time period from 1650 to 1800 UTC 31 Aug, while the model fields are sampled at 1700 UTC 31 Aug.

including the SFMR observation shown in Fig. 3. The AO is slightly weaker than UA due to the coupling to the ocean. Comparing AO and AWO, the wind–wave coupling tends to reduce the V_t outside of the RMW and increase the peak V_t near the RMW (Fig. 5a), which corresponds to an increase of radial inflow outside of the RMW (Fig. 5b). The near-surface radial inflow is driven by friction. The increase in radial inflow outside the RMW can be attributed directly to the effect of coupling to the surface waves that reduces (increases) the drag coefficient in high (low) wind speeds compared to the Charnock relationship used in UA and AO (Donelan et al. 2004; CZDT).

The characteristics of asymmetric component of winds are shown in Fig. 6. The storm is divided into four quadrants according to the direction of storm motion: front left (FL), front right (FR), rear left (RL), and rear right (RR). Near the RMW, there is a noticeable decrease in tangential wind in the RR quadrant in AO compared to that in UA (Fig. 6a), which is corresponds to the storm-induced SST cooling and the large reduction in enthalpy flux (Figs. 2d and 2e). A similar reduction is found in radial wind speed as well (Fig. 6b). Wind-wave coupling in AWO increases the mean tangential wind on the right side of the storm more than the left side (Fig. 6a). Outside of the RMW, the most noticeable difference is in the enhanced inflow in AWO, with the largest departure from UA and AO in the FL quadrant (Fig. 6b). The large difference in radial inflow from AWO compared to that of UA and AO is mostly



FIG. 4. Time series of model-simulated (a) SST anomaly, (b) azimuthally averaged peak surface wind speed at the RMW, (c) latent heat fluxes, (d) sensible heat fluxes, (e) enthalpy fluxes, and (f) momentum fluxes averaged over an annular area between radii of 0.5 and 5.0 times the RMW from 1200 UTC 31 Aug to 0000 UTC 1 Sep (UA—blue, AO—green, and AWO—red).



FIG. 5. Azimuthally averaged surface (a) tangential and (b) radial wind speeds at 1800 UTC 31 Aug from the model simulations.

due to the effect of wind-wave coupling that reduces (increases) the drag coefficient at higher wind speeds in the inner core (outer region). Figure 7 shows the drag coefficients in the four quadrants. AWO has a higher value of drag coefficient outside the inner core (>50 km) but a smaller value at the RMW compared to UA and AO. There is no significant difference from quadrant to quadrant.

5. Hurricane boundary layer structure

The GPS dropsonde data from the NOAA WP-3D flights during the CBLAST Hurricane field campaign in 2004 provides a unique opportunity to examine the HBL structure in Hurricane Frances. Here we stratify the dropsonde data into subregions in both radial and azimuthal directions around the center of the storm. The locations of all dropsondes deployed from the three WP-3D flights in Hurricane Frances from 30 August to



FIG. 6. As in Fig. 5, but averaged fields over each of the four quadrants divided based on the storm forward motion pointed to the top of the figures.

1 September are shown in Fig. 8. To compare the model simulations with observations, vertical profiles of model fields are sampled at the same storm-relative locations and times. The model simulations are verified







FIG. 8. Storm-relative locations of 34 dropsondes from NOAA WP-3D aircraft flights in Hurricane Frances used in this study. The dropsonde data are collected from 30 Aug to 1 Sep. The numbers indicate the dropsondes that will be shown individually in Figs. 10 and 11, whereas the asterisks are dropsondes included only in the mean. (left) The circles indicate radii of 50, 100, 150, and 200 km. (right) The inner core region within the 50-km radius. The storm forward motion is pointed to the top.

accordingly. Both the mean properties of the HBL and spatial variability are examined. To investigate the effects of full atmosphere–wave–ocean coupling on the HBL, a comprehensive analysis of both the traditional mixed-layer property defined by the THBL and the inflow layer defined by the DHBL will be presented. All the wind analyses are Earth relative, except section 5c, in which the influence of the storm motion on the HBL is examined in both the Earth-relative and storm-relative framework.

a. Vertical profiles of winds and temperature

1) DHBL

Azimuthally averaged vertical profiles of V_t and V_r winds from the dropsondes and the model simulations are shown in Fig. 9. Although Frances was not in a steady state, the intensity change from 31 August to 1 September was relatively small (within a 10 m s⁻¹ range). Nevertheless, it is possible that the composite could provide a somewhat skewed depiction of the storm structure depending on the data coverage in time and space. In the inner-core region, the maximum V_t occurs near but below the top of the inflow layer, that is, at a height of about 600 m at the RMW (Figs. 9a and 9d) and about 1000 m at 2 times the RMW (2RMW) (Figs. 9b and 9e). An outflow region is right above the inflow layer. These are similar to that described in Kepert (2006a,b) and Schwendike and Kepert (2008). Although all three simulations have captured the general features of the DHBL, AWO is most close to the observation with the best tangential winds and inflow strength, whereas UA overestimated V_t throughout the lower troposphere, which means that UA produced a much stronger vortex than AWO even though the surface winds are relatively close to each other (Fig. 9a). However, all model simulations have a much weaker outflow than that of observed at the RMW. The outflow is mainly a consequence of the upward-transported supergradient momentum carried by the eyewall updraft (Kepert and Wang 2001). This result indicates that the models may be underpredicting the upward momentum transport. One possible reason could be that the vertical velocity in the model is underpredicted due to the coarse vertical resolution. The inflow at 2RMW is slightly stronger than that at the RMW, as shown by both the dropsonde data and the AWO simulation (Fig. 9b). In the outer region, the inflow diminishes at about 1500 m at 5 times the RMW (5RMW), and there is no outflow above the inflow layer (Figs. 9c and 9f), which is in agreement with Kepert (2006a). One of the differences in model-simulated V_t compared to the

DECEMBER 2012



FIG. 9. The mean profiles of (a)–(c) tangential and (d)–(f) radial winds of all dropsondes shown in Fig. 8 at the (left) RMW, (middle) 2RMW, and (right) outer region greater than 5RMW, and azimuthally averaged profiles from UA (blue), AO (green), and AWO (red) simulations at 1800 UTC 31 Aug.

observations is that model tends to produce a linear, rather than a logarithmic, profile near the surface, which may be due to a problem in the implementation of the Blackadar PBL scheme in MM5 as discussed in Kepert (2012).

There is a large spatial variability that deviates from the mean properties of the boundary layer properties from the dropsonde data. Figures 10 and 11 show V_t and V_r from individual dropsondes around the center of the storm. To fit the limited space, a subset of the dropsondes shown in Fig. 8 is used, which covers four quadrants where the data are available. There were more dropsondes on the right side than on the left side. Overall, the tangential wind profiles are similar to the mean, except some do not show the low-level maximum in the inner region (i.e., from the center to 2RMW) as in the mean. There is a left–right asymmetry with the strongest V_t on the right side, which is similar to that of model simulations shown in Figs. 2a–c.

The spatial variability is more apparent in the radial wind profiles. The inflow layer is the shallowest at about 200–300 m near the RMW close to the center (sondes



FIG. 10. Observed profiles of tangential winds (m s⁻¹) from 12 dropsondes inside the 50-km radius and 9 dropsondes in the outer region. The storm-relative positions of the numbered dropsondes are shown in Fig. 8. The layout of the dropsondes here are only approximate to what is shown in Fig. 8 (e.g., the actual locations of dropsondes 5, 21, and 22 are rear right relative to the center of the storm).

10, 31, and 32), while the inflow deepens radially outward within the inner core region to 800–1000 m near 2RMW (sondes 12, 16, and 18). Some dropsondes display a clear outflow layer about the inflow layer in the inner core (Fig. 11). In the outer region, however, the inflow layer is similar or even shallower than that at 2RMW in the front quadrants (sondes 14 and 15) and much deeper in the rear quadrants where there is no outflow (sondes 19, 20, and 28). These features are different from the mean inflow layer described in Zhang et al. (2011), but they are consistent with that in Hurricane Georges (1998) shown in Kepert (2006a). This



FIG. 11. As in Fig. 10, but for the radial winds (m s^{-1}). Negative (positive) values are inflow (outflow).

asymmetric structure will be discussed further in the next section.

2) THBL

Since the THBL is essentially defined by the mixed layer, θ_v profiles are used to examine the properties of the THBL. The azimuthally averaged and spatial distributions of θ_v around the storm are shown in Figs. 12

and 13, respectively, using the same dropsondes as in Figs. 9–11. The depth of the mixed layer increases from the inner core to the outer region, from about 200 m at RMW to 600–700 m at 5RMW on average (Fig. 12). The model simulations produced THBL heights similar to that of dropsonde observations. However, the values of θ_{ν} in the models are 2–3 K higher than the observations in the inner core, whereas the values are similar to the



FIG. 12. As in Fig. 9, but for virtual potential temperature.

observations in the outer region. The coupled AO and AWO model simulations were able to reduce the high θ_{ν} value by 0.5–1.0 K, but they are still higher than the observations in the inner core (Fig. 12). It is possible that the SST in the model, initialized from the TRMM/AMSR-E satellite data, is too warm. The rear-right quadrant has the shallowest mixed layer (sondes 21 and 22, locations are shown in Fig. 8) than other quadrants at 2RMW, where the hurricane-induced SST cooling is the most pronounced (Fig. 2e).

To further examine the thermodynamic properties of the THBL, the equivalent potential temperature θ_e is used to represent the moist potential energy of the boundary layer. Again, we separate dropsondes into the inner- and outer-core regions (Fig. 14). Moreover, the dropsondes located in the RR quadrant are shown separately from other three quadrants (OQ) to examine the influence of the storm-induced SST cooling. The θ_e is about 363 K in the inner core and 353–355 K in the outer region from the dropsonde data with values in the RR quadrant 2-3 K generally lower than in OQ, except near the surface in the inner core (Figs. 14a and 14b). The model-simulated θ_e profiles are sampled in the same storm-relative locations as the dropsondes. The results show that the uncoupled UA overestimates the θ_e by more than 5 K in both the inner-core and outer regions due to the lack of storminduced cooling in the upper ocean. The fully coupled AWO produced the θ_e profiles that are closest to the observations both near the surface and up to the 2-km level (Figs. 14g and 14h).

b. Symmetric and asymmetric structures in HBL height

While the symmetric structure of the HBL has been documented in axisymmetric models (e.g., Smith 1968, 2003; Kepert 2001) and by azimuthally averaged fields from observations (e.g., Zhang et al. 2011), the asymmetric structures are difficult to examine systematically due to the lack of a full 3D observation of winds and thermodynamic fields in time and space. Given that the fully coupled AWO simulation has been verified well with the dropsonde observations as shown in Figs. 9 and 12, here we compare the azimuthally averaged THBL and DHBL (Fig. 15) with that of four quadrants (Fig. 16) from the three model simulations to examine the symmetric and asymmetric HBL structures in Hurricane Frances. The general characteristics of the model-simulated DHBL and THBL show the height of the HBL increases radially outward from the RMW to the outer region as in observations (Figs. 9 and 12), which is in agreement with the previous studies of Smith (1968), Kepert (2001), and Zhang et al. (2011). The depth of the DHBL is almost twice as high as that of the THBL (Fig. 15), which is consistent with the dropsonde data shown in this study as well as in Zhang et al. (2011). A noticeable feature is that the height of the DHBL increases rapidly from inside the RMW to 2RMW where the inflow ascends sharply into the eyewall, and then it becomes somewhat flat for radii greater than 150 km in the outer region. The height of the THBL increases outward gradually. The THBL in AO and AWO is slightly shallower than that in UA.



FIG. 13. As in Fig. 10, but for virtual potential temperature (K). The gray line indicates the THBL calculated from each sounding.

The actual HBL structures shown in all four quadrants (Fig. 16) are quite different from the azimuthally averaged mean fields as shown in Fig. 15. There is a dramatic front–rear asymmetry in the DHBL height. The inflow layer is much shallower in the front than in the rear quadrants. This strong front–rear asymmetry exists in all model simulations, which indicates it is unrelated to air–sea coupling but a possible storm motion–induced asymmetry. Unlike the azimuthally averaged DHBL,

the height of the inflow layer in the front quadrants decreases outward from 2RMW. Furthermore, the deep inflow layer is likely associated with deep convective heating in the eyewall and rainbands of the hurricane, rather than the surface friction–induced inflow. It raises two concerns: 1) the representativeness of the azi-muthally averaged HBL properties as shown in Zhang et al. (2011) and 2) whether the use of the inflow layer to define the HBL is a valid approach.



FIG. 14. Mean θ_e profiles (K) from dropsondes and model simulations. (a),(c),(e),(g) Dropsondes from inside the 50-km-radius inner-core region and (b),(d),(f),(h) from the outer region. The solid lines are the mean profiles in the RR quadrant, and the dashed lines are from OQ. Number of dropsondes in each group is indicated in parentheses. The model fields are sampled according to the storm-relative locations and times of the dropsondes.

The most noticeable asymmetry in the THBL is in the rear-right quadrant in the AO and AWO simulations, where the THBL is shallowest around the storm (Figs. 16b and 16c). It is mostly due to the storm-induced SST cooling and associated stabilizing effects where the warm air flows over the colder ocean surface.

c. Effects of storm motion and deep tropospheric inflow on HBL

To remove the influence of the storm motion on the HBL asymmetry, we compute the storm-relative inflow by subtracting the storm translation velocity. For



FIG. 15. Azimuthally averaged radial winds (shading) as a function of radius and height at 1800 UTC 31 Aug from (a) UA, (b) AO, and (c) AWO simulations. The heights of the THBL and DHBL are shown in solid and dashed contours, respectively. The gray lines mark the surface RMW.

a comparison, horizontal maps of both the earth-relative and the storm-relative inflow layer depth in Hurricane Frances from the AWO simulation are shown in Figs. 17a and 17d. Both inflow fields are averaged over a 2-h



FIG. 16. As in Fig. 15, but averaged over each of the four quadrants. The storm forward motion points to the top.



FIG. 17. The coupled model-simulated (a)–(c) earth-relative DHBL and (d)–(f) storm-relative DHBL heights in Hurricane Frances (2004), Hurricane Floyd (1999), and Typhoon Choiwan (2009). The inflow fields are averaged over a 2-h period in all three cases. The 200-km radius is marked with a black circle.

period to smooth out some temporal fluctuations. For the earth-relative inflow layer, there is a front-rear asymmetry in the inflow depth with a relative minimum in the front (mostly less than 800 m), whereas the inflow layer in the rear can be as deep as 10 km. This deep inflow layer is apparently unrelated to the direct boundary layer processes. The deep inflow in the rear quadrants is also shown in the dropsonde data up to the flight level in Fig. 11.

In comparison, the storm-relative inflow depth shows a left–right asymmetry, which varies from less than 1 km on the right to as high as 6 km on the left (Fig. 17d), which is consistent with the observed asymmetry shown in Kepert (2006a,b). Although the exact reason for the asymmetry in the inflow is unclear, there are indications that it may be related to the asymmetry in convective structure. As convective cells grow and decay in the cyclonic flow of a hurricane, they tend to evolve from convective to stratiform rain downwind from right to left as shown in Black et al. (2002). Stratiform rain regions are usually more inductive for enhanced midlevel inflow than convective regions, which is a topic beyond the scope of this study.

The asymmetry in the inflow depth in both the earth-relative (Fig. 17a) and storm-relative (Fig. 17d)

frameworks will be masked in the azimuthally averaged DHBL as shown in Fig. 15c. It is clear that the mean inflow depth does not represent the actual inflow depth or the DHBL in a 3D hurricane structure. In this regard, using the inflow layer to define the HBL will not be valid, which has also been pointed out previously in Kepert (2010).

The question is, are these characteristics of inflow unique to Hurricane Frances? It is difficult to fully address this question without a significantly large dataset. Nevertheless, with the limited space in this paper, we show two examples to demonstrate that the asymmetry in the inflow depth is independent of numerical models and/or boundary layer parameterizations and exists in other tropical cyclones. Hurricane Floyd (1999) and Typhoon Choiwan (2009) are both major tropical cyclones. The Floyd simulation is done using UMCM-MWP with the exact same AWO configuration as for Frances, while the simulation of Choiwan is from the coupled model consisting of the Weather Research and Forecasting Model (WRF) and the 3DPWP ocean model (Lee 2012). The Yonsei University (YSU) PBL scheme is used in WRF. In both cases, the front-rear asymmetry in the earth-relative inflow depth is similar to that of Hurricane Frances (Figs. 17a-c). The asymmetry

is shifted to the left–right orientation in the storm-relative inflow in both cases with a deep inflow layer greater than 6 km located on the left side (Figs. 17d–f). Comparing the inflow in the earth-relative and storm-relative frameworks, the storm motion seems to enhance the inflow asymmetry in the rear with the inflow depth up to 10 km or more in all cases (Figs. 17a–c). The inflow fields in Floyd and Choiwan are also averaged over a 2-h period, as with Frances. These coupled model simulations and the observations from Hurricane Frances shown here as well as in Kepert (2006a,b) and Schwendike and Kepert (2008) suggest that the deep inflow layer and the asymmetry in the inflow are common features in hurricanes, although they may vary in detail from storm to storm.

6. Conclusions

The characteristics of the hurricane boundary layer in Hurricane Frances (2004) are examined in both numerical model simulations and GPS dropsonde observations. The effects of the air-sea coupling on the surface winds, air-sea fluxes, and HBL structure are analyzed in detail using the fully coupled atmospherewave-ocean model UMCM-MWP (CZDT). Three numerical experiments are conducted to isolate the atmosphere-ocean and wind-wave coupling effects. Overall, the fully coupled AWO simulation of Frances produced the best storm intensity and structure in terms of the wind, surface values, and vertical profiles of θ_v and θ_e compared with the dropsonde and flight-level observations from the NOAA WP-3D aircraft research flights (Figs. 3, 9, and 13).

The coupling to the ocean with storm-induced cooling in SST resulted in an overall weaker storm with a reduced surface enthalpy flux from the ocean. It also induces a strong asymmetry in enthalpy flux with relatively lower values in the rear-right quadrant of the storm because of the presence of a persistent cold wake (Fig. 2). This feature leads to a similar asymmetry in the THBL with a shallower mixed layer in the rear-right quadrant in the coupled AO and AWO model simulations, which is absent in the uncoupled UA simulation (Fig. 16). The wind–wave coupling enhances the surface friction-induced inflow outside of the RMW due to the ocean surface wave-induced changes in the drag coefficient, which tends to produce a deeper inflow layer and DHBL in the AWO simulation compared to that without the wind–wave coupling in AO (Figs. 15b and 15c).

One of the most intriguing results of this study is that the inflow layer in tropical cyclones is highly threedimensional and can be induced by both surface friction and convective heating in hurricane eyewalls and rainbands. The azimuthally averaged inflow layer tends to misrepresent the overall inflow structure in tropical cyclones, especially the asymmetric structure (cf. Figs. 15–17), as also noted in Kepert (2010). The depth of the inflow layer can be several kilometers as shown in both the dropsonde observations and full-physics model simulations. It raises the question of validity in using the inflow depth to define the DHBL in tropical cyclones, especially because the frictionally and convectively induced inflows are impossible to separate in real storms. It also causes concern regarding the representativeness of the azimuthally averaged HBL properties as shown in Zhang et al. (2011), which mask some dominate features in the inflow depth and asymmetry.

The complex 3D structure and asymmetries in both thermodynamic and dynamic HBL properties that are in part associated with the air-sea and wind-wave coupling processes make it difficult to parameterize the atmosphere-wave-ocean coupling effects without a fully coupled model.

Acknowledgments. We thank Dr. Wei Zhao for his assistance in the model simulations of Hurricane Frances, and Drs. Ralph Foster, Peter Sullivan, Jim Price, and Brandon Kerns for their insightful discussions during the course of this study. The GPS dropsonde data were collected and supported by the NOAA Aircraft Operations Center and Atlantic Oceanographic and Meteorological Lab during the CBLAST field campaign. Comments and suggestions from three anonymous reviewers helped improve the manuscript significantly. This research was supported by the Office of Naval Research under the research grant of CBLAST Hurricane Program (N00014-01-1-0156) and a NOAA research grant (NA10NWS4680013).

REFERENCES

- Anthes, R. A., and S. W. Chang, 1978: Response of hurricane boundary layer to change of sea surface temperature in a numerical model. J. Atmos. Sci., 35, 1240–1255.
- Bao, J.-W., J. M. Wilczak, J.-K. Choi, and L. H. Kantha, 2000: Numerical simulations of air–sea interaction under high wind conditions using a coupled model: A study of hurricane development. *Mon. Wea. Rev.*, **128**, 2190–2210.
- Bender, M., and I. Ginis, 2000: Real-case simulations of hurricaneocean interaction using a high-resolution coupled model: Effects on hurricane intensity. *Mon. Wea. Rev.*, **128**, 917–946.
- Beven, J. L., 2004: Tropical cyclone report: Hurricane Frances; 25 August - 8 September 2004. NOAA Rep. [Available online at http://www.nhc.noaa.gov/2004frances.shtml.]
- Black, M. L., J. F. Gamache, F. D. Marks Jr., C. E. Samsury, and H. E. Willoughby, 2002: Eastern Pacific Hurricanes Jimena of 1991 and Olivia of 1994: The effect of vertical shear on structure and intensity. *Mon. Wea. Rev.*, **130**, 2291–2312.
- Black, P. G., and Coauthors, 2007: Air–sea exchange in hurricanes: Synthesis of observations from Coupled Boundary Layer Air– Sea Transfer experiment. *Bull. Amer. Meteor. Soc.*, 88, 357–374.

- Charnock, H., 1955: Wind stress on a water surface. *Quart. J. Roy. Meteor. Soc.*, **81**, 639–640.
- Chen, S. S., W. Zhao, J. E. Tenerelli, R. H. Evans, and V. Halliwell, 2001: Impact of the AVHRR sea surface temperature on atmospheric forcing in the Japan/East Sea. *Geophys. Res. Lett.*, 28, 4539–4542.
- —, J. F. Price, W. Zhao, M. A. Donelan, and E. J. Walsh, 2007: The CBLAST-Hurricane Program and the next-generation fully coupled atmosphere-wave-ocean models for hurricane research and prediction. *Bull. Amer. Meteor. Soc.*, 88, 311–317.
- —, W. Zhao, M. A. Donelan, and H. L. Tolman, 2012: Directional wind-wave coupling in fully coupled atmospherewave-ocean models: Results from CBLAST-Hurricane. J. Atmos. Sci., submitted.
- Donelan, M. A., B. K. Haus, N. Reul, W. J. Plant, M. Stiassne, H. C. Graber, O. B. Brown, and E. S. Saltzman, 2004: On the limiting aerodynamic roughness of the ocean in very strong winds. *Geophys. Res. Lett.*, **31**, L18306, doi:10.1029/2004GL019460.
- Doyle, J. D., 2002: Coupled atmosphere–ocean wave simulations under high wind conditions. *Mon. Wea. Rev.*, 130, 3087–3099.
- Dudhia, J., 1993: A nonhydrostatic version of the Penn State–NCAR mesoscale model: Validation tests and simulation of an Atlantic cyclone and cold front. *Mon. Wea. Rev.*, **121**, 1493–1513.
- Emanuel, K. A., 1986: An air-sea interaction theory for tropical cyclones. Part I: Steady-state maintenance. J. Atmos. Sci., 43, 585–605.
- Foster, R. C., 2009: Boundary-layer similarity under an axisymmetric, gradient wind vortex. *Bound.-Layer Meteor.*, 131, 321–344.
- Garratt, J. R., 1992: The Atmosphere Boundary Layer. Cambridge University Press, 316 pp.
- Grell, G. A., J. Dudhia, and D. R. Stauffer, 1994: A description of the fifth-generation Penn State/NCAR mesoscale model (MM5). NCAR Tech. Note NCAR/TN-398+STR, 122 pp.
- Kain, J. S., and J. M. Fritsch, 1993. Convective parameterization for mesoscale models: The Kain–Fritsch scheme. *The Representation of Cumulus Convection in Numerical Models, Meteor. Monogr.*, No. 46, Amer. Meteor. Soc., 165–170.
- Kepert, J. D., 2001: The dynamics of boundary layer jets within the tropical cyclone core. Part I: Linear theory. J. Atmos. Sci., 58, 2469–2484.
- —, 2006a: Observed boundary layer wind structure and balance in the hurricane core. Part I: Hurricane Georges. J. Atmos. Sci., 63, 2169–2193.
- —, 2006b: Observed boundary layer wind structure and balance in the hurricane core. Part II: Hurricane Mitch. J. Atmos. Sci., 63, 2194–2211.
- —, 2010: Comparing slab- and height-resolving models of the tropical cyclone boundary layer. Part I: Comparing the simulations. *Quart. J. Roy. Meteor. Soc.*, **136**, 1689–1699, doi:10.1002/qj.667.
- —, 2012: Choosing a boundary layer parameterization for tropical cyclone modeling. *Mon. Wea. Rev.*, **140**, 1427–1445.
- —, and Y. Wang, 2001: The dynamics of boundary layer jets within the tropical cyclone core. Part II: Nonlinear enhancement. J. Atmos. Sci., 58, 2485–2501.
- Lee, C.-Y., 2012: Energetic impacts of stable boundary layer on tropical cyclone structure in coupled WRF model. Preprints, 30th Conf. on Hurricane and Tropical Meteorology, Ponte Vedra Beach, FL, Amer. Meteor. Soc., 5D4. [Available online at http://ams.confex. com/ams/30Hurricane/webprogram/Paper206196.html.]
- Levitus, S., and T. P. Boyer, 1994: *Temperature*. Vol. 4, *World Ocean Atlas 1994*, NOAA Atlas NESDIS 4, 117 pp.
- —, R. Burgett, and T. P Boyer, 1994: *Salinity*. Vol. 3, *World Ocean Atlas 1994*, NOAA Atlas NESDIS 3, 99 pp.

- Montgomery, M. T., H. D. Snell, and Z. Yang, 2001: Axisymmetric spindown dynamics of hurricane-like vortices. J. Atmos. Sci., 58, 421–435.
- Pendergrass, A. G., and H. E. Willoughby, 2009: Diabatically induced secondary flows in tropical cyclones. Part I: Quasisteady forcing. *Mon. Wea. Rev.*, 137, 805–821.
- Powell, M. D., 1990: Boundary layer structure and dynamics in outer hurricane rainbands. Part II: Downdraft modification and mixed layer recovery. *Mon. Wea. Rev.*, **118**, 918–938.
- Price, J. F., 1981: Upper ocean response to a hurricane. J. Phys. Oceanogr., 11, 153–175.
- —, T. B. Sanford, and G. Z. Forristall, 1994: Forced stage response to a moving hurricane. J. Phys. Oceanogr., 24, 233–260.
- Rosenthal, S. L., 1962: A theoretical analysis of the field of motion in the hurricane boundary layer. National Hurricane Research Project Rep. 56, 12 pp.
- Rotunno, R., and K. A. Emanuel, 1987: An air–sea interaction theory for tropical cyclones. Part II: Evolutionary study using a nonhydrostatic axisymmetric numerical model. J. Atmos. Sci., 44, 542–561.
- Schade, L. R., and K. A. Emanuel, 1999: The ocean's effect on the intensity of tropical cyclones: Results from a simple coupled atmosphere–ocean model. J. Atmos. Sci., 56, 642–651.
- Schwendike, J., and J. D. Kepert, 2008: The boundary layer winds in Hurricane Danielle (1998) and Isabel (2003). *Mon. Wea. Rev.*, **136**, 3168–3191.
- Shapiro, L. J., 1983: The asymmetric boundary layer flow under a translating hurricane. J. Atmos. Sci., 40, 1984–1998.
- Smith, R. K., 1968: The surface boundary layer of a hurricane. *Tellus*, **20**, 473–483.
- —, 2003: A simple model of the hurricane boundary layer. Quart. J. Roy. Meteor. Soc., 129, 1007–1027.
- —, M. T. Montgomery, and N. V. Sang, 2009: Tropical cyclone spin-up revisited. *Quart. J. Roy. Meteor. Soc.*, 135, 1321–1335.
- Stull, R. B., 1988: An Introduction to Boundary Layer Meteorology. Kluwer Academic, 666 pp.
- Tao, W.-K., and J. Simpson, 1993: The Goddard cumulus ensemble model. Part I: Model description. *Terr. Atmos. Oceanic Sci.*, 4, 19–54.
- Tenerelli, J. E., and S. S. Chen, 2001: High-resolution simulations of Hurricane Floyd using MM5 with vortex-following mesh refinement. [Available online at https://ams.confex.com/ams/ WAF-NWP-MESO/webprogram/Paper23165.html.]
- Tolman, H. L., 1991: A third-generation model for wind waves on slowly varying, unsteady, and inhomogeneous depths and currents. J. Phys. Oceanogr., 21, 782–797.
- —, 1999: User manual and system documentation of WAVEWATCH-III version 1.18. NOAA/NWS/NCEP/OMB Tech. Note 166, 110 pp.
- —, B. Balasubramaniyan, L. D. Burroughs, D. V. Chalikov, Y. Y. Chao, H. S. Chen, and V. M. Gerald, 2002: Development and implementation of wind-generated ocean surface wave models at NCEP. *Wea. Forecasting*, **17**, 311–333.
- Wright, C. W., and Coauthors, 2001: Hurricane directional wave spectrum spatial variation in the open ocean. J. Phys. Oceanogr., 31, 2472–2488.
- Wu, C.-C., C.-Y. Lee, and I. I. Lin, 2007: The effect of ocean eddy on tropical cyclone intensity. J. Atmos. Sci., 64, 3562–3578.
- Zhang, D., and R. A. Anthes, 1982: A high-resolution model of the planetary boundary layer—Sensitivity tests and comparisons with SESAME-79 data. J. Appl. Meteor., 21, 857–861.
- Zhang, J. A., R. F. Rogers, D. S. Nolan, and F. D. Marks Jr., 2011: On the characteristic height scales of the hurricane boundary layer. *Mon. Wea. Rev.*, **139**, 2523–2535.