Directional Wind–Wave Coupling in Fully Coupled Atmosphere–Wave–Ocean Models: Results from CBLAST-Hurricane

SHUYI S. CHEN, WEI ZHAO,* AND MARK A. DONELAN
Rosenstiel School of Marine and Atmospheric Science, University of Miami, Miami, Florida

HENDRIK L. TOLMAN
Environmental Modeling Center, NOAA/National Centers for Environmental Prediction, Camp Springs, Maryland

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ABSTRACT

The extreme high winds, intense rainfall, large ocean waves, and copious sea spray in hurricanes push the surface-exchange parameters for temperature, water vapor, and momentum into untested regimes. The Coupled Boundary Layer Air–Sea Transfer (CBLAST)-Hurricane program is aimed at developing improved coupling parameterizations (using the observations collected during the CBLAST-Hurricane field program) for the next-generation hurricane research prediction models. Hurricane-induced surface waves that determine the surface stress are highly asymmetric, which can affect storm structure and intensity significantly. Much of the stress is supported by waves in the wavelength range of 0.1–10 m, which is the unresolved “spectral tail” in present wave models. A directional wind–wave coupling method is developed to include effects of directionality of the wind and waves in hurricanes. The surface stress vector is calculated using the two-dimensional wave spectra from a wave model with an added short-wave spectral tail. The wind and waves are coupled in a vector form rather than through the traditional roughness scalar. This new wind–wave coupling parameterization has been implemented in a fully coupled atmosphere–wave–ocean model with 1.67-km grid resolution in the atmospheric model, which can resolve finescale features in the extreme high-wind region of the hurricane eyewall. It has been tested in a number of storms including Hurricane Frances (2004), which is one of the best-observed storms during the CBLAST-Hurricane 2004 field program. This paper describes the new wind–wave coupling parameterization and examines the characteristics of the coupled model simulations of Hurricane Frances (2004). Observations of surface waves and winds are used to evaluate the coupled model results.

1. Introduction

The intensity of a hurricane is affected by two competing physical processes at the air–sea interface: the heat and moisture fluxes that fuel the storm and the surface friction (or momentum flux into the ocean) that dissipates the storm. Using an idealized axisymmetric tropical cyclone (TC) model Emanuel (1995) has demonstrated the sensitivity of TC intensity to the surface-exchange coefficients of the enthalpy (heat and moisture) and momentum fluxes. However, the enthalpy and momentum exchange coefficients under the high-wind conditions are difficult to determine, especially in the eyewall region of a hurricane, where very few observations of a coherent sea state and boundary layers across the air–sea interface exist. Under the high-wind conditions the stress is supported partly (~50%) by waves in the wavelength range of 0.1–10 m (Phillips 1977, Donelan 1998), which are unresolved by present wave models. Another unique feature is the swirling winds in hurricanes can induce complex wave fields that may result in wave-induced stress misaligned with the winds. The rapid increase in computer power and recent advances in technology in observations have made it possible for us to develop a new generation of high-resolution, coupled atmosphere–wave–ocean hurricane prediction models that are capable

* Current affiliation: Physical Oceanography Lab, Ocean University of China, Qingdao, China.

Corresponding author address: Dr. Shuyi S. Chen, RSMAS/University of Miami, 4600 Rickenbacker Causeway, Miami, FL 33149.
E-mail: schen@rsmas.miami.edu

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of representing the complex sea states with explicit coupling to ocean surface waves in high-wind conditions (Sullivan and McWilliams 2010). We begin by developing and examining key parameterizations including effects of the directional wave spectra and wave spectral tail on drag coefficient and storm structure and intensity at 1–2-km grid resolution.

Hurricanes rarely reach their maximum potential intensity (MPI) (Emanuel 1986, 1995; Holland 1997). Many factors prevent storms from reaching their MPI, including environmental vertical wind shear, distribution of tropospheric water vapor, hurricane internal dynamics, and air–sea interactions. The effect of the air–sea interactions (especially the wind–wave coupling under extreme high-wind conditions) on hurricane structure and intensity change was the focus of the Coupled Boundary Layer Air–Sea Transfer (CBLAST)-Hurricane program (e.g., Black et al. 2007, Chen et al. 2007).

Emanuel (1995) proposed that the storm intensity is largely controlled by the ratio of the air–sea enthalpy and momentum flux exchange coefficients, $C_K/C_D$. Using a simple axisymmetric model with idealized environmental conditions, Emanuel (1995) showed that this ratio needs to be equal or greater than one for hurricanes to intensify. As shown in many studies, although $C_D$ is largely sea-state dependent (e.g., Toba et al. 1990; Donelan et al. 1993), $C_K$ has relatively little sensitivity to sea state (e.g., Geernaert et al. 1987, Jeong et al. 2012). Laboratory experiments conducted at hurricane wind speeds have shown that $C_D$ reaches a saturation point at high wind speeds greater than about 33 m s$^{-1}$ when flow separation from the dominant waves begins to occur (Donelan et al. 2004), similar to that estimated from the GPS dropsonde data (Powell et al. 2003). However, $C_K$ remains relatively constant as wind speed increases (Jeong et al. 2012). Airborne turbulence flux measurements from the CBLAST-Hurricane also support these laboratory results, indicating that $C_K/C_D$ is at about 0.4–0.7, less than that predicted by Emanuel (1995), for intensifying storms such as in Hurricane Fabian (2003) (a major category 4 hurricane) reported by Drennan et al. (2007) and Zhang et al. (2008).

This study aims to better understand the coupling between the atmosphere, surface waves, and the upper ocean in hurricanes. The upper-ocean response to hurricane forcing has been documented in some detail in modeling and observational studies (e.g., Price 1981; Price et al. 1994; Shay and Elsberry 1987; Sanford et al. 1987; Bender and Ginis 2000; Shay et al. 2000). Effects of ocean surface waves on hurricanes have been examined in coupled models (e.g., Bao et al. 2000, Doyle 1995).

The wave stress was computed from a wave model and fed back to the wind calculation through a scalar roughness parameter in these previous studies.

This paper describes the results of a new directional wind–wave-coupling parameterization in a fully coupled model developed based on the CBLAST-Hurricane observations and laboratory measurements. Coupled model simulations of Hurricane Frances (2004) are compared with observations of surface wave spectra, surface fluxes, and vertical profiles of atmospheric boundary layer (Lee and Chen 2012) and ocean mixed-layer temperature and currents (Sanford et al. 2005). The focus of this study is on the development and testing of the directional wind–wave coupling parameterization in a fully coupled model.

2. Coupled modeling system

The University of Miami Fully Coupled Atmosphere–Wave–Ocean Modeling system (UMCM) includes three model components: atmospheric, surface wave, and ocean circulation models. Chen et al. (2007) gave a brief introduction to UMCM and an overview for the coupled modeling effort in the CBLAST-Hurricane program. Figure 1 shows a schematic of UMCM. Currently, UMCM can be configured with two different options in terms of component models: 1) coupled with the fifth-generation Pennsylvania State University–National Center for Atmospheric Research Mesoscale Model (MM5), a third-generation wave model (WAVEWATCH III), and the three-dimensional Price–Weller–Pinkel (3DPWP) upper-ocean model (UMCM-MWP); and 2) coupled with the Weather Research and Forecasting Model (WRF), the University of Miami Wave Model (UMWM; Donelan...
et al. 2012), and the Hybrid Coordinate Ocean Model (HYCOM) (UMCM-WMH). The basic coupling parameters are listed between each of the model components in Fig. 1. In this study, UMCM-MWP will be used and described in detail here.

### a. Atmospheric model

The atmospheric component of the coupled modeling system is MM5 (Grell et al. 1994, Dudhia 1993) with added features for TC prediction. To capture the long life cycle of hurricanes and to resolve the inner-core structure, we developed a vortex-following nested grid that allows the model to be integrated for 5 days or longer at very high resolution (~1–2 km) in the innermost domain. With the automatic vortex-following nest system, mesh positions are not known in advance. It is not practical to pregenerate terrain and land use files prior to running a simulation. Accordingly, we have modified the MM5 nest initialization routine so that elevation and land use data are read and placed on the fine meshes each time they are initialized or moved. The vortex-following nested grid system is described in detail in Tenerelli and Chen (2001). We use four nests with 45-, 15-, 5-, and 1.67-km grid spacing, respectively. The three inner domains move automatically with the storm. The domain sizes for each of the inner nests are 121 × 121, 121 × 121, and 151 × 151 grid points, respectively. There are 28 sigma levels in the vertical with approximately 9 levels within the atmospheric boundary layer. The model has been used to simulate Hurricane Bonnie (1998) (Rogers et al. 2003), Hurricane Georges (1998) (Cangialosi 2005), and Hurricane Floyd (1999) (Tenerelli and Chen 2001). These studies have shown that using 1.67-km grid spacing to resolve the inner-core (eye and eyewall) structure is a key in simulating hurricane evolution and intensity change.

We use an explicit moisture microphysics scheme and a slightly modified Kain–Fritsch (K–F) cumulus parameterization (Kain and Fritsch 1993) on the 45- and 15-km grids and the explicit moisture scheme only on the 5- and 1.67-km grids. The microphysics scheme used is based on Tao and Simpson (1993). The inner core of hurricanes is simulated explicitly in cloud-resolving mode. Modifications to K–F parameterization include detrain of 30% hydrometeors to the resolvable grids and a higher vertical velocity threshold for initiation of convective clouds, which is more suitable for the tropical oceanic conditions. The Blackadar PBL scheme (Zhang and Anthes 1982) is used on all grids, but over water we include a modification based upon Garratt (1992), in which we introduce different roughness scales for temperature $z_{r}$ and moisture $z_{m}$. In the original formulation of the Blackadar scheme, the roughness scales for temperature and moisture are identical to that for momentum $z_{o}$, which was not appropriate since the physics governing momentum transfer at the surface is different from that governing temperature and moisture. In the uncoupled MM5 applications, the momentum roughness length over the open ocean is calculated from the Charnock relationship (Charnock 1955):

$$z_{o} = a \frac{\tau}{\rho_a g},$$

where $\tau$ is the total stress, $\rho_a$ is the air density, and $a \ (=0.0185)$ is the Charnock parameter.

The National Centers for Environmental Prediction (NCEP) global analysis fields (6-hourly and 1° × 1° grid resolution) and the high-resolution (~9 km) Advanced Very High Resolution Radiometer (AVHRR) Pathfinder SST analysis (Chen et al. 2001) as well as the Tropical Rainfall Measurement Mission (TRMM) Microwave Imager (TMI)–Advanced Microwave Scanning Radiometer (AMSR) SST (~25 km) are used to initialize the uncoupled MM5 and provide continuous lateral and lower boundary conditions. The uncoupled MM5 using the satellite-observed SST that is kept constant with time will be referred to as “UA.”

### b. Ocean model

A three-dimensional, primitive equation, full-physics upper-ocean model (3DPWP; see Price et al. 1994) is used to simulate the upper-ocean current and temperature fields underneath a hurricane. The model domain is the same as the outer domain of the atmospheric model with 15-km grid spacing. It has 30 vertical levels with grid spacing of 5–10 m for the top 20 levels and 20 m for the rest. It is initialized using both CBLAST observations and climatological temperature and salinity profiles. The vertical structure of the ocean is initialized using observed and climatological profiles. The temperature profile is blended with a prestorm airborne expendable bathythermograph (AXBT) observation from the National Oceanic and Atmospheric Administration (NOAA) research aircraft mission and World Ocean Atlas 1994 climatologic data (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. At each time step of the ocean model (10 min) the ocean model takes surface stress and heat and moisture fluxes from the atmospheric and wave models and steps the ocean dynamics ahead. On the same times, the ocean model passes back the SST to the atmospheric model.
and the ocean surface current to the wave model. The simulations that utilize the ocean model to calculate the evolving SST will be referred to as “AO,” meaning that both the atmosphere and the ocean evolve in a coupled mode.

c. Surface wave model

The WAVEWATCH III (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. WW3 has been extensively evaluated and validated with observations (Tolman et al. 2002). The wind waves are described by the action density wave spectrum \( N(k, \theta, x, y, t) \), where \( k \) is the wavenumber and \( \theta \) is the azimuth angle. In this study, we use 25 frequency bands, logarithmically spaced from 0.0418 to 0.41 Hz at intervals of \( \Delta f = 0.1 \) and 48 directional bands (7.5° interval). The WW3 model domain is set to be approximately the same as the outer domain of MM5. The grid spacing is \( \frac{1}{8}^\circ \) in both latitude and longitude. The water depth data used in the wave model is the 5-min gridded elevation data from the National Geophysical Data Center. In the coupled system, the wave model inputs the surface wind and ocean surface current fields from the atmospheric and ocean models and outputs wave stress (Fig. 1) integrated from the wind–wave coupling parameterization discussed in detail in the next section. When the surface waves and wind are treated and coupled explicitly in this manner, the fully coupled model will be referred to as “AWO,” meaning that the atmosphere, the surface waves, and the ocean evolve concurrently.

We note that the effects of sea spray on the air–sea exchange may also be important (e.g., Fairall et al. 1994; Andreas and Emanuel 2001; Emanuel 2003). However, the lack of comprehensive field data discourages us from attempting an explicit treatment of this complex problem at this time. A separate study has been underway to examine the sensitivity of TC intensity to sea spray parameterization using wave dissipation and related spray generation in UMCM-MWP, which will be reported in a forthcoming paper.

3. Wind–wave coupling

a. Directional wave stress from two-dimensional wave spectra

The surface waves and wave stress in hurricanes are most complex. Two main issues are particularly important in hurricane conditions, which have not been addressed in a fully coupled atmosphere–wave–ocean model in previous studies. First, the swirling winds in tropical cyclones are highly variable; not only the wind speed varies anywhere from nearly 0 to more than 70 m s\(^{-1}\) in a major hurricane, but also the direction of winds varies tremendously. Second, the stress vector is not necessarily in the same direction as the local wind vector. The surface condition is mostly of the mixed wind sea and swell sea in high winds. The wave stress can be at a different direction than that of at 180° of the fast swirling winds in contrast to most previous studies treating wave stress as a scalar roughness parameter (e.g., Janssen 1991; Doyle 1995, 2002; Bao et al. 2000). Second, the wave-induced stress contributes most of the total stress in high-wind conditions (e.g., Donelan 1990). Existing third-generation wave prediction models are unable to fully represent this process as their high wavenumber cutoff is typically about 0.63 m\(^{-1}\) or 10-m wavelength, while most of the stress is supported by shorter waves. In WW3, the cutoff frequency is set to be \( 3f_p \), where \( f_p \) is the peak frequency of the wind sea as defined in Tolman and Chalikov (1996). In this study, a new wind–wave coupling method is developed to compute the directional wave stress (or form drag) using the two-dimensional wave spectra from WW3 with an appended spectral tail that is based on a direct balance of wind input and dissipation. The short waves in the spectra are assumed to be traveling symmetrically about the wind direction. The wave-stress vectors include contributions from both the long and short waves.

In the fully coupled atmosphere–wave–ocean model, the effective wind components of the surface stress vector are computed as \( \mathbf{u}_r = \sqrt{\tau_x/\rho_w C_D} \) and \( \mathbf{v}_r = \sqrt{\tau_y/\rho_w C_D} \), where \( \tau_x \) and \( \tau_y \) are the total (wave + skin) stress components. The components of the stress vector due to the waves (or “form drag”) \( \tau_{wx} \) and \( \tau_{wy} \) are computed from the integral of momentum input to the spectrum:

\[
\tau_{wx} = g \rho_p \int_0^{\infty} \int_{-\pi}^{\pi} \gamma F(k, \theta) k \, dk \, d\theta,
\]

\[
\tau_{wy} = g \rho_p \int_0^{\infty} \int_{-\pi}^{\pi} \gamma F(k, \theta) k \, dk \, d\theta,
\]

where \( F(k, \theta) \) is the energy spectrum of wavenumber \( k \) and direction \( \theta \) and \( \gamma/\omega \) is the dimensionless growth rate of waves with \( k \); \n
\[
\frac{\gamma}{\omega} = S \rho_w \left[ \frac{U(\pi/k) \cos \theta}{C(k)} - 1 \right] \left[ \frac{U(\pi/k) \cos \theta}{C(k)} - 1 \right],
\]

where \( U(\pi/k) \) is the wind speed at the height of the half wavelength.
The energy spectrum in Eqs. (2) and (3) consists of the skin, wave, and total stress, respectively.

\[
\begin{align*}
U_{(\pi/k)} &= \begin{cases} 
U_{10}, & \pi/k > 10 \text{ m} \\
U_{10} + \frac{\mu_u}{\kappa} \ln \left( \frac{\pi/k}{10} \right), & \pi/k < 10 \text{ m} 
\end{cases}
\end{align*}
\]

(5)

C is the phase speed, S is the sheltering coefficient,

\[
S = \begin{cases} 
S_1, & U_{10} < U_{10s} \\
S_2 + (S_1 - S_2)e^\left(\frac{(U_{10} - U_{10s})}{2U_{10s}}\right), & U_{10} > U_{10s}
\end{cases}
\]

(6)

S_1 = 0.28 and S_2 = 0.11 for growth and attenuation, respectively, based on Donelan (1999). The term U_{10} is the wind speed at 10 m and U_{10s} the threshold value at which the drag coefficient saturation occurs. It is currently set to equal to 33 m s^{-1} based on the results of the laboratory measurements from Donelan et al. (2004) as shown in Fig. 2a. The observed S_2 corresponds to very strong negative (wind against waves, which destroys the waves) forcing, where flow separation is expected to be prevalent. Flow separation from the dominant waves acts to shelter the short waves and reduces the stress they support (Banner and Morison 2010). Consequently, we allow S to asymptotically approach S_2 in very high winds and assumed strong flow separation. The observations from CBLAST are only valid up to about 30 m s^{-1} (e.g., Black et al. 2007), from which it is difficult to determine where flow separation from the dominant waves may occur over the open ocean.

The energy spectrum in Eqs. (2) and (3) consists of the spectrum of long waves F_{lw} from the WW3 and a short-wave spectrum F_{sw} that is unresolved by WW3. The spectrum of short waves from a fit to the tail of the long-wave spectrum is

\[
F_{sw}(k, \theta) = k^{-4} \left\{ \frac{S \rho_u}{\alpha \rho_w} \frac{U}{C} \cos(\theta - \theta) - 1 \right\}^{2} - \frac{4\nu k}{4\alpha C} \right\}^{1/n},
\]

(7)

where \(\alpha\) and n values are determined based on the wave-breaking processes observed from gravity wave spectra (Donelan and Pierson 1987). The viscosity \(\nu\) is given as a function of temperature. We first removed the built-in spectrum tail in WW3 and then patch the spectrum tail from Eq. (7) smoothly starting from 0.2 rad m^{-1}. The effect of breaking waves is implicit in this approach. More extensive discussions on breaking waves can be found in Kudryavtsev and Makin (2001) and Kukulka and Hara (2008). Makin (2005) emphasized that the effect of breaking waves on wave-induced stress is particularly important in high-wind conditions.

The skin stress \(\tau_s\), which is in the wind direction, is computed by iterating the law of the wall with the roughness Reynolds number \((Re_s = u_*z_0/\nu)\) set to 0.132; yielding both friction velocity and roughness length associated with skin stress. The denominator \(\nu\) is the kinematic viscosity of air (Monin and Yaglom 1971). The total stress vector is the sum of the wave (form) stress vector from Eqs. (2) and (3) and the skin stress:

\[
\tau_{sx} = \tau_{wx} + \tau_{sx},
\]

(8)

\[
\tau_{sy} = \tau_{wy} + \tau_{sy}.
\]

(9)

Although a similar wave-stress calculation has been suggested in Moon et al. (2004) and Fan et al. (2009) using an equilibrium spectrum formulation of Hara and Belcher (2002) appended to the long-wave spectrum calculated from WAVEWATCH III, the effect of the vector stress on winds has never been investigated in a coupled mode in these studies that do not have a fully coupled model.

b. Coupling in UMCM-MWP

The directional wind–wave coupling calculations in UMCM-MWP can be best summarized in a flowchart (see Fig. 2). The 2D wave spectrum and wind–wave stress are updated at each time step of the wave model. The total stress computed from the coupler is fed back to the atmosphere and ocean model. A detailed step-by-step calculation done in the wind–wave coupler is also given in Fig. 2.

c. Momentum and enthalpy exchange coefficients

To evaluate the coupled model results, \(C_D\) and \(C_K\) computed from the model simulations will be compared.
with observations obtained from a wind–wave tank (e.g., Donelan et al. 2004 and Jeong et al. 2012) and CBLAST field campaign (Black et al. 2007). Figure 3a [adapted from Donelan et al. (2004)] shows drag coefficients measured in the laboratory with equivalent 10-m wind speed up to 53 m s\(^{-1}\). One interesting feature shown in Donelan et al. (2004) is that the drag coefficient does not continue to increase with wind at high wind speed. It shows a remarkable “saturation” of the drag coefficient once the wind speed exceeds 33 m s\(^{-1}\). Beyond this speed the surface simply does not become any rougher in an aerodynamic sense. In the range of wind speeds of 10–26 m s\(^{-1}\) these laboratory measurements parallel the open-ocean measurements of Large and Pond (1981) but are a little lower. The 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. This corresponds to a roughness length of 3.35 mm. An adjustment to agree with Large and Pond (1981) would suggest saturation at 0.0028. Donelan et al. (2004) attributed a change in flow characteristics, leading to saturated aerodynamic roughness, to a flow separation due to continuous wave breaking where the flow is unable to follow the wave crests and troughs, as shown in Reul et al. (1999). The saturation behavior of the drag coefficient is also observed by Powell et al. (2003) using wind profiles from the airborne GPS dropwindsonde data in hurricanes.

In a recent study, Jeong et al. (2012) conducted a set of experiments using the same laboratory environment as in Donelan et al. (2004) to observe the enthalpy flux in high-wind conditions. They have shown that the exchange coefficient of enthalpy flux is relatively invariant with wind speed (Fig. 3b). The enthalpy exchange coefficient measured from the laboratory in Jeong et al. (2012) is very similar to the results from Liu et al. (1979) and Fairall et al. (2003) for wind speeds less than 15 m s\(^{-1}\) and similar to Drennan et al. (2007) for wind speeds from 15–30 m s\(^{-1}\). The observed \(C_D\) from Donelan et al. (2004) and \(C_K\) from Jeong et al. (2012) are used for the coupled model development and evaluation in this study, which will be shown in the next section.

### 4. Effects of directional wind–wave coupling on Hurricane Frances

The new directional wind–wave coupling parameterization is tested in the fully coupled modeling system UCMCM-WWP described in section 3. To isolate the effects of the surface waves from that of upper-ocean circulation including SST, we conducted three separate model simulations for each storm using the UCMCM-WWP, including the uncoupled atmosphere (UA; i.e., uncoupled MM5), coupled atmosphere–ocean (AO; coupled MM5–3DPWP), and fully coupled atmosphere–wave–ocean (AWO; coupled MM5–WW3–3DPWP) model configurations. All model simulations are done using the same multinested MM5 configuration with the highest grid resolution in the innermost domain at 1.67 km so that the model is able to capture the inner-core structure of the hurricane. The hurricane-induced ocean surface waves, upper-ocean circulations, and the effects of the full coupling on three Atlantic hurricanes—Bonnie (1998), Floyd (1999), and Frances (2004)—are described in a companion paper by S. Chen et al. (2013,
unpublished manuscript) in which the storm track, structure, and intensity are examined in detail. Here we focus on the results showing the relationship of the surface wind, waves, and stress, as well the impact of the directional wind–wave coupling on the surface wind in coupled model–simulated Hurricane Frances compared to that without the waves.


Hurricane Frances formed over the eastern Atlantic on 27 August 2004 and intensified into a major category 4 hurricane on 30 August. It eventually made landfall over the eastern coast of south Florida on 4 September. Figure 4 shows the model-simulated storm tracks in comparison with the National Hurricane Center (NHC) best-track data. The tracks from UA, AO, and AWO simulations are similar, except they diverged before landfall. The fully coupled AWO produced a track that is the closest to the best track in terms of timing, whereas UA and AO are 3–6 h slower than the best track (Fig. 4).

The model-simulated storm intensities (Fig. 5) are far more diverse and complex than the tracks. The uncoupled MM5 produced a much-lower minimum sea level pressure (MSLP) than the coupled AO and AWO simulations, which means that the uncoupled model overintensified because of the unrealistic shear force from the ocean with a constant SST. However, the maximum wind speeds (MWS) in the uncoupled model are much lower than the coupled models, indicating underintensification in terms of MWS. This contradiction in MSLP and MWS indicates an incorrect pressure–wind relationship in the uncoupled atmospheric model. This issue has been discussed in detail in Chen et al. (2013). By including the storm-induced cooling in AO, the MSLP is improved, but not the MWS. The fully coupled AWO produced the best MWS with MSLP similar to that of AO. The only difference between AO and AWO is the effect of wind–wave coupling that seems to be responsible for the improvement in the coupled model–simulated surface wind fields. More detailed discussion on this will be given in the following sections.

b. Ocean surface waves

Model-simulated ocean surface waves in hurricanes have been compared with observations from the National Data Buoy Center (NDBC) buoys and directional wave spectra from the National Aeronautics and Space Administration (NASA) Scanning Altimeter (SRA) on board the NOAA WP-3D aircraft during hurricane research flights. A detailed description of the airborne surface wave observations using SRA can be found in Wright et al. (2001). A recent study by Moon et al. (2003) compared the WW3 simulation of surface waves in Hurricane Bonnie (1998) with observations by Wright et al. (2001) and Walsh et al. (2002). Using the coupled AWO, Zhao and Chen (2005) found that the coupled model reproduces both the significant wave height (SWH) and wave periods near the U.S. coasts before and during the hurricane passages over the 5-day forecast period compared with the NDBC buoy data, as well as the observed directional wave spectra around the hurricanes over the open water by SRA in both Bonnie (1998) and Floyd (1999).

The storm-induced surface wave field is asymmetric around a hurricane as shown in observations (e.g.,
Wright et al. 2001). This is clearly evident in the AWO model–simulated SWH and wavelength in Hurricane Frances (Fig. 6). The highest surface waves, as measured by SWH, are usually observed in the right and front right of the hurricane (Fig. 6a), similar to that in Bonnie (Wright et al. 2001). Frances was heading west-northwest at 1200 UTC 31 August 2004. In a moving storm, the surface waves to the right of the storm center tend to grow into long and high waves because of the relatively long fetch compared to those to the left side of the storm (Fig. 6b). In contrast, the wind and the dominant waves are not aligned locally as shown in the directional wave spectra in Fig. 7, especially in the front left where the wind and waves are in different directions. The fully coupled AWO model simulation of the directional wave spectrum is compared with the SRA observations in Frances on 1 September 2004 (Fig. 7). The directional wave spectra observed by the airborne SRA in Hurricane Frances show features that are similar to other major hurricanes over the open ocean (e.g., Wright et al. 2001). In the rear-left quadrant of the hurricane, the directional wave spectrum is most complex with multiple spectrum peaks in both downwind and crosswind directions. The dominant wavelength is shorter than in the front-right quadrant (Fig. 7).

c. Directional wave stress in hurricanes

To examine the directional wave stress and its impact on the winds in Hurricane Frances, the wind and stress vectors are shown in Fig. 8. For a reference, the propagation directions of the mean wind-sea waves are also overlaid to compare with wind and stress vectors. Because the skin stress is aligned with the wind, the directional change in the total stress is due to the wave stress. The wave-induced stress varies in directions different from that of the wind. The storm motion is indicated in Fig. 6. The largest difference between the wind and stress vectors is in the front-left quadrant where the dominant waves propagate at about 90° from the wind direction as shown in Figs. 6 and 8. The wind and stress are mostly aligned on the right side of the storm (Fig. 8). The mean wind-sea wave vectors are to the right (outward from the storm center) of the wind vectors. The stress vectors are in between the wind and the wind-sea waves.

The relatively large difference between the wind and stress vectors in the front-left quadrant can result in a significant crosswind stress. The directional difference between the wind and stress can be as large as 25°, mostly positive on the right side of the storm and negative in the front-left quadrant (Fig. 9a). The fraction (percent) of the total wave stress in the crosswind direction is shown in Fig. 9b. The maximum value of the crosswind stress is close to 25% of the total wave stress, which can also be thought of as a reduction to the stress along the wind direction.

It is clear that the characteristics of the winds and waves vary significantly in different locations around the storm. The distributions of stress with wave frequency for pure wind seas of various ages are given in Donelan (1998). In general the stress distribution is centered on 2–4 times the peak frequency. The contributions to the wave stress by various wavelengths are shown in

![Fig. 6. The AWO model–simulated (a) SWH (color, m) and mean wave propagation direction (white vectors) and (b) mean wavelength (color, m) and surface wind (black vectors) at 0000 UTC 1 Sep 2004. The black plus sign indicates the storm center of Hurricane Frances. The arrow in the lower-left corner indicates the direction of the storm motion.](image-url)
a detailed analysis of the distribution of the wave energy spectrum as a function of wavenumber. Figures 10 and 11 show two examples from the front-right and front-left of the storm as indicated in Fig. 9 by A and B, respectively. There is significant spatial variability around the hurricane. The very long waves (swell) are mostly in the front of the moving hurricane (Fig. 7) and propagate in the direction of the storm motion as expected (Figs. 10a and 11a). They are in about the same direction as the wind in the front right of the hurricane, whereas almost directly against the wind in the front left. Swell dominates the energy spectrum, but the wind sea accounts for most of the stress. Both long and short waves contribute to the stress and the wind and the dominant waves are aligned (Fig. 10b) on the left side. In contrast, the long waves near the peak frequency do not contribute much to the stress when they propagate at large angles to the wind. Instead, they account for almost all of the crosswind stress (Fig. 11b). The shorter waves contribute more to the stress on the left side than on the right side of the hurricane.

d. Air–sea exchange coefficients

Based on the results shown above, it is expected that the directional wind–wave coupling will have an impact on the drag coefficient (and momentum flux). As a reminder, \( z_t \) and \( z_q \) are computed using the Garrat formulation (Garratt 1992) in uncoupled and coupled model simulations; they are not directly affected by the wind–wave coupling. However, they are indirectly affected by the coupling through the winds in AWO. The enthalpy exchange coefficient is computed from the heat and moisture fluxes from the model. The drag coefficient is computed diagnostically from the total stress and wind fields. Figure 12 shows \( C_D \) and \( C_K \) in Hurricane Frances from the uncoupled and fully coupled AWO simulations. There is a significant spatial variability of \( C_D \) in the hurricane, which is asymmetric around the storms owing mostly to the variation in surface waves in the fully coupled AWO (Fig. 12a), whereas \( C_D \) has a similar pattern as the wind speed, which is more symmetric around the storm center, in the uncoupled MM5 simulation (Fig. 12b). The uncoupled simulation has the largest \( C_D \) value.
(a function of wind speed through the Charnock relationship) in the front-right quadrant. In AWO, in contrast, \( C_D \) is slightly smaller in the front quadrants than the rear quadrants where the wavelength is shorter. It is also noteworthy that \( C_D \) has a relative minimum in the front left where the crosswind stress is the largest (Fig. 9).

The \( C_K \) values from model simulations varies only slightly from 1.1 to \( 1.2 \times 10^{-3} \) (Figs. 12c and 12d), which is similar to the observations from Jeong et al. (2012) for wind speeds up to 40 m s\(^{-1}\) shown in Fig. 3b. These results are also consistent with observations from Liu et al. (1979) representative of wind speeds less than 20 m s\(^{-1}\), as well as the CBLAST-Hurricane observations by Drennan et al. (2007) using airborne turbulence measurements in wind speeds up to 32 m s\(^{-1}\) in Hurricane Fabian (2003). They found that \( C_K \) changes little with wind speed and remains on average at about \( 1.1 \times 10^{-3} \). They conclude that the ratio \( C_K/C_D \) therefore varies spatially in a hurricane owing mostly to the variation in \( C_D \).

Similarly, \( C_K/C_D \) varies from 1 in the outermost part of the storm where wind speeds are less than 20 m s\(^{-1}\) to 0.4 in the eyewall region where the wind speed reaches 50–60 m s\(^{-1}\) in the fully coupled AWO simulation (Fig. 13a). In the uncoupled simulation the ratio is 0.3 or less in the eyewall region—similar to the results reported by Bao et al. (2002). Hurricane Frances continues to intensify even while the ratio was much less than 1 in the inner-core region consistently with time. The laboratory measurements of Donelan et al. (2004) and Jeong et al. (2012) have shown that the ratio is about 0.3–0.4 for wind speeds up to 40 m s\(^{-1}\). The CBLAST observations have shown that \( C_K/C_D \) is about 0.5–0.7 for wind speeds from 20 to 30 m s\(^{-1}\) (e.g., Black et al. 2007). The coupled AWO model simulation produced about the same range of this ratio compared with the observations (Figs. 13c and 13d). Note that the \( C_K/C_D \) from both the observations and the high-resolution, full-physics coupled model are less than what has been predicted by the idealized model used in Emanuel (1995). One possible explanation for the discrepancy is that Emanuel (1995) used an axisymmetric hurricane model with a bulk atmospheric boundary layer (ABL) parameterization in which the gradient winds are used.
in the calculation of exchange coefficients. Furthermore, the depth of the ABL in hurricanes decreases toward the storm center, which can be resolved in the high-resolution, full-physics models, but not in the bulk ABL in Emanuel (1995). The gradient wind at the top of the ABL, as in Emanuel’s model, is not representative of the surface wind.

FIG. 10. (a) Fully coupled AWO model—simulated 2D wave spectrum located in the rear-right quadrant (indicated by “A” in Fig. 8) and (b) (top) cumulative stress, (middle) direction of wave stress, and (bottom) power spectrum as functions of wavenumber in Hurricane Frances. The long-wave spectrum from WW3 and tail spectrum are shown in dark blue and cyan, respectively. The horizontal red line indicates the direction of the wind and the vertical magenta line indicates the peak wavenumber.

FIG. 11. As in Fig. 10, but for the point located in the front-left quadrant as indicated by “B” in Fig. 9.
e. Wind–wave coupling and storm structure

To further examine the impact of the directional wind–wave coupling on hurricanes, we compare the observed and simulated storm structures. Figures 14a and 14b show the airborne Doppler radar–observed wind speed at 3-km level in Hurricane Frances from two consecutive flights of the NOAA WP-3D aircraft during the CBLAST-Hurricane field campaign on 30 and 31 August 2004. During the period of these 2 days, Frances went through an eyewall replacement cycle that was characterized by a well-developed concentric eyewall with a primary inner eyewall and a secondary outer eyewall as shown on the airborne Doppler winds on 30 August (Fig. 14a). The secondary eyewall eventually overtakes over the inner eyewall a day later on 31 August (Fig. 14b), which may be similar to that of Hurricane Rita (2005) observed during the Hurricane Rainband and Intensity Change Experiment (RAINEX) as shown in Houze et al. (2007). Note that the “missing” value of the wind speed in the Doppler radar data is a result of the very low reflectivity (non-precipitation) regions in the eye and the “moat” region between the two concentric eyewalls as described in Houze et al. (2007). The corresponding model-simulated wind speeds at 3-km level at 1800 UTC 30 and 31 August 2004 are shown in Figs. 14c–h. The fully coupled AWO simulation captured the concentric eyewalls on 30 August (Fig. 14c), whereas the uncoupled MM5 and the coupled AO without wind–wave coupling did not (Figs. 14e and 14g). One possible reason for AWO to produce better concentric eyewalls may be attributed to the difference in radial profiles of CD, especially in the outer rainband region and inflow (Lee and Chen 2012). The model simulations reproduced the evolution of Hurricane Frances for 30–31 August as the eyewalls evolved into a large and stronger eyewall on 31 August (Figs. 14b,d,f,h). The observed asymmetry from the north–south direction in the wind field is also captured in the model simulations. However, the model-simulated inner eyewall is larger than the observation in all cases as shown in the azimuthally averaged wind speeds.
The uncoupled model overestimated the winds over a large area at the 3-km level compared to the Doppler radar observations (Fig. 15b). However, it underestimates the surface wind at the same time (Fig. 5), which again indicates the problem of unrealistically large surface stress at high wind speeds in the uncoupled model.

The surface stress vector can influence not only the surface wind speed but also the wind direction, which may be an important factor contributing to the characteristics of the near-surface inflow in hurricanes. Powell (1982) provided one of the most detailed surface wind analyses using surface observations in Hurricane Frederic (1979) over the Gulf of Mexico. The observed surface inflow angle is as large as 45° or greater in the southeastern quadrant (Fig. 16a). In comparison, the model-simulated inflow angles in Hurricane Frances (2004) are shown in Figs. 15b–d. The fully coupled AWO model increases the inflow angle in the rear-left quadrant to values greater than 45° (shaded area in Fig. 16b) comparable to the observation from Powell (1982). The inflow angles in the uncoupled and coupled AO models in the same quadrant are smaller. The impact of the directional wind–wave coupling is asymmetric around a storm, as it increases the inflow angle mainly in the rear left and decreases it in the front right. This is consistent with the values shown in the analysis of observations in Hurricane Frederic in Powell (1982). The overall azimuthally averaged inflow angle outside the eyewall is 3°–7° larger in AWO than in the uncoupled and AO simulations in Hurricane Frances (Fig. 17).

5. Conclusions

A new directional wind–wave coupling method using the wave-stress vector from two-dimensional wave spectra has been developed and tested in the University of Miami Coupled Atmosphere–Wave–Ocean Model (UMCM). The effects of the directional wind–wave coupling are examined in comparing three model experiments of uncoupled atmosphere (UA), coupled atmosphere–ocean (AO), and fully coupled atmosphere–wave–ocean.
FIG. 14. (a),(b) The NOAA WP-3D airborne Doppler radar–observed wind speed at 3-km level, and (c),(d) the corresponding model-simulated winds in Hurricane Frances from AWO, (e),(f) AO, and (g),(h) uncoupled model simulations at (left) 1800 UTC 30 Aug and (right) 1800 UTC 31 Aug 2004. Arrows indicate the directions of storm motions in observations and models.
(AWO) models. The model-simulated track, intensity, and storm structure in Hurricane Frances (2004) are compared with observations. The coupling to the ocean circulation model improves the storm intensity by including the storm-induced cooling in the upper ocean and SST, whereas the uncoupled atmosphere model with a constant SST overintensifies the storm in terms of MSLP (Fig. 5). However, without coupling to the

![Fig. 15](image_url)

**Fig. 15.** Azimuthally averaged radial profiles of the wind speed from the airborne Doppler radar and the three model simulations shown in Fig. 14 at (a) 1800 UTC 30 Aug and (b) 1800 UTC 31 Aug 2004.

![Fig. 16](image_url)

**Fig. 16.** (a) Observed surface streamline and inflow angle in Hurricane Frederic (1979) [Adapted from Powell (1982)]. Surface inflow angle in (b) AWO, (c) AO, and (d) UA simulations of Hurricane Frances (2004). The shaded area in (b)–(d) indicates inflow angles greater than 45°. The directions of storm motions in model simulations are shown in Fig. 14.
surface waves explicitly, both the uncoupled atmospheric model UA and the coupled atmosphere–ocean model AO underestimate the surface wind speed, even though the MSLP of especially the AO coupled model is close to the observed values. The full coupling with the directional wave–wind parameterization clearly improves the model-simulated surface wind in Hurricane Frances; not only is the wind speed improved, but also the inflow angle (Fig. 16), which is important for the overall storm evolution and structure.

Observations from the CBLAST-Hurricane field program provided a unique dataset to evaluate and validate the fully coupled models. The high-resolution, fully coupled model is capable of capturing the complex hurricane structure and intensity change in Hurricane Frances (2004). Fully coupled AWO simulation reproduced the concentric eyewalls and eyewall replacement observed in Hurricane Frances (Fig. 14), whereas the uncoupled UA and partial coupled AO without wind–wave coupling did not.

Hurricane-induced surface waves are highly asymmetric around the storm as observed by SRA on board the NOAA WP-3D aircraft in Frances, which is well simulated by the fully coupled AWO (Fig. 7). Directional wind–wave coupling can represent the complex features in wave stress (form drag) varying significantly from the right side of the storm where the wind and dominant waves are aligned to the left side where they are misaligned (Fig. 8). Both long (swell) and short (wind sea) waves contribute to the stress when the wind and the dominant waves are aligned (Fig. 10). In contrast, the long waves near the peak frequency do not contribute much to the stress when they propagate at large angles to the wind. Instead, they account for almost all of the crosswind stress (Fig. 11). The shorter waves contribute more to the stress on the left side than on the right side of the hurricane.

The directional wind–wave coupling parameterization has been implemented recently in UMCM-WMH (fully coupled WRF–UMWM–HYCOM; see section 2). Further investigation on the impact of directional wind–wave coupling has been ongoing. Comparisons of uncoupled WRF simulation using wind speed–dependent $C_D$ parameterization based on Donelan et al. (2004) with the fully coupled model result will be described in a forthcoming paper.

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