Numerical Simulations of Air–Sea Interaction under High Wind Conditions Using a Coupled Model: A Study of Hurricane Development

J.-W. Bao
Cooperative Institute for Research in Environmental Sciences, University of Colorado, and NOAA/Environmental Technology Laboratory, Boulder, Colorado

J. M. Wilczak
NOAA/Environmental Technology Laboratory, Boulder, Colorado

J.-K. Choi and L. H. Kantha
Department of Aerospace Engineering Sciences, Colorado Center for Astrodynamics Research, University of Colorado, Boulder, Colorado

(Manuscript received 30 March 1999, in final form 18 August 1999)

ABSTRACT

In this study, a coupled atmosphere–ocean wave modeling system is used to simulate air–sea interaction under high wind conditions. This coupled modeling system is made of three well-tested model components: The Pennsylvania State University–National Center for Atmospheric Research regional atmospheric Mesoscale Model, the University of Colorado version of the Princeton Ocean Model, and the ocean surface gravity wave model developed by the Wave Model Development and Implementation Group. The ocean model is initialized using a 9-month spinup simulation forced by 6-hourly wind stresses and with assimilation of satellite sea surface temperature (SST) and altimetric data into the model. The wave model is initialized using a zero wave state. The scenario in which the study is carried out is the intensification of a simulated hurricane passing over the Gulf of Mexico. The focus of the study is to evaluate the impact of sea spray, mixing in the upper ocean, warm-core oceanic eddies shed by the Gulf Loop Current, and the sea surface wave field on hurricane development, especially the intensity.

The results from the experiments with and without sea spray show that the inclusion of sea spray evaporation can significantly increase hurricane intensity in a coupled air–sea model when the part of the spray that evaporates is only a small fraction of the total spray mass. In this case the heat required for spray evaporation comes from the ocean. When the fraction of sea spray that evaporates increases, so that the evaporation extracts heat from the atmosphere and cools the lower atmospheric boundary layer, the impact of sea spray evaporation on increasing hurricane intensity diminishes.

It is shown that the development of the simulated hurricane is dependent on the location and size of a warm-core anticyclonic eddy shed by the Loop Current. The eddy affects the timing, rate, and duration of hurricane intensification. This dependence occurs in part due to changes in the translation speed of the hurricane, with a slower-moving hurricane being more sensitive to a warm-core eddy. The feedback from the SST change in the wake of the simulated hurricane is negative so that a reduction of SST results in a weaker-simulated hurricane than that produced when SST is held unchanged during the simulation. The degree of surface cooling is strongly dependent on the initial oceanic mixed layer (OML) depth. It is also found in this study that in order to obtain a realistic thermodynamic state of the upper ocean and not distort the evolution of the OML structure during data assimilation, care must be taken in the data assimilation procedure so as not to interfere with the turbulent dynamics of the OML.

Compared with the sensitivity to the initial OML depth and the location and intensity of the warm eddy associated with the loop current, the model is found to be less sensitive to the wave-age-dependent roughness length.

1. Introduction

The atmosphere and the ocean are coupled dynamically and thermodynamically by momentum and enthalpy exchanges at the air–sea interface. The interaction of the atmosphere and the ocean has been long recognized as an important element of atmospheric and oceanic circulation on a wide range of scales. Although numerical modeling of the atmosphere and the ocean as separate geophysical fluid systems has made tremendous progress over the past four decades, significant uncertainties remain in how these two systems influence one another. These uncertainties concern the physical pro-
cesses that are important for coupling the atmosphere through the transfer of momentum, heat, and moisture across the air–sea interface.

One manifestation of the interaction between the atmosphere and the ocean is ocean surface gravity waves, which can have a significant impact on the transfer of momentum and enthalpy across the air–sea interface (see, e.g., Geernaert 1990; Kraus and Businger 1994). Conventionally, in atmospheric modeling the effect of ocean waves on flux transfer at the air–sea interface is taken into account by using an average roughness length that is linked to the surface stress through the Charnock relationship (see, e.g., Garratt 1992; Kraus and Businger 1994). The surface stress is taken to be a function of only wind speed, the roughness length, and the stability of the surface layer above the air–sea interface. Recently, both observations and numerical modeling studies indicate that the surface stress is also a function of sea state; in other words, it is dependent on the wind wave spectrum (see, e.g., Komen et al. 1994). Part of the momentum imparted by the wind stress is propagated away by surface gravity waves instead of driving local currents. Ocean waves consist of a distribution of moving elements that are related to the wind stress, which together determine the rate of input of energy from wind to waves. The nonlinear, resonant, wave–wave interaction mechanism redistributes energy in the wave spectrum. Despite the complexity, advances have been made over the past two decades in understanding and modeling these surface wave processes (Komen et al. 1994).

Another manifestation of air–sea interaction is the change of sea surface temperature (SST) due to wind-driven mixing in the upper ocean, and to the heat, moisture, and radiative fluxes at the air–sea interface. The dynamical response of atmospheric models can be very sensitive to temporal changes in the thermodynamic fluxes at the air–sea interface, which in turn can be sensitive to SST changes. For example, idealized simulations of tropical cyclone–ocean interaction within a coupled atmosphere–ocean model (Bender et al. 1993) indicate that the cooling of the sea surface produced by the tropical cyclone results in a significant reduction of the hurricane intensity. At the other extreme, air–sea fluxes under low wind conditions are sensitive to variations of SST induced by the cool skin and the dynamics of the diurnal warm layer at the air–sea interface (Fairall et al. 1996).

SST changes are strongly dependent on turbulent mixing in the upper ocean. Both the surface momentum flux and an upward enthalpy flux across the interface will produce turbulent mixing in the upper ocean, resulting in an ocean mixed layer (OML) within which horizontal velocity, temperature, and salinity are nearly constant with depth. The OML can have large variability on a wide range of temporal and spatial scales resulting either from changes in atmospheric forcing or from changes in the ocean circulation. Because all oceanic changes are communicated to the atmosphere through the OML, any numerical model intended to simulate air–sea interaction must include an accurate description of ocean mixed layer dynamics.

Flux transfer processes over the ocean are commonly parameterized in atmospheric models using Monin–Obukhov similarity theory and a specification of the surface characteristics (e.g., surface roughness length). Recent observations and modeling studies indicate that the sea spray generated by ejection of droplets from breaking waves should be taken into account in the air–sea flux parameterization used in atmospheric models, especially under high wind conditions (see Andreas et al. 1995; Kepert et al. 1999). High winds produce sea spray droplets that modify the mean thermodynamic state of the air surrounding them. As a result, the physics involved in enthalpy transfer across the air–sea interface is different from that in situations where the sea spray is absent. If the lowest model level is well above the droplet evaporation zone, the effects of the sea spray may be included as a modification of the surface fluxes that can be applied within the Monin–Obukhov similarity framework. If, however, the lowest model level is within the droplet evaporation zone, a modification of Monin–Obukhov similarity theory may be necessary.

Although it is believed that the surface fluxes can be strongly altered by intensive mixing within the OML, by sea spray, or by sea surface waves, the collective impact of these processes on air–sea interaction has not been assessed in field programs to the same degree as has each of these processes individually. This is due in part to the difficulty of making simultaneous observations of the momentum, and sensible and latent heat fluxes across the air–sea interface along with measurements of all the atmospheric and oceanic parameters affecting the fluxes. Therefore it is appealing to use a numerical model to evaluate the combined impact of OML dynamics, sea spray, and sea surface waves on the surface fluxes. Obviously it is a basic requirement that in the model, air–sea interaction be treated in a two-way fashion, with mutual feedback from the ocean and the atmosphere. Numerical evaluations such as this, although they cannot replace observational studies of air–sea interaction, can indicate the potential importance of the parameterized physical processes as well as highlight processes for which additional observational field studies would be useful.

This study describes numerical simulations in which a limited-area, coupled air–sea modeling system is used to evaluate the impact of air–sea interaction on the development of a hurricane. The modeling system consists of a regional atmospheric model, an ocean surface gravity wave model, and an ocean circulation model that are directly coupled at each integration time step. The purpose of these numerical experiments is to assess the impact of mixing in the OML, sea spray, and sea surface waves on air–sea interaction under high wind conditions.
The methodology used for carrying out the numerical experiments is similar to that of Bender et al. (1993). All the experiments use the same atmospheric boundary conditions and hurricane vortex initialization so that the atmospheric environmental conditions that predominantly control the hurricane track remain constant. The initial and boundary conditions are prescribed by analyzed fields for Hurricane Opal (1995). The model domain contains realistic geographic features such as coastlines, the bathymetry of the Gulf of Mexico, and land topography.

Observations suggest that hurricane development may be strongly affected by warm oceanic eddies shed by the Loop Current in the Gulf, which possesses a relatively deep and warm OML (Shay et al. 1998). In this study, a realistic oceanic warm eddy is introduced into the numerical experiments through assimilation of satellite data in order to evaluate how the existence of the warm eddy affects the intensification of the hurricane.

It should be noted that it is not the purpose of this study to explore the details of the internal dynamics of the hurricane or to investigate the interaction of the large-scale atmospheric environment with the hurricane. Because the purpose of this study is to evaluate the impact of mixing in the OML, sea spray, and sea surface waves in numerical simulations of storm evolution, idealized experiments are used. Liu et al. (1997) have shown that in order to simulate reasonably well the intensity and inner-core structures of hurricanes, not only is it required that the atmospheric model have high resolution (~6 km) and realistic cloud physics, but it is also necessary that the hurricane vortex (depth, size, and intensity) be properly initialized in relation to large-scale atmospheric conditions. Their results also indicate that an apparently realistic simulation of hurricane evolution can be obtained even if the model does not have a physically faithful description of fluxes at the air–sea interface, so long as the amount of momentum and enthalpy fluxes from the model’s lower boundary fortuitously is sufficient to sustain the simulated hurricane evolution.

The paper is organized as follows. Brief descriptions of individual model components in the coupled modeling system are presented in section 2. Parameterizations used for the surface roughness lengths and turbulent fluxes are described in section 3. A parameterization scheme is then introduced in section 4 that takes into account sensible and latent heat fluxes contributed from sea spray droplets. This is followed in section 5 with a discussion of the coupling procedure for all the model components. The assimilation procedure for the satellite data and the numerical experiments are presented in section 6, followed by a discussion and summary in section 7.

2. Model description

The Pennsylvania State University–National Center for Atmospheric Research (Penn State–NCAR) Mesoscale Model version 5 (MM5) (Grell et al. 1994) is a regional, hydrostatic or nonhydrostatic, sigma-coordinate model designed to simulate or predict mesoscale and regional-scale atmospheric circulation. It has been developed at Penn State and NCAR as a community mesoscale model and is continuously being improved by contributions from users at numerous universities and government laboratories. The model has a variety of resolvable-scale microphysics and subgrid-scale cumulus parameterization schemes for precipitation physics, along with several options for the parameterization of planetary boundary layer and surface-layer processes. The model also includes three well-tested atmospheric radiation schemes.

The ocean wave model (WAM) is a third-generation wave model (WAMDI Group 1988) which solves the wave transport equation explicitly without any prior assumptions about the shape of the spectrum. It represents the physics of the wave evolution in accordance with today’s knowledge for the full set of degrees of freedom of a 2D wave spectrum. WAM describes the evolution of the directional wave spectrum by solving the wave energy equation. The forcing in the wave energy equation includes the wind input and dissipation terms from the quasi-linear wind-wave generation theory (Janssen 1991; Komen et al. 1994), in which the wind input term includes the square of the inverse of the wave age (defined as \(c/u^*\), \(c\) being the wave phase speed and \(u^*\) the friction velocity), and the dissipation term is proportional to the fourth power of the frequency. The nonlinear wave–wave interaction term in the forcing is based on the theory of Hasselmann et al. (1985). The model was developed at the Max Planck Institute for Meteorology in Hamburg, Germany, has been installed at many institutions worldwide, and is used for both research and operational applications. It is also being applied for interpretation and assimilation of satellite wave data. So far, four cycles of the wave model have been issued. It is the last cycle, cycle 4 (see Gunther et al. 1992; Komen et al. 1994), that is coupled to MM5.

The Princeton Ocean Model (POM) is used in the coupled modeling system to simulate the OML processes and oceanic circulation. POM is a sigma-coordinate, free-surface, primitive equation ocean model, which includes a turbulence submodel. POM uses the following three assumptions: 1) the flow is incompressible; 2) the vertical accelerations are negligible compared to gravity, thus, the vertical pressure distribution satisfies the hydrostatic approximation; and 3) density is approximated by its mean value except when multiplied by gravity, that is, when the Boussinesq approximation is used. POM was developed in the late 1970s by G. Mellor’s group (Blumberg and Mellor 1987; Kantha and Piacsek 1996) at Princeton University, with subsequent contributions from many other institutions. The model has been applied to modeling of estuaries, coastal regions, and open oceans, and is being used routinely by many worldwide institutes for both research and op-
3. Sea surface roughness and turbulent flux parameterization

The transfer of momentum flux across the so-called wave boundary layer is described by the bulk parameterization based on the Monin–Obukhov similarity theory, in which the stress, \( \tau \), is related to the resolved wind via the drag coefficient, \( C_d \). By definition, the drag coefficient is fully determined by the roughness length of the ocean surface. Data from field and laboratory experiments indicate that the roughness length of the ocean surface depends upon the stage of wind–wave development. In the coupled modeling system, the following parameterization is used to describe the dependence of the drag coefficient on the wave state.

Assume that the total stress close to the surface can be expressed as the sum of contributions from surface wave and turbulence: \( \tau = \tau_w + \tau_r \), where \( \tau \), \( \tau_w \), and \( \tau_r \) are the total stress, wave-induced stress, and atmospheric turbulence–induced stress, respectively. The mean wind profile above the wave surface is assumed to be logarithmic. Following Janssen (1991), we assume that the effective roughness length of gravity–capillary waves can be modeled by means of the Charnock relation

\[
\zeta = z_0 = \alpha \frac{\tau}{\rho_s g}, \tag{1}
\]

where \( \alpha \) is a constant, \( g \) the acceleration of gravity, and \( \rho_s \) the air density. The effective wave-induced roughness length of the developing gravity waves is modeled by a parametric height, \( z_1 \), so that the mean wind profile under neutral stratification is given by

\[
U(z) = \frac{\sqrt{\tau / \rho_s} \ln \frac{z + z_1}{z_0 + z_1}}{\kappa}, \tag{2}
\]

where \( \kappa \) is the von Kármán constant. Consider the steady-state stress balance of airflow over sea waves. Following Janssen (1991), the atmospheric turbulence–induced stress at \( z = z_0 \) is given by

\[
\tau(z_0) = \left( \frac{z_0}{z_0 + z_1} \right)^2 \tau_r, \tag{3}
\]

in which it is assumed that \( \tau_r \) is modeled by

\[
\tau_r(z) = l \frac{\partial U(z)}{\partial z} \frac{\partial U(z)}{\partial z}, \tag{4}
\]

where \( U(z) \) is the wind profile given by (2) and \( l \) is the mixing length (given by \( \kappa z \) for neutral stratification). Substituting (3) into the expression of the total stress and using (1) then gives

\[
z_1 = \frac{\alpha \tau \ln \frac{1}{1 - \tau_r / \tau}}{g \rho_s \sqrt{1 - \tau_r / \tau}} = 1. \tag{5}
\]

With the above parameterization of the sea-state-dependent roughness length, the coupling of MM5 and WAM takes place in the following sequence. First, the total stress, \( \tau = \mu w^* \), is calculated in MM5 using a stability-dependent scheme based on the Monin–Obukhov similarity theory,

\[
u = \frac{\kappa U}{\ln(z/z_0) - \psi_w}, \tag{6}
\]

where \( \psi_w \) is a nondimensional stability correction factor. The total stress is then entered into WAM to obtain \( \tau_w \). Then both \( \tau_w \) and \( \tau \) are used to compute the new sea-state-dependent \( z_1 \); the sum of \( z_0 \) and \( z_1 \) is then used at the next time step in MM5 as the new roughness length to obtain the new total stress \( \tau \).

In recent experiments of ocean wave models that are coupled to atmospheric models (e.g., Doyle 1995), the wave model uses the wind at the lowest level of the atmospheric model to approximate the 10-m wind. The wave model then makes a stability-independent estimate of the effective roughness length \( z_1 \), total stress, and wave-induced stress (which drives the wave field). Only the effective roughness length is then used by the atmospheric model, which calculates its own stability-dependent stress for use in the atmospheric model. This results in a total stress that is not continuous across the air–sea interface. This approach to coupling is improved with the aforementioned approach based on (1)–(5), which for neutral stratification will now produce the same total stress in the atmospheric and ocean models. Although this parameterization is derived for neutral stratification, it is easy to show that the equation for \( z_1 \) holds under nonneutral stratification if the wind profile follows Monin–Obukhov similarity and can still be expressed as

\[
\frac{\kappa(z + z_1)}{\sqrt{\tau / \rho_s}} \frac{\partial U(z)}{\partial z} = \frac{\kappa}{l} = \varphi(z/L), \tag{7}
\]

where \( \varphi \) is the well-known Monin–Obukhov similarity function and \( L \) is the Monin–Obukhov length.

It has been recognized that the roughness length used for the surface momentum flux calculation need not be
the same as that used for the surface enthalpy flux. Strictly speaking, the roughness length obtained by using the parameterization obtained as the sum of (1) and (5) is only applicable to the momentum flux. Surface roughness lengths for heat and moisture could be obtained using a surface renewal model (e.g., Liu et al. 1979; Fairall et al. 1996), or using aerodynamically smooth scaling laws (Beljaars 1995), although for hurricane force winds there is little heat flux data to validate any parameterization choice. Recognizing that considerable uncertainty exists for the scalar roughness lengths, in the present simulations we simply use the same roughness length in the calculation of the surface thermal and momentum fluxes. The surface sensible and latent heat fluxes are then given by

\[ H_s = -\rho c_p \kappa u T_m \]  
\[ Q_l = -\rho L_m \kappa u q, \]

where

\[ T_m = \frac{\Delta T}{\ln(z/z_0)} - \psi_h \]  
\[ q = \frac{\Delta q}{\ln(\kappa u z/K_n + z/z_0)} - \psi_q, \]

where \( \Delta T \) and \( \Delta q \) are the air–sea temperature and moisture differences, \( \psi_h \) is a stability correction factor for thermal fluxes, and \( K_n \) is a background diffusivity (see Grell et al. 1994).

4. The spray contribution to sensible and latent heat fluxes from the ocean

When the wind speed is in excess of approximately 15 m s\(^{-1}\), a substantial amount of sea spray is produced by breaking waves, bursting bubbles, and wind gusts (e.g., Kraus and Businger 1994; Andreas et al. 1995). This results in the development of a near-surface layer that can be characterized as a droplet evaporation zone. Both theory and observations suggest that, at high wind speeds, evaporation from sea spray is significant. Above the droplet evaporation zone, the droplet contribution to the flux takes the form of an enhanced turbulent flux of water vapor, and a corresponding change in the turbulent flux of sensible heat that is dependent on the enhanced water vapor flux. Simple scaling models to parameterize the spray effect on sensible and latent heat fluxes have been developed based on recent observations (see, e.g., Andreas 1992, 1998; Fairall et al. 1994). The thermodynamic effect of sea spray in these models is characterized as a competition between their evaporation rate and their lifetime in the air. Two factors determine the lifetime of sea spray in the air: the fall velocity of the droplets and their vertical transport by turbulence. The evaporation rate of sea spray is controlled by the amount of heat available for evaporation and interactions between droplets of different sizes. The heat for the spray evaporation comes from two sources: 1) upward transfer of heat from the ocean by turbulence and the spray droplets, and 2) downward heat transfer from the atmosphere by MBL turbulence and, when cumulus convection exists, convective downdrafts. These heat sources adjust internally until some equilibrium is reached, with the spray evaporation acting as a moisture source and as a heat source or sink to the MBL. As a result, the MBL structure is adjusted accordingly and this adjustment feeds back on the dynamic processes above the MBL. Obviously, a coupled air–sea model with an accurate, well-verified MBL parameterization is ideal for investigation of the net effect of sea spray on air–sea interaction.

A bulk parameterization of the sea-spray-mediated sensible and latent heat fluxes based on Fairall et al. (1994) is used in the coupled modeling system. This parameterization includes three physical processes: 1) the cooling of spray droplets from sea surface temperature \( T_s \) to air temperature \( T_m \) that occurs by conduction of heat to the atmosphere, producing spray-sensitive heat flux \( Q_s \); 2) the cooling of spray droplets from \( T_s \) to the wet-bulb temperature \( T_w \) (corrected for salinity and droplet curvature effects) that occurs by evaporation, producing latent heat flux \( Q_l \); 3) additional evaporation of the spray droplets that occurs at the expense of cooling the atmosphere, producing additional latent heat fluxes \( Q_{ls} \). The first two processes occur rapidly (tenths of a second), while the timescale for the third process can be much longer.

From Fairall et al. (1994), droplet-mediated evaporation and sensible fluxes, \( Q_s \) and \( Q_l \), are described by

\[ Q_s = 3.0 \times 10^{-6} u^3 q(T_s) H(t) \]  
\[ \div \left[q(T_a) - q\right] \]

\[ Q_l = 2.7 \times 10^{-5} u^3 q(T_a) \]

where \( u \) is the 10-m wind speed (approximated by the wind speed at the lowest level of MMS), \( B \) is a parameter that is related to the fact that the sea spray droplets are at the wet-bulb temperature \( T_w \) varies from 0.59 at 273 K to 0.21 at 303 K), \( q \) the specific humidity, \( q_s \) the saturation specific humidity, \( H \) the turbulent latent heat flux, \( H_l \) the turbulent sensible heat flux, and \( Q_h \) is the total latent heat flux generated by droplet evaporation. We note that \( Q_s \) and \( Q_l \) are the droplet-mediated fluxes that would occur if the sea spray does not alter the normal logarithmic profile of mean \( q \) and \( T \) in the droplet evaporation zone. Fairall et al. (1994) argue that (12) and (13) represent upper limits, and that the actual spray-dependent fluxes will be reduced by a factor \( a \) due to the fact that mean profiles of \( q \) and \( T \) in the droplet zone do not remain logarithmic, but are modified by the presence of the spray. Based on Andreas (1998) reanalysis of his 1992 results, the actual spray-dependent fluxes can be a factor of 2 smaller than as

Since sea spray droplets evaporate at the expense of the sensible heat that they carry, and the sensible heat available in the surrounding air, the flux boundary conditions at the air–sea interface become

\[
total \text{ sensible heat flux } = H_s + \alpha Q_s - \alpha \beta Q_l \quad (14)
\]

and

\[
total \text{ latent heat flux } = H_l + \alpha Q_l, \quad (15)
\]

where \( \alpha \) is the profile-change feedback parameter and \( \beta \) is the evaporation partitioning parameter. Equations (14) and (15) are the same as given by Fairall et al. (1994), except for the inclusion of \( \beta \), which is described below. Because the sea spray parameterization is derived from data with wind speed less than 30 m s\(^{-1}\), Fairall et al. (1994) suggest that an upper bound of 30 m s\(^{-1}\) should be used in numerical model simulations of weather and climate. This upper bound is used in this study.

In this study, a simple ad hoc parameterization for \( \alpha \) (see Kepert et al. 1999) is used,

\[
\alpha = \frac{H_{tot}}{H_{tot} + Q_{tot}}, \quad (16)
\]

where \( H_{tot} = H_s + H_l \) and \( Q_{tot} = Q_s + Q_l \). The profile-change feedback parameter thus defined varies from 0 to 1. It limits the total production of water vapor from sea spray, reflecting the feedback effect of sea spray to reduce the evaporation.

The evaporation partitioning parameter \( \beta \) is defined as the ratio

\[
\beta = \frac{Q_{2}}{Q_{1} + Q_{2}}. \quad (17)
\]

For the range of wind speeds and droplet sizes considered by Fairall et al. (1994), it was assumed that \( Q_{2} \gg Q_{1} \), and \( \beta = 1 \). However, for extremely high wind speeds, the droplet size may be so large that they fall back to the ocean before further evaporation can extract heat from the atmosphere, in which case \( Q_{2} \) and \( \beta \) approach zero.

Andreas and Emanuel (1999) consider the case \( \beta = 0 \) by assuming that only a very small fraction (\( \leq 1\% \)) of the spray mass evaporates, just sufficient to bring the spray droplets to a new temperature, \( T_{eq} \), that is close to the wet-bulb temperature. In this case, the enthalpy of the air is increased in part by conduction of sensible heat from the spray droplets to the air when they undergo a temperature change from their initial temperature to the air temperature, together with the transfer of latent heat from the droplets to the atmosphere as the droplets temperature continues to fall to \( T_{eq} \). Using an idealized hurricane model, Andreas and Emanuel (1999) show that in this case sea spray evaporation can significantly increase hurricane intensity.

Although it has long been postulated that the evaporation of spray droplets influences the energy budget at the air–sea interface, the process of heat transfer mediated by spray droplets is still poorly understood due to the complex nature of the interactions between the droplets and the air–sea system. Numerical experiments carried out in this study indicate that different values of \( \beta \) lead to different impacts of sea spray on the enthalpy flux from the ocean to the atmosphere, and thus on the simulated hurricane development.

5. Coupling the atmospheric model with the ocean circulation model and the ocean wave model

Figure 1 shows schematically how MM5 is coupled with WAM and CUPOM at the time step when the coupling takes place. Each model computes its prognostic variables using its own prognostic equations and integration time steps. Because the computational stability conditions of the three models are different, the integration time steps used in both WAM and CUPOM are larger than that used in MM5 for the same horizontal resolution. In all the numerical experiments presented here, the horizontal resolutions of both WAM and CUPOM are no finer than that of MM5. Each time step of either WAM or CUPOM contains a few time steps of MM5. Therefore, variable passing required for the coupling takes place only when either WAM or CUPOM advances one time step in the integration. Because all three models have their own horizontal grids, linear interpolation is applied to those variables that are passed between the models. WAM takes the surface stress computed in MM5 as input at the most recent time step. It uses the diagnosed wave-induced stress and the diagnostic relation from (5) to compute the wave-induced roughness length. The roughness length then feeds back to MM5 for the computation of the surface fluxes at the next MM5 time step, while the wave-induced stress is taken by CUPOM along with the total stress from MM5. After CUPOM is integrated for one time step, the most recent SST is passed into MM5. During the MM5 in-
integation, both the SST and the wave-induced roughness length are kept constant until WAM or CUPOM provides a new value. Salinity fluxes associated with evaporation (including sea spray evaporation) and freshwater influx by precipitation at the air–sea interface, and enthalpy flux modulation by sea spray are taken into account in the surface flux calculation.

The momentum transfer from the atmosphere to the ocean is complicated by the presence of waves. Without waves, all momentum from the atmosphere is transferred directly into the ocean current. When waves are present, some of the momentum is transferred to the waves. The wave momentum is partly dissipated to the surface ocean current and partly radiated away from the area of wave generation. For a steady wind, after the wave field has reached equilibrium all of the momentum transferred to the waves immediately is transferred to the currents through wave breaking, as the waves are no longer growing. As this limit is approached, it can be assumed that all of the surface stress goes into driving the surface current (Lionello et al. 1993, 1998). However, for waves that are far from equilibrium, as are the waves that grow and radiate out from the center of a hurricane, an appreciable fraction of the momentum imparted to the waves remains in the growing wave field. This fraction will vary spatially, decreasing with distance from the center of the hurricane. In theory, with an accurate description of the wave dissipation at the surface as a function of frequency and wavenumber from a wave model, one could determine the amount of wave momentum lost to the current. However, for simplicity, in the present coupled model hurricane simulation we assume that only the part of the total stress left by subtracting the wave-induced stress (i.e., $\tau - \tau_w$) directly drives the ocean circulation in CUPOM.

The three models are initialized independently. MM5V2 is initialized by performing a successive-scan Cressman objective analysis on conventional surface and rawinsonde observations (including available ship reports), together with the first-guess fields from the gridded global analyses of wind, temperature, geopotential, and relative humidity at the mandatory levels from the National Centers for Environmental Prediction (NCEP). CUPOM is initialized using results from a multimonth spinup simulation forced by 6-hourly wind stresses from the U.S. Navy Fleet Numerical Meteorology and Oceanography Center (FNMOC). During the simulation, CUPOM assimilates Ocean Topography Experiment (TOPEX) and European Remote Sensing Satellite (ERS) altimetric data and the SST of the model is nudged to weekly composite MCSSTs derived from the National Oceanic and Atmospheric Administration’s (NOAA) Advanced Very High Resolution Radiometer (AVHRR) data to provide a realistic state of the ocean at the start of the coupled model simulation. The initialization of WAM can be carried out either by assigning a known spectrum at all grid points, or by computing a spectrum at each grid point from the initial wind fields according to fetch laws with a $\cos^2$ directional distribution (see Komen et al. 1994). In applications when reliable observations of waves are not available, WAM is initialized by simply assigning a zero state at each grid point. This is what is used in this study (with high winds it usually takes a few hours for the waves to become fully developed).

6. Numerical experiments

Numerical experiments in this study are designed to examine the sensitivities of the atmospheric circulation system to physical processes at the air–sea interface. Because it is well known that the intensity of a hurricane is controlled by the enthalpy flux across the air–sea interface for a given large-scale environment, we choose to simulate the scenario of a hurricane passing over an initially warm water surface to be used for the sensitivity experiments. Specifically, atmospheric analyses from NCEP for the period surrounding the intensification and landfall of Hurricane Opal (1995) beginning at 1200 UTC 2 October 1995 are used to provide boundary conditions for MM5. The initial conditions are constructed by incorporating a Rankine vortex into the analysis at 1200 UTC 2 October 1995, with the center of the vortex at the center of Hurricane Opal (based on the NCEP best track information). All model simulations are carried out for 72 h.

In this study, MM5 uses a nested grid system of two meshes, with grid resolutions of 45 km and 15 km. The finer mesh covers the entire Gulf of Mexico. Both meshes contain 25 sigma levels with the lowest level 15 m above the surface. The model physics includes the Betts–Miller parameterization scheme (Betts and Miller 1986) for subgrid water condensation, an explicit scheme (Reisner et al. 1998) for grid-resolvable water vapor condensation (taking into account cloud water, rainwater, and ice), and a Monin–Obukhov scheme for the surface flux [including the parameterized sea spray effect by Fairall et al. (1994) as an option]. The Blackadar scheme (Blackadar 1979; Grell et al. 1994) is used for the PBL mixing processes and for vertical diffusion.

a. Initialization of CUPOM

The grid of CUPOM used in this study has a horizontal resolution of $\frac{1}{5^\circ}$ in longitude (about 20 km) and $\frac{1}{25^\circ}$–$\frac{1}{5^\circ}$ in latitude (about 4–20 km with the higher resolution occurring near the coastline) and consists of 86 $\times$ 87 grid points (see Fig. 2). A total of 21 vertical sigma levels is used with corresponding physical depths of 0, 1, 2, 5, 10, 20, 40, 70, 100, 150, 250, 400, 600, 850, 1150, 1500, 2000, 2500, 3000, 3500, 4000 m in 4000-m water.

CUPOM is initialized with the output of a 9-month spinup run (ending 0000 UTC 1 October 1995) in which CUPOM is forced by 6-hourly wind stresses from the U.S. Navy FNMOC. The spinup run is initialized using
temperature and salinity profiles indicative of the Gulf water masses. During the spinup run, TOPEX- and ERS-2-derived altimetric sea surface height (SSH) anomaly data, as well as weekly composite MCSSTs derived from NOAA AVHRR satellite data, are assimilated into the model. Time-independent temperature and salinity profiles are prescribed at the basin inflow–outflow boundaries and the inflow–outflow are assumed to be geostrophic. The average inflow at the Yucatan Channel is constrained to be 28 Sv ($1 \text{ Sv} = 10^6 \text{ m}^3 \text{s}^{-1}$) and the inflow rate at the Yucatan Channel is varied monthly during the spinup run based on historical hydrographic data.

The MCSSTs used in this study are a 7-day composite with a horizontal resolution of about 14 km. Assimilation of MCSSTs is carried out by nudging the model SST to the observed value (Clifford et al. 1997; Horton et al. 1997). Altimetric SSH anomalies are assimilated track by track into the model by using the following approach. First, the altimetric SSH anomalies are converted into temperature profile anomalies using a statistical relationship between the dynamic height and temperature profile anomalies at each model grid point. This relationship is derived by carrying out empirical orthogonal function analysis (see, e.g., Carnes et al. 1990) using the SSH anomalies and temperature profile anomalies from a 10-yr model run. Second, the temperature profile anomalies are interpolated to the model grids using a statistical interpolation scheme (Horton et al. 1997). Third, the interpolated temperature profile anomalies are assimilated into the model in the upper 1000 m of the ocean using the nudging method (Choi et al. 1995; Bang et al. 1996).

A major feature of the large-scale Gulf circulation is the shedding of large clockwise-rotating warm-core eddies (often referred as Loop Current eddies, or LCEs). These are shed periodically by the Loop Current, which enters the Gulf from the Caribbean and exits through the Florida Straits to eventually become the Gulf Stream. It has been observed that the Loop Current in the Gulf of Mexico sheds roughly one to three warm-core eddies a year with a diameter of 100–400 km and a depth of about 1000 m (see, e.g., Elliott 1982; Kirwan et al. 1988). A number of numerical simulations of eddy shedding have appeared in the literature (see, e.g., Hurlburt and Thompson 1980; Arango and Reid 1991; Dietrich and Lin 1994).

Physical and dynamical characteristics of LCEs have been studied through hydrographic surveys (Merrell and Morrison 1981; Elliott 1982; Lewis and Kirwan 1987; Cooper et al. 1990), drifting buoys (Kirwan et al. 1988; Lewis and Kirwan 1987), and satellite infrared images (e.g., Vukovich and Crissman 1986). By the time the Loop Current passes through the Yucatan Channel, this flow achieves characteristics of a western boundary current. Because of its high momentum, it may penetrate several degrees northward into the Gulf of Mexico (about 26°N) before looping around anticyclonically to

![Grid system of the Gulf of Mexico](image)
Figure 3 shows the ocean model temperatures at 200-m depth at the end of the spinup from three CUPOM spinup runs (see Table 1 for a summary). SST data assimilation is performed in all three runs. In the first run (denoted as eddy 1), there is no assimilation of altimetry-derived temperature anomalies during the entire spinup period. In the other two runs (denoted as eddy 2 and eddy 3), assimilation of altimetry-derived temperature anomalies is carried out, respectively, for the layer between 100 and 1000 m, and for the entire upper 1000 m of the ocean. It can be seen that without assimilation of altimetry-derived temperature anomalies, the model produces an LCE at the end of the run (Fig. 3a), but the LCE is located to the south of the observed Eddy Aggie that was centered at about 26°N and 89°W (see Black and Shay 1998; Shay et al. 2000 hereafter SGMCB). When the assimilation takes place within the layer between 100 and 1000 m, or within the entire upper 1000 m, the resulting LCE is at the observed location of Eddy Aggie (Figs. 3b,c). However, when the assimilation is carried out within the entire 1000-m layer, the size of the LCE is

---

**Table 1. Summary of CUPOM spinup runs.**

<table>
<thead>
<tr>
<th>Name</th>
<th>FNMOC wind stress</th>
<th>Assimilation of MCSST</th>
<th>Assimilation of altimetry data</th>
</tr>
</thead>
<tbody>
<tr>
<td>Eddy 1</td>
<td>Yes</td>
<td>Yes</td>
<td>No</td>
</tr>
<tr>
<td>Eddy 2</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes, but only for the layer of 100–1000 m</td>
</tr>
<tr>
<td>Eddy 3</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes, for the entire upper 1000 m</td>
</tr>
</tbody>
</table>

---

An LCE named Aggie was present in the Gulf of Mexico during and prior to the development of Hurricane Opal. This eddy, which AVHRR data indicate was approximately 250 km in diameter, has been presumed to have had a considerable influence on the evolution of Hurricane Opal over the Gulf (see Shay et al. 1998; Black and Shay 1998). In the CUPOM spinup run, the purpose of assimilating both altimetry and MCSST data is to ensure that the ocean model state, including the LCE location and size, is close to that observed. The altimetry and SST data assimilation are found to be necessary to obtain a realistic representation of Eddy Aggie in the model state at the end of the spinup run. Results from several test spinup runs indicate that the shedding time and the size of an LCE are strongly dependent on how the altimetry and SST data assimilation is carried out. As shown later, when the data assimilation procedure is not performed properly, it is difficult to obtain a simulated Eddy Aggie that is in good agreement with its observed location and size.

Elliott (1982) calculated a mean radius of 183 km and translation speed of 2.1 km day⁻¹ for LCE’s using hydrographic data during 1965–72.
noticeably reduced and the temperature anomaly decreases by about 3°C. These results indicate that although the assimilation of the altimetry-derived temperature anomalies is needed to obtain the right location of the LCE in the ocean model initialization, whether or not the assimilation is carried out within the OML leads to different model initialization states. The implication of these results will be discussed later in the section on discussion and summary.

The upper-100-m mean temperature profiles averaged over the Gulf for the above three CUPOM spinup runs are presented in Fig. 4. It is seen that the mean temperature of the upper-100-m layer in eddy 3 is cooler than those in the other two (with the difference being a little less than 1°C at the surface and the 100-m depth, and more in between). The difference between eddy 1 and eddy 2 is much smaller than that between either of them and eddy 3. Note that the mean OML in eddy 3 barely exists, while in both eddy 1 and eddy 2 it is about 20 m, which is consistent with the shallow late summer seasonal thermocline in the Gulf (Monterey and Levitus 1997). Further examination of the results indicates that the failure in the OML development in eddy 3 is caused by the fact that the forcing introduced by the altimetry data assimilation interferes with the intrinsic mixing dynamics in the OML. In eddy 1, assimilation of altimetry data is not carried out; in eddy 2, it is applied only below the 100-m depth. Therefore, the OML is well developed in these two runs and the altimetry assimilation in eddy 2 generates an LCE with the correct location and size without interfering with the OML development. The impact on air–sea interaction of the differences in the upper-100-m mean temperature profiles for these three CUPOM spinup runs will be discussed later.

b. Initialization of WAM

The horizontal resolution of WAM is 0.4° (~40 km). The wave spectrum is discretized into 25 frequency bands and 24 directional bands. The frequency bands are logarithmically spaced from 0.042 Hz to 0.41 Hz at intervals of \( \Delta f/f = 0.1 \), while the directional bands are spaced evenly by 15°. WAM is initialized from a zero wave state. Under high wind conditions, the wave state described by WAM adjusts rapidly to the input wind forcing. Except for nowcasting applications, an elaborate initialization of WAM seems unnecessary for simulations in which the time integration of the model is carried for more than 1 day. Furthermore, because observational data are not routinely available, such an initialization is usually not possible.

c. Experiment design

Table 2 summarizes the numerical experiments carried out in this study. The purpose of these experiments is to examine the sensitivity of the simulated hurricane to the physical processes associated with air–sea interaction under high wind conditions.

1) Sensitivity to sea spray effect

Experiments with and without the sea spray parameterization (expt 1 and expt 2) are carried out to evaluate how sensitive the air–sea coupled modeling system is...
to the enthalpy flux contributed by sea spray. Three different values of $\beta$ (1, 0.5, and 0) are used in experiment 1 to reveal the sensitivity of the simulated hurricane to the uncertainty in the sea spray evaporation partitioning parameter. However, $\beta$ is set to be 0 in all the other experiments. The initial state of the ocean model resulting from the spinup run eddy 2 is used in experiments 1 and 2. The LCE in eddy 2 is located at the observed location of Eddy Aggie and has the approximately correct size and location. Although recent studies using idealized models (e.g., Fairall et al. 1994; Andreas and Emanuel 1999) have shown that sea spray described by existing parameterizations can have substantial effects on the evolution of tropical cyclones and the atmospheric boundary layer under high winds, it is still uncertain whether the inclusion of sea spray will have a significant impact on hurricane development in a realistic coupled air–sea model. Because the present coupled modeling system realistically takes into account atmosphere–ocean feedbacks (including atmospheric convection), it is a promising tool to assess the importance of sea spray evaporation.

2) Sensitivity to OML

Three experiments (expts 3, 4, and 5) are carried out to examine the sensitivity of the coupled model simulation to different prescribed horizontally uniform depths of the OML (i.e., 10, 30, and 50 m). The initial states of the ocean model used in these three experiments are specified by modifying the output of the spinup run eddy 3, such that the model state within the OML of a prescribed depth is equal to that in the top model layer. The LCE in eddy 3 is located at the observed location of Eddy Aggie, but it is smaller and cooler than that in eddy 2. The reason for modifying the initial state from eddy 3 is that the mean OML depth in it barely exists (Fig. 4). These experiments are motivated by the following considerations. The first is that an important aspect of air–sea interaction under high wind conditions is SST changes resulting from wind stirring and its feedback to the atmospheric circulation. The SST changes simulated by the coupled modeling system are strongly dependent on the distribution of the OML depth at the initial time of the coupled modeling. For the same wind condition, in areas where the OML is shallow, the SST change will take place more rapidly than in areas where the OML is deep.

The second consideration is that there are significant uncertainties in the initialization of the distribution of the OML depth. Presumably the spinup of CUPOM would produce a realistic distribution of the OML depth at the end of the spinup (i.e., the initial time of the coupled simulation) if the imposed forcing at the surface is realistic. However, due to the lack of wind observations over the Gulf, the forcing imposed during the spinup run comes from analyses that are strongly model dependent. Even if the surface forcing is realistic, it may not make up for a poor assimilation system that distorts the OML dynamics. Furthermore, the resulting OML depth of the spinup run is very much dependent upon the data assimilation procedure and the present algorithm has been calibrated mainly with deep sea observations of LCEs. Unfortunately, there have been no observations of the OML depth during and prior to Opal and, in general, expendable bathythermograph or conductivity–temperature–depth soundings in the Gulf of Mexico are sparse. Therefore, it is important to evaluate the sensitivity of the coupled model simulations to the initial OML depth.

3) Sensitivity to LCE: Location and Intensity

The energy to sustain a hurricane comes from the enthalpy input from the sea surface (Emanuel 1986; Rotunno and Emanuel 1987; Holland 1997), which is strongly dependent on the SST distribution along the passage of the hurricane. Numerical simulations (see, e.g., Bender et al. 1993; Ginis et al. 1997) have demonstrated that the cooling of the sea surface can have a profound influence on the intensity of hurricanes. It has also been observed that the decrease in SST in the wakes of hurricanes ranges from 1.5° to 9°C, depending on the translation speed of the hurricane, the intensity of the hurricane, and the underlying OML structure (Anthes 1982; Hodur 1996; Emanuel 1998, Sakaïda et al. 1998). The decrease of SST is caused by extraction of enthalpy at the sea surface by the hurricane, and entrainment of the colder water below in the OML forced by strong wind. Previous studies (e.g., Anthes 1982) indicate that the effect of entrainment mixing is much larger than the effect of enthalpy extraction at the sea surface. The role of warm LCEs on hurricane intensification over the Gulf of Mexico has recently been examined by Shay et al. (1998) using satellite-derived altimetry and SST data. Within warm LCEs such as Eddy Aggie, the OML is warm and deep compared with the surrounding water mass; the decrease of SST caused by the wind-driven mixing is slower; thus, the enthalpy flux to the air can be larger. The results from Shay et al. (1998) suggest that the enthalpy flux into Hurricane Opal from the sea was enhanced when the hurricane passed over Eddy Aggie. To examine the impact of oceanic warm eddies on the evolution of hurricanes, two experiments (expts 6 and 7) are conducted with different ocean model initial states, respectively, from the spinup runs (eddy 1 and eddy 3), in which the location and intensity/size of the LCE are different. To further examine the impact of the LCE itself on the simulated hurricane development, a third numerical experiment (expt 8) is carried out in which the water column in and near the warm eddy is modified by linearly interpolating the physical characteristics of the water neighboring the LCE. Comparison of this experiment with experiment 1 will be discussed in section 6d.
4) Sensitivity to Waves

The existence of sea surface waves alters the momentum and thermal fluxes at the air–sea interface. Since the development of hurricanes depends upon the energy fluxes across the air–sea interface, it is of interest to determine the importance of sea surface waves on hurricane development when air–sea interaction is simulated by existing model components. For this purpose, one experiment without WAM (expt 9) is carried out to compare with experiment 1 in order to evaluate how sensitive the coupled model simulation is to the wave-age-dependent roughness length. In this experiment, the ocean model is initialized with the spinup run eddy 2.

d. Results

Figure 5a shows the time series of minimum sea level pressure (hPa) sampled every 6 h for the numerical experiments with and without the sea spray parameterization. When the effect of sea spray is included, different \( \beta \) (1, 0.5 and 0) values are used. The time series of surface maximum wind speed (also sampled every 6
h) is shown in Fig. 5b for different $\beta$ values to compare with the simulation without sea spray. For $\beta = 0$, the track of the simulated hurricane with the inclusion of sea spray effect is shown in Fig. 5c, together with the sea level pressure at 51 h into the simulation. Note that for the first two days the simulated hurricane moves slowly in the southern half of the Gulf (i.e., south of 25°N) with translation speeds as small as 1.5 m s$^{-1}$, while it moves as rapidly as 11 m s$^{-1}$ as it crosses the northern half of the Gulf during the third day, reaching the coast at about 0009 UTC 5 October 1995. It can be seen from Figs. 5a and 5b that the inclusion of sea spray increases the intensity of the simulated hurricane significantly when $\beta = 0$.

Figures 6 and 7 depict the time series of surface stress and thermodynamic fluxes for the experiments with and without sea spray at the fixed location of the minimum sea level pressure (hereafter referred to as point C) shown in Fig. 5c. The maximum wind speed at the lowest model level at the same point is about 30 m s$^{-1}$ in the simulation without sea spray (not shown). When sea spray is included, for $\beta = 0$ the maximum surface stress increases by about 57% and 227% at the peak winds on either side of the hurricane eye (Fig. 6), while the maximum latent heat flux increases by about 68% and 160% (Fig. 7a), and the sensible heat flux by about 23% and 229% (Fig. 7b). When $\beta$ increases to 0.5, the maximum surface stress increases by about 50% and 70% at the peak winds, while the latent heat flux increases by about 58% and 78%, and the sensible heat flux decreases to small values that are slightly above zero. When $\beta = 1$, the maximum surface stress at point C increases by about 22% and 27% at the peak winds, while latent heat flux increases by about 31% and 42%. In this situation, however, the sensible heat flux decreases and becomes negative at the peak winds, for spray evaporation consumes more sensible heat in the air than both turbulence and sea spray can supply. It is interesting to note that when $\beta = 1$, the intensity of the simulated hurricane is not significantly affected by sea spray in comparison with the experiment without sea spray. This apparently is the result of the feedback of the spray evaporation that readjusts the sensible heat flux so that the lower atmospheric boundary layer be-
and prescribed with the distribution obtained in the simulation in Fig. 8. In this simulation, the SST is held constant to either CUPOM or WAM (expt 10) is also included. The uncoupled simulation in which MM5 is run alone without coupling is also included for comparison. For comparison, the result from the simulation in which the ocean mode was initialized with different, prescribed, horizontally uniform depths of the OML: 10 m (dot), 30 m (square), and 50 m (cross). The result from the uncoupled simulation (triangle) is included for comparison.

As the hurricane moves over the sea spray the turbulent energy flux is greater than the turbulent latent heat flux. This reduces the turbulent enthalpy flux from the ocean to the air, which is compensated by the additional spray-mediated flux so that the total enthalpy flux from the ocean to the air remains nearly the same as when sea spray is neglected. Finally, we note that the change in hurricane track among these different simulations is negligible and that the region of high velocity winds at this model resolution is quite broad. Therefore the differences shown in Figs. 6 and 7 are in fact due to differences in \( \beta \), and not to changes in hurricane position relative to the sampled grid point.

Sensitivity of the coupled simulation to different depths of the OML is illustrated by the time series of minimum sea level pressure (hPa) sampled every 6 h (Fig. 8). It can be seen that deeper OMLs lead to a significantly more intense hurricane, at least for the one shown in this study, which has a slow translation speed in its early stages. This is consistent with the results from earlier idealized numerical studies (see, e.g., Chang and Anthes 1978; Sutyrin and Khain 1984; Khain and Ginis 1991). We note that sensitivity of the hurricane to OML depth is very large, with a change of 35 mb in minimum sea level pressure for a 50% variation in OML depth about a nominal value of 20 m that is similar to that expected in the Gulf at the start of the hurricane season. For comparison, the result from the simulation in which MM5 is run alone without coupling to either CUPOM or WAM (expt 10) is also included in Fig. 8. In this simulation, the SST is held constant and prescribed with the distribution obtained in the spin-up run eddy 2, and the sea spray effect is turned off. A temporally constant SST is equivalent to an infinitely deep OML. In this case, the minimum sea level pressure reaches 910 mb.

In contrast to intensity, for the simulation shown in Fig. 8 it is found that the depth of the OML does not have significant impact on the hurricane track, with the hurricane’s position at landfall changing by no more than one to two grid points (not shown). However, the timing of landfall does change by as much as 6 h, with the most intense hurricane moving the most rapidly.

Figure 9 shows the simulated SST reduction at 72 h into the simulation for three different initial OML depths. Under hurricane conditions, SST changes are caused mainly by cold water entrainment across the thermocline (see, e.g., Anthes 1982, section 6.3). As the initial OML depth increases from 10 to 50 m, the area of SST cooling decreases substantially. However, the maximum change of SST decreases only slightly, remaining at approximately 4–5°C. SSTs at earlier times in the simulations (not shown) also indicate that the area of SST decrease in the wake of the hurricane is more spatially confined and that the change in SST is only slightly smaller for a greater initial OML depth. The magnitude of SST cooling is in agreement with results from the numerical study of Price (1981) for a slowly moving hurricane, but somewhat larger than that found by Elsberry et al. (1976) in a similar numerical study.

The small variation in maximum SST change shown in Fig. 9 can be explained by the fact that a deeper OML generates a stronger hurricane and greater ocean mixing, which tends to counter the effect of the greater thermal inertia of the deeper OML. Although the ocean mixing will not lead to a SST change for an infinite OML depth, it obviously can cause a significant SST response even for the 50-m OML depth. This illustrates how the ocean mixing works as an effective regulator to constrain hurricanes from reaching their maximum potential intensity.

For comparison, the SST reduction in experiment 1 (initialized with eddy 2) is included in Fig. 9d. It is seen that the deep OML depth within the warm eddy results in less SST cooling than the area outside of the warm eddy even though the hurricane passes right over the LCE. The magnitude of the SST cooling in Fig. 9d is close to that observed for Opal from AVHRR imagery (Black and Shay 1998), although the area of cooling is larger than observed.

Figure 10 depicts the sensitivity of the intensity of the simulated hurricane to the location and size of the LCE (i.e., eddy 1–expt 6, eddy 2–expt 1, and eddy 3–expt 7 in Table 2) in terms of the time series of minimum sea level pressure (hPa) sampled every 6 h. Because the OML depth within the LCE is deeper than outside the LCE, the spatial distributions of the OML depth in these simulations are different. Also, since the Gulf-averaged mean temperature profiles are nearly the same for eddy 1 and eddy 2 (Fig. 4), the difference between these two
Fig. 9. Simulated SST reduction in the wake of hurricane with different prescribed horizontally uniform depths of the OML at 72 h into the simulation: (a) 10, (b) 30, and (c) 50 m. The contour interval is 1°C. The SST reduction in expt 1 (initialized with eddy 2) is shown in (d) for comparison.

Runs shown in Fig. 10 reflect the effects of the spatially varying deviation of OML depth. It is interesting to note that when CUPOM is initialized with the output of eddy 1 (in which the LCE is located to the south of the observed location of Eddy Aggie), the simulated hurricane intensifies at a different rate than when CUPOM is initialized with the output of eddy 2 (in which the LCE is at the observed location). With the more southerly positioned LCE (eddy 1–expt 6), the rate of intensification is almost linear before 0000 UTC 5 October 1995, while with the LCE at the location of Eddy Aggie (eddy 2–expt 1) there is a rapid intensification that apparently accompanies the passage of the hurricane over the LCE (which started at about 1000 UTC 4 October 1995). Because the size and temperature excess of the LCEs in runs eddy 1 and eddy 2 are similar, the major difference in the time series of minimum sea level pressure between these two runs can be attributed to the difference in position of the LCE and the difference in translation speed of the hurricane. The more southerly position of the LCE in the run eddy 1 allows for the hurricane to intensify at an earlier time when the hurricane is moving slowly. It is also interesting to note that when CUPOM is initialized with the spinup run eddy 3, the simulated hurricane fails to intensify because in this spinup run the upper-ocean temperature is cooler and
More stably stratified than the other two runs (see Fig. 4). Despite this result, it is not yet clear that the LCE in the simulation initialized with eddy 2 is responsible for the rapid intensification of the simulated hurricane. This will be further investigated below with the results from experiment 8.

Figure 11 shows the initial bulk heat content of the upper ocean relative to the depth of the 26°C isotherm as defined in SGMCB by modifying the original definition in Leipper and Volgenau (1972). The magnitude of the heat content along the simulated hurricane track for eddy 2 is qualitatively consistent with the altimeter-derived values presented by SGMCB, although the peak value within the LCE is somewhat less (−25 vs. −30 Kcal cm⁻²). For eddy 3 the heat content is significantly lower. With hurricane forcing, the upper ocean loses heat to the atmosphere through the enthalpy flux across the air–sea interface, and to the deeper ocean beneath the thermocline through entrainment. The latter, based on SGMCB and references cited therein, accounts for 75%–90% of the upper ocean cooling. The cooling of the upper ocean leads to the reduction of enthalpy flux from the ocean to the atmosphere. For the same wind condition, the greater initial heat content the upper ocean has, the slower it cools, and the more enthalpy flux it can provide to the atmosphere. Therefore, the results shown in Fig. 11 provides a further explanation why the simulated hurricane initialized with the output of eddy 3 does not intensify: the upper ocean heat content is insufficient to fuel the storm.

Figure 12 shows the time series of minimum sea level pressure for the numerical experiment in which the wa-
extratropical cyclones. This result suggests that although reduced surface roughness does reduce the surface friction, it does not necessarily lead to the intensification of a hurricane because of either the spatially variable distribution of roughness produced by WAM, or because the enthalpy flux is reduced as well. This result also suggests that the interaction of a changing sea state with a hurricane is dynamically complicated; simply changing the Charnock parameter may not be sufficient to describe the dynamics involved in wave–air interaction.

7. Discussion and summary

In this study, a coupled air–sea modeling system is used to simulate air–sea interaction under high wind conditions. This coupled modeling system consists of three well-tested model components: the Penn State–NCAR Mesoscale Model, the University of Colorado version of the Princeton Ocean Model, and the WAMDI wave model. The scenario in which the study is carried out is the intensification of a simulated hurricane passing over the Gulf of Mexico. The focus of this study is to evaluate the impact of air–sea interaction on hurricane intensification and evolution.

The results from the experiments with and without sea spray effects show that in a coupled air–sea model the impact of sea spray evaporation on the simulated hurricane development depends on the evaporation efficiency of sea spray. When only a small amount of sea spray evaporates at the expense of cooling the remaining spray, the increase of enthalpy flux due to the evaporation produces stronger surface winds, which in turn increases the surface enthalpy flux. This positive feedback results in a significantly more intense hurricane. The model results, however, show that sea spray evaporation does not affect the hurricane intensity significantly when the evaporation efficiency is so high that the spray droplets evaporate at the expense of sensible heat from the ambient air. This is because the droplet evaporation cools the lower atmospheric boundary layer, and produces stable stratification so that the turbulent enthalpy flux within it decreases, which compensates the spray-mediated enthalpy flux. As indicated by Kepert et al. (1999), the effect of sea spray in this situation
will not directly change the total sea-air enthalpy flux to any great degree for a fixed wind speed; it will alter the partitioning of this flux into latent and sensible components.

For the atmospheric model’s horizontal resolution used in these experiments (15 km), the enthalpy flux at the air-sea interface calculated using the traditional Monin-Obukhov theory alone (i.e., without sea spray parameterization) appears to be insufficient to sustain an intense simulated hurricane when the negative feedback of ocean SST cooling is present. Sea spray may play an important role in the enthalpy transfer from the ocean to the atmosphere under extreme wind conditions. However, while the significant increase in intensification rate found when the sea spray parameterization is introduced is interesting, we caution that our conclusions here are based on a somewhat idealized study of a single case, with one parameterization scheme of the sea spray effect. Because systematic observations are not available to verify our results, uncertainties still remain, especially with respect to the source of sensible heat consumed by spray evaporation. Additional observations and modeling studies are necessary to fully elucidate the role of sea spray in tropical cyclone dynamics.

Proper initialization of the coupled modeling system is crucial to air-sea interaction studies. Although it is well known that successful simulations and forecasts of the structure and movement of hurricanes using an atmospheric model alone are strongly dependent on good initialization (see, e.g., Kuribara et al. 1995; Liu et al. 1997), it is one of the findings in this study that the development of hurricanes simulated by a coupled air-sea modeling system is quite sensitive to the initial mean OML depth. This sensitivity seems large compared to previous coupled model simulations (e.g., Hodur 1996), in which a larger OML depth of 50 m was used. However, the sensitivity found in the present simulations is similar to that found by Bender and Ginis (2000) in coupled model simulations of Hurricane Opal using realistic, climatological values of OML depth for the Gulf of Mexico. Results of the sensitivity experiments carried out in this study also demonstrate that the impact of the warm eddy shed from the Loop Current in the Gulf of Mexico on the development of the simulated hurricane is dependent on the location of the eddy relative to the hurricane path, and the structure of the eddy. The warm eddy changes the timing, rate, and duration of hurricane intensification.

It is shown in this study that the size, intensity, and position of the warm eddy associated with the Loop Current in the Gulf of Mexico are sensitive to the method of assimilating satellite altimetry data during the ocean model spinup run. Results of the ocean model spinup runs using different ways of assimilating altimetry data also suggest that in order not to distort the evolution of the OML structure during data assimilation, care must be taken during the procedure to preserve the physics in the OML. That is, the forcing introduced by data assimilation should not interfere with the intrinsic dynamics of the OML. In this study, when the satellite altimetry data were not assimilated into the top 100 m of the ocean, the OML depth and the characteristics of the warm eddy shed from the Loop Current appear to be reproduced more accurately.

In agreement with previous studies, the present results
indicate that the intensification of the model-simulated hurricane depends on the SST cooling due to the wind forcing associated with the hurricane. In contrast with the positive feedback from sea spray, the feedback from the SST change is negative in the sense that the reduction of SST results in a weaker simulated hurricane than when the SST is held constant during the simulation. The spatial extent of surface cooling is found to be strongly dependent on the initial OML depth, while the magnitude of the SST cooling is more weakly dependent on initial OML depth. Results from this study demonstrate that any numerical model intended to simulate air–sea interaction with physical soundness must include an accurate depiction of the oceanic mixed layer, including variations in the OML due to oceanic dynamic processes. However, again due to the scarcity of observations at high wind speeds, considerable uncertainty exists on the accuracy of presently available ocean mixing parameterizations in hurricane conditions. Our results also suggest that assimilation of both MCSST and satellite altimetry data is necessary to accurately describe the initial upper ocean structure in a region of large mesoscale variability.

The degree to which the hurricane is modified by SST cooling is dependent on the hurricane’s movement. Hodur (1996) discussed the relationship between the movement of a hurricane and the SST change caused by the hurricane in a coupled modeling study: the SST change has less effect on a faster moving hurricane than on a slower moving one (see, also, Chang and Anthes 1978; Sutyrin and Khain 1984; Khain and Ginis 1991). Because Hurricane Opal exhibited a wide range of translation speeds as it traversed the Gulf, the importance of SST feedback also varied significantly. The large sensitivity to mean OML depth found in our experiments is due in part to the exceptionally slow (1.5 m s\(^{-1}\)) translation speed of Opal in its early stages.

It has been demonstrated in the numerical experiments that the development of the simulated hurricane is dependent on the location and size of the warm eddy shed from the loop current and that this sensitivity depends on the translation speed of the simulated hurricane. Since the hurricane is moving rapidly as it passes over the LCE and approaches the coast in the eddy 2 simulation, it is affected little by the potential positive feedback of the LCE and by the potential negative feedback of the lower heat content water to the north of the LCE. This suggests that the existence of the warm LCE was not a primary contributing factor in the sudden intensification and weakening of Hurricane Opal. We note, however, that heat content estimated from satellite altimeter data using statistical regression has a considerable uncertainty, as seen in the 5 Kcal cm\(^2\) difference between our assimilation result and the analysis of SGMGB for the core of the LCE. The larger heat content found by SGMGB would have produced a somewhat larger hurricane response to the LCE. Given that the observed sudden intensification of Opal cannot be attributed solely to the effect of the LCE, it appears that atmospheric forcing or hurricane eyewall contraction dynamics must have played an important role. Bosart et al. (1998) provide evidence that an upper-level atmospheric trough helped intensify Opal. Our results also suggest that the sudden weakening of Hurricane Opal before it made landfall (shown by the NCEP best track information) may have been caused by atmospheric
forcing or by internal hurricane dynamics rather than air–sea interaction. Further research effort is required to understand what are primary factors to cause Opal’s sudden intensification and weakening.

Compared with the sensitivity to the initial OML depth and the location and intensity of the warm eddy associated with the Loop Current, the model is less sensitive to the wave-age-dependent roughness length. However, although the results from this study suggest that the Charnock parameter used in MM5 for obtaining the roughness length over the open sea may be too large under high wind conditions, it appears that the interaction of changing sea state with hurricane is dynamically complicated. Simply changing the Charnock parameter with different values may not be sufficient to describe the dynamics involved in wave–air interaction. The best way to evaluate the roughness length over the open sea is through use of a good ocean surface wave model.

Finally, it should be pointed out that this study addresses only one of several important factors contributing to hurricane intensity change: air–sea interaction. Hurricane development is a synergistic process that involves not only air–sea interaction but also the interaction of the hurricane with its atmospheric environment. In addition, the small-scale internal dynamics of spiral rainbands and eyewall contraction plays a vital role in hurricane intensity change. Even within the restricted scope of an air–sea interaction study, we caution that the results obtained in this are derived from a single event that is simulated only by using one model resolution configuration and a limited set of model physics. Uncertainties still remain with respect to the parameterization of sea spray effect, the initialization of the ocean model using altimetry data, and the degree to which one model component responds and feeds back to the other two components in the coupled modeling system. For example, whether the sensitivity of the simulated hurricane intensity change to the OML depth is realistic is a question that needs to be answered with further observations and modeling studies of high wind speed events.

Acknowledgments. We are grateful to several reviewers for their careful reading of this paper and their helpful comments. We also wish to acknowledge helpful discussions with C. Fairall and P. Lionello.

REFERENCES


