The effects of atmosphere-ocean-wave coupling during tropical cyclone

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**THE EFFECTS OF ATMOSPHERE-OCEAN-WAVE
COUPLING DURING TROPICAL CYCLONE**

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SCHOOL OF MECHANICAL AND AEROSPACE
ENGINEERING**

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**THE EFFECTS OF ATMOSPHERE-OCEAN-WAVE
COUPLING DURING TROPICAL CYCLONE**

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Abstract

It is well recognized that air-sea interaction affects the development and modulation of tropical cyclone (TC). However, open questions remain as to how and to what extent air-sea interaction affects the structure and intensity of a TC. This research project investigates and provides a better understanding of the coupling of atmosphere, ocean and ocean wave systems at the interface (*i.e.* sea surface) and its impacts on the TC structure. It is intuitive to think that, at the earth's surface, the fluid (both air and water) layer belongs to a single system, thus, justifying the construction of a single dynamically coupled model where atmosphere and ocean are interacting on different temporal and spatial scales. Within this coupled system, it can be thought that the ocean waves generated at the sea surface are due to the momentum transfer from atmosphere, which subsequently gets transferred to ocean. This transfer of momentum (from atmosphere to ocean) is not a one-way link, *i.e.* , there are feedback effects from ocean (both in terms of enthalpy and momentum) on the atmospheric boundary layer.

Earlier studies using coupled models have shown improvements in TC modelling and prediction. Studies using theoretical analyses and laboratory experiments have argued that the state of the art bulk parametrisation schemes that are currently implemented in these models are unable to account for the ocean wave related processes (*e.g.* , sea state dependent surface stress, sea spray mediated heat and momentum flux). In summary, the atmosphere-ocean-wave coupling affects the boundary layer and the TC structure. However, the highly complex and nonlinear nature of these processes makes them difficult to parametrise, which necessitates more work in the field of air-sea interaction.

Moreover, given the growing interest in renewable energy and wider acceptance of numerical modelling for the assessment of available renewable energy resources, improved modelling systems with better representation of air-sea interaction have the potential to contribute towards better forecasts of wind, wave and ocean currents. These models also have the potential to be used within the early warning systems, and to aid in human activities in areas such as fisheries, coastal constructions, marine transportation and oil exploration.

This research project utilises a dynamically coupled atmosphere-ocean-wave model to quantify the effects of air-sea interaction on the TC. A new fully coupled modelling system was designed and implemented. This coupled model was designed with the capability to conduct simulations in coupled and uncoupled configurations, so that the effects of different physical processes can be isolated. In addition to the coupling interface, the modelling system also includes several modules and modifications for including effects of processes such as sea spray and ocean currents. A series of simulations using different model configurations are carried out to study the effects of coupling wave model, sea spray mediated heat and momentum fluxes, sea surface temperature and ocean currents on a TC named Hurricane Arthur, that traversed through North Atlantic in 2014.

During the course of this study, special attention was paid to the often inadvertently overlooked factors that can modulate the intensity of a TC. For instance, recent studies using field observations and laboratory measurements have concluded that the momentum exchange coefficient saturates at wind speed $> 33\text{ms}^{-1}$. However, little is known about the behaviour of coefficient of enthalpy at such wind speeds. Furthermore, a number of studies have argued that the model skill can be potentially improved by including the effects of sea spray. However, most if not all, of the sea spray parametrisation depend only on the wind speed. Due to this wind speed dependency, these sea spray parametrisations have a tendency to overestimate the spray flux contributions. It can be posited that the sea spray generation should depend on both wind speed and sea state. A sea state dependent spray parametrisation was developed, where the spray generation was estimated from the wave energy dissipated through whitecapping was developed as part of this study.

Furthermore, the effects of atmosphere-ocean coupling on a TC are investigated, where the influence of sea surface temperature and ocean currents is considered. Within the atmosphere-ocean coupled model, the effects of surface waves were also considered, where the surface waves affected ocean model by means of radiation stress, while affecting the momentum flux in atmosphere model. Model results are evaluated using a comprehensive set of atmosphere and oceanic observations obtained from airborne radar, dropwindsonde, satellite altimetry and moored buoys. Model

results have shown that coupling a wave model enhances the momentum flux from atmosphere to ocean, where this enhanced momentum flux acts to reduce the wind speeds in high wind region, while increasing the enthalpy flux. Additionally, the coupling of ocean model affects the intensity of modelled TC, where the reduced sea surface temperature decreases the intensity of TC. The coupling of ocean currents also influences the intensity of TC, by reducing the momentum flux from atmosphere to ocean.

In conclusion, coupling of atmosphere, ocean and wave models affects the intensity and structure of TC. The coupling modifies the low-level boundary layer structure and the physical properties of near surface flow in TCs, which subsequently affects the physical mechanisms such as convective organisation. Lastly, it was shown that at the air-sea interface, there are a number of physical processes that can affect the TC. Unfortunately, these complex physical processes cannot be completely parametrised, thus, obviating the usage of fully coupled models.

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Dedicated to My Parents

Nikhil Garg

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List of Acronyms

AAM Absolute Angular Momentum.

ABL Atmospheric Boundary Layer.

AFWA Air Force Weather Agency.

AOML Atlantic Oceanographic and Meteorological Laboratory.

APDRC Asia-Pacific Data-Research Centre.

ASGAMAGE Air-Sea Gas Exchange and MAGE project.

ASMD Air-Sea Moisture Difference.

ASTD Air-Sea Temperature Difference.

AXBT Airborne EXpendable BathyThermograph.

BEP Building effect parametrization.

BSBT Backward in space, Backward in time.

CAM Community Atmosphere Model.

CAPE Convective Available Potential Energy.

CCN Cloud Condensation Nuclei.

CCSM Community Climate System Model.

CESM Community Earth System Model.

CFSR Climate Forecast System Reanalysis.

CISK Convective Instability of Second Kind.

COAMPS Coupled Ocean/Atmosphere Mesoscale Prediction System.

CRM Coastal Relief Model.

DEL Droplet Evaporation Layer.

DHI Danish Hydraulic Institute.

DTU Technical University of Denmark.

EKE Eddy Kinetic Energy.

ERA-Interim European Centre for Medium Weather Forecasting re-analysis project.

ESMF Earth System Modelling Framework.

FASTEX The Fronts and Atlantic Storm Track EXperiment.

GEBCO General Bathymetric Chart of the Oceans.

GFDL Geophysical Fluid Dynamics Laboratory.

GMTED Global Multi-resolution Terrain Elevation Data.

GOM Gulf of Mexico.

GPS Global Positioning System.

HEXOS Humidity Exchange over the Sea.

HRD Hurricane Research Division.

HYCOM Hybrid Co-ordinate Ocean model.

IFREMER Institut français de recherche pour l'exploitation de la mer.

IOWAGA Interdisciplinary Ocean Wave for Geophysical and Other Applications.

ITCZ Inter Tropical Convergence Zone.

LSM Land Surface Model.

LW Long wave.

MABL Marine Atmospheric Boundary Layer.

MAGE Marine Aerosol and Gas Exchange programme.

MCICA Monte Carlo Independent Column Approximation.

MCT Model Coupling Toolkit.

MERRA Modern-Era Retrospective Analysis for Research and Application.

MODIS Moderate-resolution Imaging Spectroradiometer.

MPI Maximum Potential Intensity.

MSLP Minimum Sea Level Pressure.

NCAR National Centre for Atmospheric Research.

NCEP-GFS National Centre for Environment Prediction Global Forecasting System.

NDBC National Data Buoy Centre.

NHC National Hurricane Centre.

NOAA National Oceanic and Atmospheric Administration.

NWP Numerical Weather Prediction.

OBL Oceanic Boundary Layer.

OML Ocean Mixed Layer.

OSTIA Operational Sea Surface Temperature and Sea Ice Analysis.

PBL Planetary Boundary Layer.

POM Princeton Ocean Model.

RH Relative Humidity.

RMSE Root Mean Square Error.

RMW Radius of Maximum Wind.

ROMS Regional Ocean Modelling System.

RRTM Rapid Radiative Transfer Model.

RSMC Regional Specialized Meteorological Centre.

SAR Synthetic-aperture Radar.

SAS Simplified Arakawa-Schubert.

SEF Surface Enthalpy Flux.

SFMR Stepped Frequency Microwave Radiometer.

SSGF Sea Spray Generation Function.

SSH Sea Surface Height.

SST Sea Surface Temperature.

SW Short wave.

SWAN Simulating WAVes Nearshore.

TC Tropical Cyclone.

TKE Turbulent Kinetic Energy.

UNESCO United Nations Educational, Scientific and Cultural Organization.

USAF United States Air Force.

USGS United States Geological Survey.

UTC Coordinated Universal Time.

UW University of Washington.

VHT Vertical Hot Tower.

VWS Vertical Wind Shear.

WISHE Wind Induced Surface Heat Exchange.

WNP Western North Pacific.

WRF Weather Research and Forecasting.

WSM WRF single moment.

List of Symbols

$(\)_0$ base state sea-level constant.

A horizontal eddy viscosity.

C_d Coefficient of drag [-].

C_g wave group velocity [m/s].

C_h Coefficient of heat [-].

C_q Coefficient of humidity [-].

D_h horizontal diffusion coefficient.

D_v vertical turbulent diffusion coefficient.

F forcing term for U, V, W, θ .

F_c horizontal diffusion term for momentum.

F_s horizontal diffusion term for salinity.

F_t horizontal diffusion term for temperature.

H_l Latent heat flux [Wm^{-2}].

H_s Sensible heat flux [Wm^{-2}].

L_0 Monin-Obukhov length [m].

L_v Latent heat of vaporization, 2.5×10^6 [Jkg^{-1}].

Q_l Spray latent heat flux [Wm^{-2}].

Q_s Spray sensible heat flux [Wm^{-2}].

S magnitude of discharge due to point sources [s^{-1}].

S_{ds} Wave dissipation source function [m^2s^{-1}].

- S_{in} Wind input source function [$m^2 s^{-1}$].
- S_{nl} Non-linear transfer source function [$m^2 s^{-1}$].
- T temperature [K].
- T_s temperature source term.
- U coupled horizontal component of velocity in x-direction [m/s].
- V coupled horizontal component of velocity in y-direction [m/s].
- W coupled vertical component of velocity [m/s].
- Ω Angular rotation rate of the earth, 7.2921×10^{-5} [s^{-1}].
- $\delta_{t,x,y,n}$ partial differential operator $\frac{\partial}{\partial t}, \frac{\partial}{\partial x}, \frac{\partial}{\partial y}, \frac{\partial}{\partial n}$.
- η Water surface elevation [m].
- $\hat{H}$ source term due to heat exchange with atmosphere.
- κ Von Kármán constant, 0.4.
- λ Wave length [m].
- μ hydrostatic pressure difference between surface and top of model.
- ω angular frequency [$rad s^{-1}$].
- $\overline{w'\theta'}$ Vertical turbulent flux of potential temperature [$ms^{-1}K$].
- $\overline{w'u'}$ Vertical turbulent flux of horizontal velocity [$m^2 s^{-2}$].
- ϕ geopotential.
- ψ_h Atmospheric stability function for heat [-].
- ψ_m Atmospheric stability function for momentum [-].
- ρ density [kg/m^3] e.g. ρ_a is air density, ρ_w is water density.
- τ Stress [Nm^{-2}].
- τ_w Wave induced stress [Nm^{-2}].
- θ potential temperature [K].
- θ_e equivalent potential temperature [K].
- θ_v virtual potential temperature [K].

- c_p wave phase speed [m/s].
- f coriolis parameter.
- f Frequency [s^{-1}].
- h Mean water depth [m].
- n ration of RK3 time step to the acoustic time step.
- p pressure [$kg\ m^{-3}$].
- p_h hydrostatic pressure.
- $p_{hs,ht}$ hydrostatic pressure at surface and top of the model.
- q_s Saturation specific humidity [m^2s^{-1}].
- q_{ref} Specific humidity at reference height [m^2s^{-1}].
- r_e Radius of earth, 6.370×10^6 [m].
- s salinity [PSU].
- s_s salinity source term.
- t time.
- u horizontal component of velocity in x-direction [m/s].
- u_* friction velocity [m/s].
- v horizontal component of velocity in y-direction [m/s].
- w vertical component of velocity[m/s].
- z_0 surface roughness length [m].
- z_Q Roughness length for humidity [m].
- z_T Roughness length for temperature [m].
- z_{ch} Charnock Constant.
- g acceleration due to gravity, $9.81\ [ms^{-2}]$.

Chapter 1

Introduction

1.1 Background and Motivation

Hurricane in the north Atlantic and eastern Pacific, *typhoon* in the western north Pacific and *tropical cyclone* in the Indian ocean and south Pacific ocean refer to the same phenomenon - an organised system of clouds with deep convection and closed low level wind circulation rotating around a well defined centre (commonly referred to as the *eye*). It is a warm core, non-frontal low pressure system usually originating from warm tropical oceans. Tropical cyclones (TCs) are arguably the strongest phenomena of the tropical atmosphere, and are widely recognised for their potential to destroy lives and property. The high socio-economic costs associated with TCs are well documented. The wind speeds in strong TCs can reach anywhere between 50-80 ms^{-1} . The wind intensity can last for several days, affecting the underlying ocean surface which then generates current and massive ocean waves.

An increasing demand for energy has led to the development of offshore infrastructure such as wind turbines and oil platforms in regions affected by TCs, like Gulf of Mexico (GOM) and Western North Pacific (WNP). The specific environmental conditions needed for the development and maintenance of TCs are present in tropical convergence zones. TCs originating from inter tropical convergence zone (ITCZ), which is to the east of WNP, drift westward, as the trajectory of TCs is mainly driven by the mean tropospheric flow, which is generally poleward. TCs that make landfall affect the East Asian countries of the People's Republic of China, Japan, South Korea,

Figure 1.1: The Pacific Tropical Cyclone Climate (a) Cyclone tracks for 1980-2005, (b) Cyclone tracks for 2011, RSMC Tokyo - Typhoon Center (2013)

Taiwan and the Southeast Asian countries of the Philippines and Vietnam. Figure 1.1 (a) shows the cumulated tracks of TC originating from the WNP region over a period of 25 years from 1980 to 2005. Meanwhile Figure 1.1 (b) tells us that within one year up to 20 cyclones (in 2011) can potentially hit the WNP.

To mitigate the risks posed by the TCs, it is essential to accurately forecast the intensity and the tracks of TCs. In the last two decades, with the tremendous increase in computing power and surge in modeling and observation activities, there has been a sustained improvement in the forecasting of TC track (see Figure 1.2 (a)). However, there has not been much improvement when it comes to forecasting hurricane intensity, as Figure 1.2 (b) shows.

Figure 1.2: Annually averaged official NHC 48h forecast errors for TC (a) track (nmi) and (b) intensity (kt) for TCs in the Atlantic basin from 1990 to 2005, Rogers et al. (2006)

Predicting the TC intensity remains a major challenge despite the sustained improvements in computing power, Numerical Weather Prediction (NWP) models and data assimilation techniques. Unfortunately, it is unclear whether the improvements in TC intensity forecast are limited by the lack of understanding of physical processes, deficiencies in NWP models or better observations.

It is widely accepted that the TC track is mainly influenced by the large scale atmospheric environment, while the TC intensity is influenced by the myriad of physical processes acting on different scales. Although it has been known that TCs are maintained by the enthalpy fluxes from the warm ocean surface, there is still an ongoing debate on whether the TC intensity is controlled by external factors such as sea surface temperature (SST) or by internal processes intrinsic to the storm. These physical processes include:

- air-sea interaction and physical processes in the boundary layer (BL);
- interaction between vortex and its large scale environment, which includes vertical wind shear (VWS), environment moisture, environment lapse rate for conditional instability etc.; and
- internal vortex dynamic processes.

Additionally, the three dimensional structure of the winds and convection modulate how these processes affect the intensity of a TC. According to the Wind-Induced Surface Heat Exchange theory (WISHE) (Emanuel, 1995), convection is self fuelled through coupling between surface fluxes and winds in a convective system, where, the TC growth is restricted by the magnitude of surface heat and moisture fluxes. Therefore, it can be said that air-sea interaction is one of the fundamental processes, as it leads to surface enthalpy - the source of energy - and momentum fluxes - the sink of the energy. Furthermore, these fluxes are locally determined by the dynamics of and thermal disequilibrium between the atmospheric boundary layer and upper ocean layers.

By using simplified model experiments, Emanuel (1986) and Rotunno and Emanuel (1987) have shown using simplified model experiments how the maximum potential intensity a TC can attain is linked to the large scale environmental factors such as ocean

Figure 1.3: An illustration of radar reflectivity in mature hurricane in the northern hemisphere (Houze, 2010)

surface temperature. These model experiments were developed within the paradigm that a TC can be regarded as a “Carnot Engine”, this paradigm is known as *maximum potential intensity* (MPI) theory. The MPI theory was subsequently extended by Bister and Emanuel (1998) to include small scale processes, which were not considered earlier. As the MPI theory is based on the large scale thermodynamical processes, it implies a long predictability of TC intensity. But because TCs rarely attain the intensity predicted by the MPI theory, other studies have focused instead on the role of processes intrinsic to the TC. For instance, the physical processes responsible for converting the sensible and latent heat flux extracted from the ocean to kinetic energy involve processes from wide range of scales such as those on the scale of convective clouds ($O(1\text{ km})$) and those on the scale of mean vortex ($O(100\text{ km})$). As most of the

heating is released in the form of convection, the arrangement of convective clouds must be important in determining the intensity of a TC.

Figure 1.3 which is taken from Houze (2010), shows the convective features in a TC, which include the inner core region comprising of the eye, eyewall and the heavy rain band (circling around the eye). There also exists an outer core region which is immediately outside the inner core and is made up of distant spiralling rain bands and the anvil clouds.

Figure 1.4: A schematic of air-sea interaction in TCs. It is a vertical cross-section from the storm centre to outer rainband, as shown in Figure 1.3. The black arrows in the eyewall indicate the secondary circulation (in-up and out). Red arrows indicate the upward heat fluxes and blue arrows indicate the downward momentum fluxes. The black arrows in the primary and secondary rainband indicate convective updrafts and downdrafts.

Figure 1.4 shows a schematic of air-sea interaction in TCs. It shows that the winds spiral inwards to the eye to reach their maximum (purple line) in the eyewall at a radius referred to as the radius of maximum wind (RMW). From the RMW, the wind speeds decrease rapidly inward with minimum at the center of the eye, while decreasing gradually outwards. In a TC, there are two prominent circulations - the primary circulation and the secondary circulation. The primary circulation results in the tangential components of wind (into/ out of) the plane in Figure 1.4 while the secondary circulation (“in-up-out”) creates the radial (inward or outward) components. As the air is pulled closer to the storm center by the strong radial pressure gradient,

by virtue of conservation of angular momentum, its rotational velocity increases. At a distance of 10-100 km, the direction of the motion of air suddenly changes from inwards to upwards in a ring of intense convection which is referred to as the eyewall. This air rises in the eyewall and turns outwards and moves away from the storm center when it encounters strong, stable air aloft (*i.e.* , in the stratosphere).

The radial inflow, which is generated in response to the frictional retardation and the gradient wind imbalance, is referred to as the frictional inflow. Apart from the frictional inflow, the secondary circulation also contains the convective updraft in the eyewall and the tropopause outflow due to the diabatic heating present in the eyewall.

Figure 1.5: White-caps (both active and residual) and the wave breakers as observed during hurricane Isabel on Sept. 12, 2003. The sea surface observation corresponds to a wind speed of 26 ms^{-1} .

Moreover, there exists a weak mid-level inflow between the boundary layer and the tropopause. This inflow is driven by the convective heating in the eyewall. There are convective updrafts and downdrafts in the rainbands, which entrain mid-level air into the BL. The frictional inflow, as it moves gradually towards the center of the TC, acquires surface fluxes (*i.e.* , both heat and moisture flux). The convective updrafts seen in the rainbands redistribute these surface fluxes due to the lifting of unstably stratified air to the higher levels. This redistribution of the surface fluxes necessitates

further investigation of the air-sea interaction and the resulting consequences for the TCs. A summary of the schematic of air-sea interaction is in order. The strong winds produce upwards enthalpy fluxes that allow the BL to gain energy from the ocean, which then gets transported inward to the eyewall by the frictional inflow. This heat energy when released by means of condensation (in eyewall updraft) helps intensify/maintain the TC. At the same time, the downward momentum flux from atmosphere to ocean enhances mixing in upper ocean layers and generates ocean surface waves. This enhanced mixing helps reduce the sea surface temperature (SST), which in turn reduces the TC intensity by consequential reduction in the surface fluxes. Finally, the ocean surface generated as a consequence of the downward momentum flux increases the surface roughness, which further increases the momentum flux transfer from atmosphere to ocean.

Here, we further elaborate on the dynamical coupling between the atmosphere and ocean. Due to the dynamic nature of water surface, evolves at different time scales. For instance, at a very low wind intensity, the interaction is predominantly governed by the molecular level interaction and density difference between the two fluids, while at high wind speeds, several relatively large scale processes come into the picture. In the presence of persistent intense winds, surface waves continue to grow and eventually break. These breaking waves entrap air in the form of bubbles, which can generate spray droplets that are then ejected from the sea surface into the air aloft. Moreover, at such high wind speeds, the crest of the surface waves can get sheared off resulting in so called “spume” droplets. An example of such phenomenon can be seen in Figure 1.5, which shows the white (due to entrapped air bubbles) foam being generated by the actively breaking wave as well as the residual white-cap from the previous breaker.

In summary, it can therefore be said that atmospheric and oceanic systems are coupled at air-sea interface through two physical processes, namely, the momentum and enthalpy flux. However, the interactions between these processes are complex and highly non-linear for conditions such as high wind speeds and huge direction variability prevalent during TCs. In numerical modeling (NWP), momentum flux is parametrised with an empirical expression for coefficient of drag C_d . Such expressions are usually

determined from field observations or laboratory measurements. For reasons such as complex sea state (*i.e.* , growing wind generated waves, direction variations) and the difficulty in carrying out field measurements in hurricane conditions, there is no unique relation for C_d as a function of wind speed U_{10} .

Usually, the two systems (*i.e.* , ocean and atmosphere) are modeled separately using a semi-empirical treatment of air-sea interaction. With the advent of spectral wave models in the last 20-30 years, it has become possible to predict ocean state with reasonable accuracy. These spectral waves models use balance between wave growth, wave energy dissipation, and non-linear wave-wave interaction to predict wave state. In order to forecast the wave field, these models use wind speeds as an input and can therefore provide the wave induced stress on the winds. Due to the inherent coupling between the wind and the surface waves, earlier studies by Doyle (1995) and Janssen et al. (2001) coupled mesoscale atmosphere model with an ocean wave model in order to address the deficiencies attached to the coefficient of drag based momentum exchange. Recent work by Chen et al. (2012) has affirmed the importance of coupled atmosphere-ocean-wave model for a holistic study of three systems and air-sea interaction however, current models still rely on assumptions which are conservative and underestimate the conditions during high wind speeds.

Studies such as those by Peake and Cox (2013) have developed a methodology to model extreme wind climates using publicly available storm track data, which are usually derived from historical satellite data, from sources like RSMC Tokyo - Typhoon Center (2013). There is burgeoning research that aim to improve the storm forecasting by using a 3D coupled numerical model to ensure less dependency on satellite data. However, past studies have shown that current models under-predict the storm intensity for a variety of reasons, including grid resolution to unresolved physics. Studies (Jenkins et al., 2012; Chen et al., 2013) have utilized a fully coupled Atmosphere-Ocean-Wave model to investigate the effects of ocean feedback mechanism on the storm intensity and structure. It has been observed that the use of coupled numerical model improves storm prediction but, as highlighted in Andreas et al. (2012), current models for coupling atmosphere and ocean rely on bulk parametrisation for exchange of flux at the interface. Andreas et al. (2012) refers to past studies to point out the

existing parametrisations are insufficient due to a lack of understanding of physical mechanism at the interface, as can be seen from the huge variability in the empirical relations from dataset to dataset.

The present study collaborated with DHI in the X-WiWa project¹ that looked at various coupling schemes and their effects in modeling of TCs. X-WiWa project utilized WRF model for the atmospheric component of the model and proposed to use DHI MIKE SW and DHI MIKE 3 for wave and ocean components respectively. Moreover, the X-WiWa project focused on the extreme wind and wave climate modeling in coastal areas in North Sea locations. In contrast, the present study was aimed at studying the effects of coupled model on TCs and extreme climate in tropics. It is worth mentioning that the climatic conditions in tropics and higher latitudes vary drastically, thereby, making it important to study the effects of coupled model and parametrisations in such conditions. Also, extreme climatic conditions in tropical storms pose another unique challenge, because assumptions such as in-plane wind and waves are not applicable.

1.2 Objectives

Recent literature has shown that the air-sea exchanges of heat, moisture and momentum influence how a TC gains its strength and intensity. The processes at the air-sea interface are influenced by a number of factors, including but not limited to the sea surface temperature, sea surface current, bottom orography, wind speeds and atmospheric temperature. All of these processes vary in both spatial and temporal scales ranging from few micro meters to several hundred kilometer. In order to completely resolve and account for all the physical processes, a “unified” high-resolution atmospheric-ocean-wave model is required. However, a review of the previous literature highlights that no model with such capabilities exists because of the fact that the governing theories and equations for all three systems (*i.e.* , atmosphere, ocean and wave) have been developed separately with certain commonalities. It is also important to mention that such a model will involve huge computing and development costs. In

¹ <http://www.xwiwa.dk/>

order to come up with a cheaper, working coupled atmosphere-ocean-wave model, researchers have developed coupling strategies that couple individual models at the interface, (*i.e.* , variables from each individual model are transferred at the coupling time step).

Studies like Lee and Chen (2012, 2014) utilized a coupled model to investigate the effects of air-sea interaction on TC boundary layer. Lee and Chen (2012, p. 3580) (see their “Figure 1”) have shown the prediction of storm track and intensity, suggesting that a coupled model shows some improvement over an uncoupled model. Lee and Chen (2012) have also concluded that the bias in results among other factors could be due to the poor surface layer parametrisations, which are incapable of accurately modeling the complex three dimensional structure present in the boundary layer.

The main objective of this research project is to investigate the role of a number of physical processes facilitating the interaction at the air-sea interface in modulating/affecting the structure of TCs. Numerous studies in the past have proposed methods and techniques, which include simple parametrisations and fully coupled three dimensional numerical models. This present study further examined the various physical processes, which connect the three systems - atmosphere, ocean and wave. There was emphasis on the smooth handling and transfer of variables and interfacial fluxes among the three models, whereby the role of other factors such as coupling time and grid resolution on the coupled model results was not investigated.

1.3 Scope

During the course of this project, the development of a coupled atmosphere-ocean-wave model was carried out using three different state of the art numerical models. The development of model coupling took a significant amount of time, with the focus on the development and application of adequate coupling strategy. This research project specifically looked at the role of feedback effect of waves on the atmospheric boundary layer (specifically TC boundary layer) and their effects on the intensity and strength of TCs. Due to time constraints, no experiments were carried out during the duration of this project. A wide variety of publicly available data was instead used for

the validation of results. The present project however did not cover the validation of modeling and parametrisation of some of the micro physical processes (for example sea spray) at the air-sea interface due to a lack of experimental data at such levels under a TC.

1.4 Research Significance

This research project is structured to provide new insights into and analysis of the role of surface layer parametrisation and the air-sea interaction processes by utilizing a coupled atmosphere-ocean-wave model. The coupled model was developed and improved during the course of this project. It is expected that this project will contribute to the vast literature on air-sea interaction, TC modeling and mesoscale modeling (of atmosphere and ocean). The development and validation of the coupled model will be done in collaboration with DHI (Singapore) and DTU (Denmark). This work draws on the existing knowledge base in the field of atmosphere, ocean and wave modeling in general, and interaction among atmosphere, ocean and ocean waves in particular. This research addresses the literature gap in field of hurricane modeling and prediction by providing useful methodology and tools for coupled model development and application and by showing how such methodology and tools can be applied.

1.5 Outline of Thesis

The structure of this thesis is as follows: Chapter 1 outlines the main objectives and scope of this research project while Chapter 2 provides a review of the literature on air-sea interaction and different coupling methodologies. Chapter 3 describes the different models used and the methodology adopted to develop a coupled model as well as the different sources of field observations used in the present study. Chapter 4 discusses the results obtained from a coupled atmosphere-wave model while Chapter 5 presents the sea spray model, which was developed during the course of this project, with an emphasis on the sea state dependent spray generation function and the effects of coupling sea spray on a TC. Chapter 6 further shows the effects of atmosphere-

ocean coupled model on TC, while also highlighting the effects of TC on ocean. Thereafter, Chapter 6 explains the scheme used to include the effects of ocean currents in atmosphere model. Lastly, conclusion, together with the implications of this research project for future work as well as the limitations of the current study, are summarized and briefly outlined in Chapter 7.

Chapter 2

Literature Review

2.1 General Theory

Csanady (2004) stated that processes like transfer of momentum, heat and mass are all irreversible processes. An irreversible process increases the entropy of the system. The Second Law of Thermodynamics tells up that the rate of entropy generation is always positive. When only one force is present, the entropy production rate is directly proportional to the product of Flux and Forces. De Groot and Mazur (1963) related “Flux” and “Forces” in following form

$$Flux = \frac{Forces}{Resistance} \quad (2.1)$$

where “Force” behaves like a potential gradient and “Resistance” like an inverse conductivity. Csanady (2004) posited that the transfer laws of air-sea interface (see Figure 2.1) can be viewed as a relationship between fluxes and forces. These laws are dominated by a blend of turbulent interaction between atmosphere, ocean flows and wind waves on the sea surface and are more complex than a linear relationship between “Flux” and “Forces”. Csanady (2004) stated that the most useful formulation for air-sea transfer laws is a relationship between a property difference (*e.g.* , density) across a layer of air above the interface and flux (*e.g.* , momentum) across the interface.

According to Babanin et al. (2012) and Chen et al. (2013), local wave effects are needed to provide coupling between large scale oceanic and atmospheric systems.

Figure 2.1: Illustration of air-sea interface, Janssen (2009)

Wind and pressure changes at the air-sea interface result in ocean wave generation, which varies over time scale, ranging from fraction of a second to a time scale of a few years, as can be seen in Figure 2.2.

Figure 2.2: Classification of ocean waves (Source: <http://www.physocean.icm.csic.es/Intro0c/lecture09-en.html>)

Figure 2.2 shows the classification¹ of ocean waves, *i.e.*, wind and tidal waves in temporal and spatial scales. Wave modelling takes into account the large scale and long term simulation of climate and general ocean circulation. The problem related to wave modelling has generally been classified into two main categories, namely:

1. Wind generated waves at sea surface;

- **Waves in wind conditions:** When wind speeds are higher than the phase speed of dominating waves;

¹ <http://www.es.flinders.edu.au/~mattom/Intro0c/notes/figures/fig9a2.html>

- **Waves during swell conditions:** When dominating waves are travelling faster than wind speeds; and

2. Internal and inertial waves driving ocean circulation.

Babanin et al. (2012) discussed the current state-of-the-art in surface waves and effects of wind-wave coupling. They described wind wave generation as a blend of turbulent theory of atmosphere and ocean boundary layer, wind-wave two way interaction, weakly non-linear resonant and quasi-resonant interaction mechanism complemented by strongly non-linear wave breaking. The various challenges associated with wind wave coupling as mentioned in Babanin et al. (2012) are:

- Timespace scale variation:** Where resonant wave-wave interaction takes place over a time scale of thousands of period while wave breaking has a time scale of fraction of a second.
- Sporadic and intermittent processes:** As some processes such as wind-wave exchanges above the surface are continuous processes while white-capping and wave induced turbulence are sporadic or intermittent processes.

Also, for shallow depths, the presence of internal waves and currents have to be taken into account. Similarly, during very strong winds, other processes have to be considered.

2.2 Wind driven air-sea interaction

The groundwork for turbulence and atmospheric research is based on the laboratory scale fluid mechanics research, which mean that any work in fluid mechanics starts from the Bernoulli's equation, which can be written as

$$p + \frac{1}{2}\rho u^2 + \rho g z = \text{constant} \quad (2.2)$$

The above equation states that the energy per unit volume is conserved along a streamline. For cases like a bluff body in flow path, Bernoulli's equation is not

constant because of a drag force exerted on the body by the fluid flow. This drag force can be written as

$$D = \frac{1}{2} \rho A C_d U^2 \quad (2.3)$$

Here, ρ is fluid density; A is the frontal area of the bluff body; C_d is drag coefficient and U is fluid velocity. Early researchers like Charnock (1955) and Wu (1982) have parametrised air-sea coupling in terms of this drag coefficient C_d , which is expressed as a function of wind speed. In the case of air-sea coupling, C_d is assumed to take into account both momentum and heat flux at the interface. According to Kraus and Businger (1994), as wind speed increases, sea surface changes from smooth to rough, thereby, resulting in air flow separation when the waves become too steep or start to break. Simultaneously, surface stress increases, resulting in a rougher surface in comparison to the other areas. To obtain the sea surface roughness from the measured local wind stresses, Charnock (1955) proposed a relation between sea surface roughness length and wind speed given as

$$z_0(\text{roughness}) = \frac{z_{ch} u_*^2}{g} \quad (2.4)$$

where z_{ch} is a non dimensional Charnock constant, g is acceleration due to gravity, u_* is friction velocity and z_0 is roughness length. Charnock (1955) derived this relation using dimensional analysis. Charnock constant z_{ch} is obtained from measured wind data by using curve fitting, whereas u_* is related to surface stresses by

$$\tau = \rho u_*^2 \quad (2.5)$$

Experimental works like the one by Wu (1982) suggested the value of z_{ch} as $0.011 < z_{ch} < 0.018$. To obtain a relation between wind stress coefficient C_z and wind velocity, we use logarithmic law given as

$$U_z = \frac{u_*}{\kappa} \log \left[\frac{z}{z_0} \right] \quad (2.6)$$

where U_z is wind speed at height z and κ is von kármán constant. Using eqs. (2.4) and (2.6), we can derive a relation

$$C_z = \left(\frac{\kappa}{\ln\left(\frac{1}{z_{ch} C_z F_z}\right)} \right)^2 \quad (2.7)$$

where C_z can be written as $(u_*/U_z)^2$ and F_z can be written as $(U_z/\sqrt{gz})^2$. For most wind speeds in neutral conditions, the above relation shows a very good correlation as can be seen in Wu (1982). However, it is important to note from the discussion above that z_0 is the only parameter used to represent interaction between wind and waves. Here, roughness length z_0 represents combined effects of wave formation, wave breaking and wave energy dissipation on the momentum transfer from wind to ocean. Kraus and Businger (1994) have argued that C_d would be a good indicator of wind wave interaction, where effects of fetch, sea state, static stability, secondary and non-steady flows have been taken into account. Wave models use energy balance equation for modelling the evolution of ocean waves. An energy balance equation for two dimensional wave spectrum $F(f, \theta; x, t)$, where F depends on wave frequency f , direction θ , location x and time t can be written as

$$\frac{\partial}{\partial t} F + \nabla(C_g F) = S_{in} + S_{nl} + S_{ds} \quad (2.8)$$

where C_g is group velocity for surface waves. S_{in} is a source term for atmosphere input while S_{nl} is a source term for four-wave interaction² and S_{ds} is a source term for dissipation. Janssen et al. (2001) described the role of ocean surface waves in the interaction between atmosphere and oceans. They stated that the ocean waves receive momentum from atmosphere through atmospheric input term S_{in} , whereas transfer momentum and energy to ocean by means of wave breaking and dissipation. From Charnock's relation, we have assumed that the drag coefficient is linearly varying with wind speed. However, this behaviour is in contrast with turbulent flow over smooth surface where drag coefficient decreases as wind speed increases. Janssen et al. (2001) proposed that wave induced stress should also depend on sea state. They obtained a

² Four-wave interaction is a process which conserves total energy and momentum

relation between wave induced stress τ_w and S_{in} as given by

$$\tau_w = \frac{\rho_w g}{C} \int S_{in} df d\theta \quad (2.9)$$

where wave induced stress depends on the wave spectrum. Also, the high frequency part of wave spectrum is shown to depend on sea state, thereby, concluding that the wave induced stress should depend on sea state. A relation between wave induced stress with Charnock's constant was given by Janssen et al. (2001):

$$z_0 = \frac{z_{ch} u_*^2}{g} \left(1 - \frac{\tau_w}{\tau}\right)^{-1/2} \quad (2.10)$$

Using experimental data from fetch limited and open seas, they also showed that Charnock's constant is sensitive to geographical location. Till now we have considered atmosphere to be of neutral³ stratification, whereas for stable and unstable stratification, buoyancy flux has dramatic effects on surface layer. This has been explained in depth by Kraus and Businger (1994) by means of Monin-Obukov similarity theory. For steady state and uniform horizontal conditions, turbulent kinetic energy (TKE) budget equation given by Kraus and Businger (1994) as well as Smedman et al. (1999) can be written as

$$\underbrace{\overline{u'w'} \frac{\partial u}{\partial z}}_P - \underbrace{\frac{g}{T_\theta} \overline{w'\theta'}}_B + \underbrace{\frac{\partial}{\partial z} \frac{\overline{w'e'^2}}{2}}_{T_t} + \underbrace{\frac{1}{\rho_0} \frac{\partial}{\partial z} \overline{p'w'}}_{T_p} + \underbrace{\overline{\varepsilon}}_\varepsilon = 0 \quad (2.11)$$

where P is mechanical production term; B gives production or loss term due to buoyancy; T_t is Turbulent transport term; T_p is pressure transport term; and ε is dissipation term for TKE. Under neutral condition, TKE equation reduces to

$$\overline{u'w'} \frac{\partial u}{\partial z} = -\varepsilon \quad (2.12)$$

³ Neutral stratification is rarer in actual conditions, where it occurs when a transition from stable to unstable stratification takes place. According to Stull (1991); Kraus and Businger (1994), neutral stratification is present on cloudy days, where no heat flux or buoyancy flux exist in surface layer.

where $\overline{u'w'} = \tau/\bar{\rho} = u_*^2$. The equation for TKE in neutral condition can be further modified using logarithmic law to give a relation between height and dissipation term by removing velocity gradient term.

From the above discussion of TKE budget equation we can see that mechanical and heat flux terms influence the surface layer as mentioned in Rutgersson et al. (2001). Monin-Obukov similarity theory (MO) associates flow turbulence with these two mechanisms, where stability of boundary layer is described by Monin-Obukov length scale (L) given as

$$L = -\frac{u_*^3 T_0}{\kappa g \overline{w'T'_v}} \quad (2.13)$$

where T'_v is virtual potential temperature; $\overline{w'T'_v}$ is buoyancy flux term; and T_0 is the mean temperature of the surface layer. As per MO, thermal and mechanical forcings are expressed by means of a non dimensional term z/L . A modified form of logarithmic law as given in Drennan et al. (2003) taking into account stability terms can be written as

$$\frac{U_z - U_0}{u_*} = f\left(\frac{z}{z_0}, \frac{z}{L}\right) \quad (2.14)$$

Weakly and moderately stable boundary layer can contain continuous turbulence near the surface and globally intermittent turbulence at higher level and can generally be explained by MO theory, according to Rutgersson et al. (2001). Several relation of the form as mentioned in eq. (2.14) have been proposed but their applicability is dependent on wind speed range and parameters such as location and sea state.

Apart from sea state and statistical stability, effects of turbulence have also been studied by several authors. Kraus and Businger (1994) stated that sea surface is not in equilibrium with wind input. Also, gustiness increases the energy transfer from wind to waves and effects of direction variability of waves with respect to wind should be taken into account, as they can modify sea state and consequently u_* . From the previous discussion, we can summarise that the drag coefficient C_d can be considered as a function of several parameters, which include:

1. The effects of wave age, c_p/u_* ;
2. The effects of stability, z/L ;

3. The effects of turbulence, σ_u/U ;
4. The effects of fetch, F/z_0 ; and
5. The effects of sea state given by 2D wave spectrum, where θ_w is the angle between wind and wave direction.

An expression for C_D given by Kraus and Businger (1994) can be written as

$$C_D = C_{Dn} K \left(\frac{c_p}{u_*}, \frac{z}{L}, \frac{\sigma_u}{U}, \frac{F}{z_0}, \theta_w, \dots \right) \quad (2.15)$$

where C_{Dn} is neutral-equilibrium drag coefficient and K is a complex coefficient accounting for several parameters. Fetch F is defined as the length of water over which given wind has blown *i.e.* length of water surface along a particular direction over which particular wind speed is present. Here, it is essential to mention that the terms in bracket on right are not independent of each other. For example, direction between wind and wave will affect both sea state and turbulence, other parameters similarly affect each other.

2.3 Air-sea interaction under swells

Air-sea interaction when the dominating waves are faster than the wind speed are less well studied than when the wind speed is higher than that of dominating waves. According to Rutgersson et al. (2001), swell can be defined by means of wave age. Wave age for full developed waves is given by

$$\frac{c_p}{U_{10}} = 1.2 \quad (2.16)$$

where U_{10} is wind speed measured at 10 meters height above the surface. Waves with wave age above 1.2 can be referred to as swell. According to Smedman et al. (1999), during swell conditions, surface waves transport momentum upwards into the atmosphere by means of pressure fluctuations inside the inner surface present just above the sea surface. They also showed that the mechanical production term in TKE budget equation is nearly zero and that the flow is dominated by buoyancy and pressure

fluctuation terms. Smedman et al. (1999) stated that for young wave conditions, momentum deficit dominates the flux plot while for swell conditions, outward interaction is in same order of magnitude as momentum deficit. These variations, which include flux behaviour, were linked to a phenomenon called “Inactive Turbulence”; where inactive turbulence is defined when there is little change in turbulence with the drop in momentum. Some key characteristics of "Inactive Turbulence" listed by Smedman et al. (1999) are:

1. It does not aid in momentum transfer.
2. It does not interact with active turbulence in inner layer.
3. It is of relatively large scale and is present in upper part of boundary layer.

Smedman et al. (1999) also showed that the higher turbulence near the surface layer is due to the entrainment of inactive turbulence from upper surface layer. Also, it was shown that the transport of momentum has little to no effect on heat flux in the surface layer, thereby concluding that momentum is transported by means of pressure-velocity correlations, which is incapable of transporting scalars. An example of such condition where dominating waves are faster than the wind speed would be just after a gale, *i.e.* , when the wind speed has dropped down. An important takeaway from this study was the evidence of transfer of momentum from ocean to atmosphere by means of pressure fluctuations, simultaneous to the transfer of momentum from atmosphere to ocean.

2.4 Surface Heat Budget

In addition to the momentum flux exchange at the air-sea (or air-land) interface as described in the preceding section, atmosphere and ocean also exchange heat, mass and gases. According to Kantha and Clayson (2000), the total heat fluxes at the interface comprise of radiative and turbulent fluxes. The radiative energy from sun is emitted on the shorter wavelengths (*i.e.* , ultraviolet, visible and infra-red) in comparison to the radiation emitted from relatively much cooler earth surface. Therefore, the radiative fluxes are traditionally classified into long wave (LW) and short wave (SW) radiative

fluxes. The long wave radiation is emitted from the atmosphere (*e.g.* , clouds) and earth surface (*e.g.* , land). The turbulent fluxes consist of latent heat flux (*i.e.* , due to the evaporation) and the sensible heat flux (*i.e.* , convection). These heat fluxes (*i.e.* , turbulent) are directly influenced by the turbulence in the near surface (or boundary) layer, in contrast to the radiative fluxes which are not affected by the boundary layer turbulence.

$$LW \downarrow - LW \uparrow + SW \downarrow - SW \uparrow - H_s - H_l = 0 \quad (2.17)$$

The heat flux balance at the surface (*i.e.* , air-sea or air-land interface) is described in eq. (2.17), where the upwelling radiative fluxes are shown with $\uparrow$, while the downwelling radiative fluxes with $\downarrow$. A detailed illustration of the earth's annual mean energy budget can be seen in Figure 2.3. The negative sign of latent H_l and sensible H_s heat flux indicates the loss of heat flux from the ocean (or land) surface.

Figure 2.3: Schematic of the Earth's mean energy budget, taken from Kantha and Clayson (2000). The units are in W/m^2

It is needless to say that the illustration in Figure 2.3 is highly simplified, where the various parameters such as the type of land surface, ice, type of vegetation, urbanisation and their affects have not been described.

2.5 Air-sea interaction in a tropical cyclone (TC)

2.5.1 Energetics of Tropical Cyclone

It can be unequivocally said that there exist two major theories for describing the conceptual model for intensification of a TC, where the first (older) one is the Cooperative Intensification theory of hurricanes (Ooyama, 1964, 1982) and the second (newer) one is the Wind Induced Surface Heat Exchange (Emanuel, 1986). It should also be pointed out that the cooperative intensification theory succeeds its more widely known predecessor, *i.e.*, the Conditional Instability of Second Kind (CISK), which is attributed to both Charney and Eliassen (1964) and Ooyama (1964). These two theories differ mainly in terms of the stress laid upon the importance of convection and the heat fluxes for the intensification of TC, Smith (1997). Over the last decade, a newer model of TC intensification known as rotating convective towers (or vortical hot towers, VHT) (Montgomery et al., 2006) has emerged. This newer paradigm has emerged as an extension of the cooperative intensification theory (Ooyama, 1982), where the conceptual model has been extended from a two-dimensional axis-symmetric viewpoint to a three-dimensional one. A brief description of the two theories (*i.e.*, cooperative intensification theory and WISHE) is provided below. A detailed description of the various paradigms for TC intensification can be found in Montgomery and Smith (2014).

Cooperative intensification theory of hurricanes

In the cooperative intensification theory, it was envisaged that the buoyancy force in the field of cumulus clouds present near the centre of a rotating flow would lead to the intensification of the vortex. Under this paradigm, it was hypothesised that the rate of latent heat release by cumulus convection and the moisture convergence in the boundary layer of the vortex would work in tandem (or cooperation). Figure 2.4 illustrates the conceptual model of the cooperative intensification theory, where the rate of latent heat release leads to the intensification of the vortex, and the intensified vortex results in increased moisture convergence, and hence, an increased rate of latent heat release.

Figure 2.4: An illustration of the schematic of CISK theory for the intensification of TC, taken from Laing and Evans (2011)

Also, in CISK, the surface friction plays two competing roles, where it results in the deceleration of surface winds (momentum loss) while enhancing the moisture convergence within the boundary layer of the vortex. In order for TC to intensify, the boundary layer moisture convergence and hence the rate of latent heat release must be greater than the energy dissipation due to surface friction.

Wind Induced Surface Heat Exchange (WISHE)

Emanuel (2003) presents a detailed introduction to the theory of air-sea interaction developed by Emanuel (1986), where he explained the process by means of a Carnot engine cycle (see Figure 2.5). As described in Emanuel (2003), air spirals in isothermally from (A to B), during which it suffers a drop in pressure and increase in entropy while losing its kinetic energy to the atmospheric boundary layer and angular momentum due to the frictional torque with the sea surface. From B (eye wall), air turns

upwards following adiabatically a surface of constant entropy and angular as it reaches the point of low pressure at O. From O, air descends (isothermally) to O' losing entropy gained during the inflow leg (from A to B) via electromagnetic radiation, and eventually reaches starting point A, thereby closing the loop. Here, it is important to note that this is an idealised explanation of the energetics of a TC. Emanuel (2003) notes that this explanation deviates from the actual scenario where in TCs the air flow moving outward (*i.e.*, starting from the knee part of B - O curve) experiences strong exchanges with the environment (*i.e.*, an open cycle), contrary to the numerical idealisation described above.

Figure 2.5: The hurricane Carnot cycle, here vertical axis represent vertical elevation above sea surface, where $Z = h$ is the top of surface/ lower boundary layer and horizontal axis represent radial distance from the centre of storm. Air spirals in from A to B, where at B (eye wall) flow turns upwards and ascends adiabatically to a low pressure point O, from O air descends isothermally to O' while losing heat by electromagnetic radiation and then from O' air returns back to starting point A completing the full cycle, Emanuel (1986)

According to Emanuel (2003), TCs are driven by the transfer of enthalpy fluxes from sea while controlled or limited by the surface drag (momentum flux) near the surface. Geernaert et al. (1986), using the valid field data (*i.e.* wind $< 30\text{m/s}$) collected during TCs, showed that the drag coefficient increases with the wind speed while a little change is observed in enthalpy fluxes. Also, Emanuel (2003) by means of

numerical experiments concluded that if the conventional surface layer parametrisation (for enthalpy and heat fluxes) were to be used then it would not be possible to sustain a TC higher than marginal intensity. From the above discussion, we can construe that there must be an array of mechanisms (not only enthalpy flux) at play near the ocean surface (*i.e.* , where atmosphere and ocean meets) influencing the energy gain by the storm which if modelled using conventional techniques would not provide sufficient energy (heat and momentum) transfer to the storm to reach its potential intensity. Emanuel (2003) suggested that sea spray could be one of those processes near the ocean surface which might explain to some extent the additional enthalpy flux from the ocean to the TC. Andreas and Emanuel (2001) also concluded that for accurate prediction of hurricane (or TC) intensity, it is necessary to couple atmosphere model with a prognostic surface wave model, in contrary to author's previous works where it was thought that a simple coupled model with specified fluxes should be able to model the TC intensity and strength, which leads us to our first conclusion that there is clearly an issue regarding the surface flux parametrisation at wind speeds higher than 20m/s .

2.5.2 Ocean response to Tropical cyclone

It is widely accepted that the passage of storm over the ocean surface decreases the sea surface temperature (SST) by means of storm induced upper ocean mixing. This induced upper ocean mixing results in deepening of the mixed layer propelling the cooler thermocline water from higher depths to upper mixed layer. It is also well established that this cooling of SST which is estimated to be as high as 6°C affects intensity of the TC (*i.e.* , weakening of TC). Price (1981) showed by means of numerical results that this cooling is due to the entrainment caused by the turbulence generated at the base of mixed layer with strong shear of near inertial currents. Researchers recognising this feedback mechanism between ocean and atmosphere, used coupled models to simulate the response of ocean to the storm induced momentum transfer and then back on the TC.

Fan et al. (2009, 2010), using a coupled axis symmetric winds (idealised storm) together with an ocean-wave model, simulated the response of ocean and waves to

the storm for different scenarios like a steady storm, moving storm, effects of ocean currents on waves. Lee and Chen (2012, 2014) and Chen et al. (2013) used a coupled model for investigating the effects of ocean on the hurricane boundary layer using test cases of actual TCs. It was found from the studies carried out using idealised and actual TCs that much of the cooling takes place on the right side (for Northern Hemisphere) of the storm. When coupling the models (*i.e.* , atmosphere, ocean and wave), it is normally assumed that the momentum exchange from atmosphere to waves gets transferred to the ocean (*i.e.* , no net momentum gain by waves). However, Fan et al. (2009) found that this assumption is not valid for all the cases (like growing waves or mixed type surface waves *i.e.* , fully developed and young waves). Under TC conditions, the wave field is complex and fast varying and can affect the air-sea flux budget. They also showed that when the storm is moving, due to the misalignment between the directions of ocean current vectors and wind stress, there is a reduction of 0.2 m/s in ocean current velocity on the right side of the TC, which results in a reduction in SST cooling of up to 0.5° C .

2.5.3 Influence of surface waves on TC intensity

As mentioned in previous sections, downward momentum flux from atmosphere induces surface waves. Using data collected from airborne scanning radar altimetry during Hurricane Bonnie, Wright et al. (2001) showed that longer waves with higher significant wave heights were observed in front right quadrant while underdeveloped shorter waves with lower wave heights were observed in the rear left quadrant. Moon et al. (2004a) and Chen et al. (2012, 2013) observed similar asymmetry in wave field under a TC. They associated this asymmetry in wave field to the presence of stronger winds in the front right quadrant of the storm, translation speed of the storm, angle between wind and wave propagation direction.

From Figure 2.6 taken from Chen et al. (2013, p. 3205), we can see that the wind and wave direction vectors are aligned in the front right quadrant while significantly misaligned in the rear left quadrant.

As described in section 2.2, surface momentum flux has been thought to be dependent on factors like sea state and wind speeds among others. From Figure 2.7

Figure 2.6: The atmosphere-wave-ocean model simulated (a) Colour contours of significant wave heights [m] and mean wave propagation direction (white arrows), (b) Colour contours of mean wave length and direction of surface wind (black vectors). Storm center is represented by black sign. Chen et al. (2013)

Figure 2.7: The estimates of drag coefficient versus wind speeds derived from aircraft measurements during CBLAST project conducted in hurricanes Edouard, Isadore, Lili (2002), Fabian and Isabel (2003), Frances and Jeanne (2004), together with published data from both laboratory and field level studies (from Black et al. (2007, p. 363))

taken from Black et al. (2007), we can clearly see the behaviour of coefficient of drag with wind speeds (at 10m). We can make a few observations from the results shown in Figure 2.7. Firstly, we can notice that there is a huge amount of scatter between different datasets. Secondly, we can see that the measurements are still under 30 m/s.

Lastly, we can see that some empirical relations such as Donelan et al. (2004) (black dash dot line) and Powell et al. (2003) (red dash dot line) fall within the scatter but the saturation of C_d as predicted with empirical relations is not noticeable in the field data.

The surface layer parametrisation used in the mesoscale atmospheric models such as Weather Research and Forecasting (WRF) model uses the relation obtained from Donelan et al. (2004), which essentially treats the drag coefficient as a function of the wind speed. Soloviev et al. (2014) recently published results where they used the data collected in the laboratory together with field observations and the results obtained from Large Eddy simulations to derive a relation of C_d versus wind speeds for wind speeds above 30 m/s . Soloviev et al. (2014, p. 3) showed two different relations for C_d based on the different estimation methods (*i.e.*, surface stress and surface roughness), where one relation shows that after wind speed exceeds 30 m/s , the change in C_d versus wind speeds is reduced, while in the second relation C_d decreases with the increase in wind speed, which can also be noticed in the results of Troitskaya et al. (2012). The decrease in C_d is thought to be due to the flattening of waves caused by the higher wind speeds blowing away the crests of steep surface waves, thereby reducing the surface roughness. We can also see some contrasting behaviour of C_d versus wind speeds in results of Kudryavtsev et al. (2014), where they presented a wide range of estimates of C_d with the wind speeds. From the above discussion, it can be concluded that the effects of TC on surface waves and the feedback effects of waves on TC are not conclusive and vary among different studies.

2.5.4 Influence of sea spray on TC intensity

Fairall et al. (2009) stated that the current air-sea flux parametrisations are reasonably valid for wind speeds (at 10m above sea surface) up to 20m/s. Due to the lack of field observations, to obtain values for air-sea fluxes at wind speeds greater than 20m/s, usually an extrapolation of the data (or the curve fitted to the data) is performed, where this extrapolation of fitted curve and the fluxes obtained from it have been subjected to criticism over the years. Emanuel (1995) conducted a detailed study using an axis-symmetric hurricane model to investigate the sensitivity of TCs to the available flux

parametrisations for the lower boundary (*i.e.*, near the sea surface) and concluded that the fluxes obtained by extrapolating values to hurricane wind speeds are inconsistent with the potential strength of TCs. Emanuel (1995), through his study, also noted that for the intense TCs (or hurricanes) to attain their respective intensity, the ratio of C_k/C_d is an important parameter, where the value of C_k/C_d in the high wind speed regions must range between 1.2 - 1.5, here C_k and C_d are coefficient of transfer of enthalpy and coefficient of drag (or coefficient of transfer of momentum) respectively.

Observation studies such as Powell et al. (2003) have reported that in hurricane level winds, coefficient of drag does not monotonically increase with wind speed as would be implied by the relations such as Charnock (1955). They found that drag coefficient levels off at wind speed above 34ms^{-1} . Powell et al. (2003) hypothesised that such a behaviour of drag coefficient could be due to the transfer of momentum by the sea spray. Also, Bao et al. (2011) in their study reported that the value of ratio C_k/C_d is between 0.6-0.7 using the field data collected in moderate wind speeds ($U_{10} < 30\text{ms}^{-1}$), whereas traditional parametrisations give a value which is lower than 0.7 for wind speeds greater than 40ms^{-1} . They attributed this discrepancy to the presence of sea spray in the lower boundary layer (*i.e.*, above the water surface).

Spray generation mechanism

Andreas and DeCosmo (1999) state that sea spray droplets form mainly through two processes, where the formation of sea spray in first process is described as the trapping of air during the wave breaking, where this trapped air gets distributed in the form of bubbles. With the rising surface, these bubbles result in white caps which form jet and film droplets. These film droplets thus formed from the bubble caps rupture and shoot upwards, similar to the jet droplets which are directly formed from the bubbles (when the bubble cavity collapses). Both of these droplets (*i.e.*, jet droplets and film droplets) are called bubble droplets. Additionally, during high wind speeds the sea spray is also produced by a second process. When the wind becomes too high (*i.e.* greater than $18\text{--}20\text{ m/s}$), it creates spray bubbles by tearing it from the wave crests. These type of spray droplets are known as spume droplets.

Figure 2.8 provides a very insightful illustration about the generation of sea spray

Figure 2.8: An illustration of the heat flux (interfacial and spray mediated) transfer within the droplet evaporation layer (DEL), where DEL extends to one significant wave height from the sea surface. Q_s, Q_l represent droplet radius specific spray mediated sensible and latent heat flux, ρ_s is the density of sea water and L_v is latent heat of vaporisation of sea water. Interfacial sensible and latent heat flux are represented by H_s, H_l .

droplets and its interaction with the marine atmospheric boundary layer. We can see that, when the temperature of sea is higher than that of the Droplet Evaporation Layer (DEL), sea spray loses heat in the form of sensible heat to the DEL. If the DEL is not saturated (*i.e.* relative humidity $< 98\%$), then the spray droplet evaporates by extracting heat from the DEL in the form of latent heat and results in the change in moisture content of the DEL (or the Marine Atmospheric Boundary Layer (MABL)). According to Andreas and Emanuel (2001), when these cooled droplets enter ocean (*i.e.* , re-entrant spray), they result in the reduction in ocean sensible heat (as the temperature of ocean is higher), thereby completing the loop (*i.e.* , spray mediated sensible heat transfer from ocean to atmosphere).

As mentioned above, there are two major processes of droplet production - from bubbles and from tearing of wave crests. In order to calculate the droplet production rate, it is also important to distinguish the various droplets based on their sizes. A rough estimate of droplet size is presented in Kepert et al. (1999) and Andreas and

Figure 2.9: A plot of sea spray generation function in terms of volume of spray droplets produced per square meter of ocean surface per second per micrometer of increase in droplet radius (volume flux) as a function of wind speed at 10m for wind speed of 15 m/s . The size distribution of droplet size can be seen on the x-axis. (Andreas and DeCosmo, 1999)

DeCosmo (1999), where the radius of film and jet droplets ranges from a few μm to a few tens of μm , while the radius of spume droplets is greater than $20\ \mu\text{m}$. Figure 2.9 illustrates a profile of volume flux of spray droplet production as a function of droplet radius based on the micro-physical model developed by Andreas (1992). Apart from the size of droplet, the other important factor is the residence time of spray droplet in the MABL, which according to Andreas and DeCosmo (1999), leads them to believe that only spray droplet under $500\ \mu\text{m}$ can result in the additional heat and momentum flux transfer from the ocean to the atmosphere. Andreas and DeCosmo (1999) provided the profiles of droplet temperature and droplet radius as a function of residence time in the MABL in their Fig 13.3, where they showed that the time scale of drop in temperature and size (radius of droplet) of a spray droplet can differ up to three orders of magnitude, which means that the transfer of heat from the droplet

to the MABL (*i.e.* , sensible heat)⁴ takes place faster than the transfer of heat from the MABL to the sea spray droplet (*i.e.* latent heat), thereby resulting in evaporation of sea spray droplet to equilibrium size (*i.e.* the size of spray droplet which attains equilibrium with its environment).

Impact of sea spray on air-sea fluxes

Bianco et al. (2011, p. 362) summarizes the complexity involved with the evaluation of the effects of sea spray on the air-sea fluxes by stating that “The combined feedback effect of sea spray to overall air-sea momentum and enthalpy flux have been debated since such a combination adds to the complexity of the air-sea interaction”. Bianco et al. (2011) posed a question based on the presumption that if the effects of sea spray on the air-sea fluxes are not negligible (*i.e.* , as evidenced from the background literature), then how do these fluxes affect the vertical structure of the boundary layer, and more specifically, how do they affect the vertical distribution of temperature, moisture and the turbulent fluxes of sensible heat and moisture over breaking waves. Bianco et al. (2011) used a one dimensional boundary layer model with a turbulence closure based on the Mellor-Yamada 1.5 formulation together with the sea spray model described in Kepert et al. (1999). They showed that the size of sea spray droplet increases with the increase in wind speeds, where the droplets produced with wind speeds up to 40m/s tend to evaporate thus cooling the surface layer. However, the droplets produced at higher wind speeds are much bigger in size and don’t have enough residence time before falling back in the sea, thereby facilitating the sensible heat flux exchange between the ocean and the surface layer. Also, it was shown that the suspended droplets reduce the buoyancy by increasing the density (via moisture content), thereby making the surface layer more stable. Lastly, it was shown that the droplets get displaced (facilitated by turbulence in surface layer) up to one significant wave height in the surface layer, thereby emphasising on the need to account the effects of sea spray up to one significant wave height above the ocean surface in the logarithmic relation of wind and thermal fields.

⁴ In the conditions prevalent for the genesis and propagation of a TC, the temperature of sea water is higher than that of atmospheric boundary layer next to it, as has been said in different parts of this report

In order to investigate the effects of sea spray on the TCs, Bao et al. (2011) used a sea spray model in conjunction with an uncoupled idealised WRF model following the work of Bianco et al. (2011). For their sea spray model, Bao et al. (2011) took into account the average effects of wave field (with full developed sea state assumption), thus, ignoring the effects of complex and growing wave field. For their investigation, they take the droplet source function given by Fairall et al. (2009), while noting that for better parametrisation of sea spray effects it is important to know sea spray production at various droplet radii as a function of wind speeds and wave state. Aside from the earlier conclusions of Bianco et al. (2011), they find that the surface drag values are reduced due to the spray mass loading of the surface layer. They also find that on one hand the depth of the boundary layer gets reduced due to the increase in the flow stability, while on the other hand the depth of the boundary layer increases due to the spray mediated enthalpy exchange, thus concluding that the effects of spray mass and thermal loading are in contrast with each other. The spray enhanced reduction of surface drag and the spray mediated enhanced heat flux make the boundary layer under TC more buoyant, which, then intensifies the tropical cyclone. Finally, they pointed out that the numerical coupling of sea spray effects with the wave model in a fully coupled modelling system, where the effects of wave breaking and sea spray generation on the air sea exchange of momentum and thermal fluxes, where both models (*i.e.* sea spray and wave) are interacting dynamically, still remains a complicated and challenging task.

2.6 Strategies for Model Coupling

The limitations posed by the uncoupled models in representing the processes affecting (see Section 2.5) the life cycle of a TC, as elucidated in Chapter 1, necessitate the adoption of a coupled model approach. However, the coupling of various models (*e.g.* atmosphere, ocean, wave) is not a straightforward task, as it is necessary to maintain the continuity of heat and momentum flux exchange between the respective model components.

A coupled model consists of various components, where apart from the individ-

ual atmosphere, wave and ocean model, parametrisations of turbulent air-sea flux exchange, a coupling infrastructure and a coupling algorithm are required. Over the years, numerous parametrisation schemes for air-sea flux exchange have been formulated where the most prominent and widely used ones are Fairall et al. (1996) and Smith (1988). These parametrisations are merely empirical relations obtained from the data collected from field and laboratory levels studies complemented with some modelling studies.

In regards to coupling infrastructure, several coupling libraries have been developed, including the Model Coupling Toolkit (MCT) Larson (2005), Earth System Modelling Framework (ESMF) Hill et al. (2004). A good summary of various coupling approaches and libraries can be found in Valcke et al. (2012). Despite the availability of a vast array of libraries, and the choice of a specific library for the development of a particular coupled model is subjected to the choice of programming language used by the models (ocean, atmosphere), the desired coupling algorithm (described in succeeding sections), computing platform used for deployment of such models (*e.g.* , MS Windows, Linux) and the complexity of the coupling library itself. In certain cases, it might be necessary to develop a specialised library for coupling when the existing libraries cannot be used.

Besides the uncertainties both numerical and physical present in the individual model components used in a coupled model (*e.g.* , ocean, wave), the performance of a coupled model also depends strongly on the parameters like initial conditions, coupling frequency (*i.e.* , time interval of exchange of variable among the models) reported in Lebeaupin Brossier et al. (2009) and the air-sea flux formulations as shown by Lebeaupin Brossier et al. (2008).

2.6.1 Coupling Problem

Lemarié et al. (2015) in their recent paper described the model coupling problem in detail. In Figure 2.10, we can see the description of a model coupling in a rather simple schematic. Here, the atmospheric domain is given by Ω_a , ocean domain by Ω_o , and the air-sea interface is given by Γ . The coupled earth system therefore can be

described as

$$\Omega = \Omega_a \cup \Omega_o \quad (2.18)$$

The differential operators for the model components are: Δ_a for atmosphere and Δ_o for ocean domain, and the boundary operators are Π_a , Π_o . The equation for the integration of an atmosphere model over a time period of $[0, \tau]$ can then be simply described as

$$\begin{aligned} \Delta_a U^a &= f_{atm} \\ \Pi_a U^a &= g_{atm} \\ F_a U_a &= F_{oa}(U^o, U^a, R) \end{aligned} \quad (2.19)$$

where U^a is used for variables like velocity and temperature. Furthermore, f and g represent forcing terms where g represents the forcing terms obtained from the global models, and is applied at the model boundaries. Similarly, the equation for ocean model can be written as

$$\begin{aligned} \Delta_o U^o &= f_{oce} \\ \Pi_o U^o &= g_{oce} \\ F_o U_o &= F_{oa}(U^o, U^a, R) \end{aligned} \quad (2.20)$$

F_{oa} in eqs. (2.19) and (2.20) is used to represent the air-sea flux exchange at the interface, which is carried out using parametrisation schemes in surface layer. Also, the interfacial terms are given as

$$\begin{aligned} F_a \bullet &= \rho_a K_a \partial_z \bullet \\ F_o \bullet &= \rho_o K_o \partial_z \bullet \end{aligned} \quad (2.21)$$

where K is used to represent eddy viscosity K_m and turbulent diffusivity K_t . Therefore, in a coupled model, the flux exchange between the models can be simply written as

Figure 2.10: Schematic of a coupled model framework with its non-overlapping domains

interface

$$\begin{aligned} F_a u_a^h &= F_o u_o^h \\ F_a T_a &= F_o T_o \end{aligned} \tag{2.22}$$

where superscript “h” is used to emphasise that u represent horizontal velocity. In eq. (2.22), for simplicity, it is assumed that the exchange of momentum between atmosphere and ocean takes place immediately through wave breaking. In reality, the momentum exchange will consist of two parts: (1) transferred to ocean and (2) used in wave growth. Further details on the above mathematical formulation of a coupled problem can be found in Lemarié et al. (2014, 2015).

2.6.2 Coupling Algorithms/ Schemes

There are a wide variety of coupling libraries which have been developed in order to develop multi-physics models by coupling different models together. However, in the

literature it has been noticed that the coupling algorithms used in these libraries can be broadly classified into two subcategories, namely, Asynchronous and Synchronous. These subcategories are largely done based on the features the model is being used to resolve such as diurnal pattern, future climate projections, TCs, etc.

Figure 2.11: Schematic of an asynchronous coupling algorithm. The operator $\langle \cdot \rangle_i$ indicates temporal average over the time window i .

Asynchronous Coupling

Asynchronous coupling is largely used in climate models such as those in IPCC⁵ studies, where the coupling interval varies anywhere between 2 hours to 1 day. In asynchronous type coupling, the atmospheric and oceanic models solve respective prognostic equations in their respective domains, and exchange time average quantities over a given time window (coupling interval).

The advantage of such coupling algorithm is that it uses minimal communication between models. However, from Figure 2.11, a drawback of using such an algorithm can be noted, where the values exchanged between the models result in synchronicity issues as the fluxes exchanged have a time lag, where the model (ocean model) state at the previous coupling interval is provided to the atmospheric model at the next interval (*i.e.*, the feedback only takes place at the next coupling timestep). The second drawback of this method is due to the size of coupling time interval which results in an averaged model state over a certain predetermined time window.

⁵ Intergovernmental Panel on Climate Change

Synchronous coupling

The atmosphere and ocean in reality are exchanging fluxes continuously and at a range of temporal and spatial scales. In order to establish a more accurate coupled model, it is necessary to use as small as possible coupling interval. In synchronous coupling, such coupling is carried out by using the largest time steps of the different model components being coupled. From Section 2.6.2, following can be noticed

Figure 2.12: Schematic of a synchronous coupling algorithm

$$t + N\Delta t_a = t + \Delta t_o \quad (2.23)$$

which can be further simplified as

$$N = \frac{\Delta t_o}{\Delta t_a} \quad (2.24)$$

where N is the factor by which the timestep in atmosphere model differ from that of the ocean model. The ocean and atmosphere models are integrated at their respective time steps, and at the coupling interval the fluxes are exchanged instantaneously. Although this process seems to be an appropriate solution to the coupling problem, Lemarié et al. (2015) have raised some numerical issues that arise with this type of coupling algorithm, where it is suggested to apply iterative procedure during the coupling interval to use the converged fluxes rather than the instantaneous fluxes. This algorithm is able to overcome the issue of synchronicity among the models raised by

the earlier approach. However, synchronous coupling is computationally intensive (due to increased communication between the models).

Chapter 3

Methodology

The primary objective of this study is to investigate the effects of processes at the air-sea interface on the TC. In order to carry out this investigation, it is needed to have a coupled model which is capable of simulating TCs. To that end, a coupled model was developed using respective model components for simulating atmosphere, ocean and waves. These model components were coupled using a framework which was simple yet capable enough to test different processes. This chapter discusses the different model components and the coupling framework used in the present study.

3.1 Atmosphere Model

This study uses the Advanced Research WRF (ARW) version 3.4.1 of the Weather Research and Forecasting model (Skamarock et al., 2008), which was developed for application at higher resolution. It has been extensively used in research and operational applications.

WRF model uses terrain following vertical coordinate system, where the hydrostatic pressure coordinate is given by η and is defined as

$$\eta = \frac{p_h - p_{ht}}{p_{hs} - p_{ht}} \quad (3.1)$$

where p_h is hydrostatic pressure bounded by p_{ht} at the top and p_{hs} at the bottom surface. The vertical coordinate system follow terrains near the surface, which blends with the

pressure levels near the top of the atmosphere. The model solves fully compressible, non-hydrostatic Euler equations and provides several options for modelling surface layer, cumulus convection, micro-physics, short wave and long wave radiation.

Within the WRF model, it is assumed that the reference state is in hydrostatic balance and is a function of height $\bar{z}$, so the governing equations are expressed in their perturbation form p' by removing the hydrostatically balanced $\overline{p(\bar{z})}$ portion. The momentum equations in their perturbation form are written as:

$$\begin{aligned}
 F_U &= \frac{\partial U}{\partial t} + m_x \left[\frac{\partial(Uu)}{\partial x} + \frac{\partial(Vu)}{\partial y} \right] + \frac{\partial(\Omega u)}{\partial \eta} \\
 &\quad + \frac{m_x}{m_y} \frac{\alpha}{\alpha_d} \left[\mu_d \left(\frac{\partial \phi'}{\partial x} + \alpha_d \frac{\partial p'}{\partial x} + \alpha'_d \frac{\partial \bar{p}}{\partial x} \right) + \frac{\partial \phi}{\partial x} \left(\frac{\partial p'}{\partial \eta} - \mu'_d \right) \right] \\
 F_V &= \frac{\partial V}{\partial t} + m_y \left[\frac{\partial(Uv)}{\partial x} + \frac{\partial(Vv)}{\partial y} \right] + \frac{m_y}{m_x} \frac{\partial(\Omega v)}{\partial \eta} \\
 &\quad + \frac{m_y}{m_x} \frac{\alpha}{\alpha_d} \left[\mu_d \left(\frac{\partial \phi'}{\partial x} + \alpha_d \frac{\partial p'}{\partial y} + \alpha'_d \frac{\partial \bar{p}}{\partial y} \right) + \frac{\partial \phi}{\partial y} \left(\frac{\partial p'}{\partial \eta} - \mu'_d \right) \right] \\
 F_W &= \frac{\partial W}{\partial t} + \frac{m_x}{m_y} \left[\frac{\partial(Uw)}{\partial x} + \frac{\partial(Vw)}{\partial y} \right] + \frac{\partial(\Omega w)}{\partial \eta} \\
 &\quad - m_y^{-1} g \frac{\alpha}{\alpha_d} \left[\frac{\partial p'}{\partial \eta} - \overline{\mu_d}(q_v + q_c + q_r) \right] + m_y^{-1} \mu'_d g
 \end{aligned} \tag{3.2}$$

the mass conservation equation is given as:

$$\frac{\partial \mu'_d}{\partial t} + m_x m_y \left[\frac{\partial U}{\partial x} + \frac{\partial V}{\partial y} \right] + m_y \frac{\partial \Omega}{\partial \eta} = 0 \tag{3.3}$$

and the geo-potential equation as:

$$\frac{\partial \phi'}{\partial t} + \mu_d^{-1} \left[m_x m_y \left(U \frac{\partial \phi}{\partial x} + V \frac{\partial \phi}{\partial y} \right) + m_y \Omega \frac{\partial \phi}{\partial \eta} - m_y g W \right] = 0 \tag{3.4}$$

Using the perturbation variables, the hydrostatic relation is written as:

$$\frac{\partial \phi'}{\partial \eta} = -\overline{\mu'_d} \alpha'_d - \alpha_d \mu'_d \tag{3.5}$$

The equation of state, *i.e.* the diagnostic relation for full pressure in WRF model is

expressed as:

$$p = p_0 \left(\frac{R_d \theta_m}{p_0 \alpha_d} \right)^\gamma \quad (3.6)$$

where α_d is the inverse density of dry air *i.e.* $(1/\rho_d)$ and inverse density $\alpha = \alpha_d(1 + q_r + q_v + q_c + \dots)^{-1}$, which takes into account the composition of full parcel *i.e.* the mixing ratios for rain, water vapour, cloud etc. It can be noted that the equation of state is not written using perturbation variables, due to the exponential term in the expression.

Equations (3.2) to (3.5), together with the equation of state eq. (3.6) form the set of governing equations solved in ARW. The LHS terms $F_{U,V,W}$ in eq. (3.2) include the Coriolis term, mixing term and the terms from physics parametrisations. Furthermore, within the ARW solver, in order to transform from computational space to physical (geographical) space, map scale factors (m_x, m_y) are used, where they represent the ratio of distance in computational space to the distance on earth's surface. These map scale factors are included in the governing equations by redefining the variables (u, v, w, η) to following form:

$$U = \mu_d u / m_y, V = \mu_d v / m_x, W = \mu_d w / m_y, \Omega = \mu_d \dot{\eta} / m_y \quad (3.7)$$

3.1.1 Exchange Coefficient Parametrisations

In the atmosphere model, air-sea fluxes which include wind stress, sensible heat and latent heat are the primary mechanism by which exchange of momentum, heat and freshwater (*e.g.* rain) between atmosphere and ocean takes place. These fluxes are calculated using bulk formulas, which depend on number of variables, such as wind speed, atmospheric stratification and characteristics of underlying surface (*e.g.* land surface, ocean surface).

Due to the limited range of observational data (where observation data are limited because of the difficulty in measurements during extreme condition like TC) and the uncertainties in the datasets arising from the differences in the data sources, there exists a wide array of bulk formulas. These bulk formulas differ both in terms of approximation of parameters as well as the various constants (or value of coefficients).

Due to the foregoing deficiencies and limitations, atmospheric model like WRF provides a number of bulk formulas which can be used (or tuned) so as to obtain better model results.

In the WRF model, the coefficient terms used in exchange of momentum, sensible and latent heat are commonly referred to as coefficient of drag C_d , coefficient of heat C_h and coefficient of humidity C_q . The bulk formulas can be written as:

$$\begin{aligned}\tau &= \rho \overline{(w'u')} = \rho C_d (U_{ref} - U_0)^2 \\ H_s &= \rho c_p \overline{(w'\theta')} = \rho c_p C_h U_{ref} (\theta_{ref} - T_s) \\ H_l &= \rho L_v \overline{(w'q')} = \rho L_v C_q U_{ref} (q_{ref} - q_s)\end{aligned}\tag{3.8}$$

where H_s is sensible heat, H_l is latent heat and $\overline{w'q'}$, $\overline{w'\theta'}$, $\overline{w'u'}$ are covariance terms for humidity, potential temperature and velocity respectively. These covariance terms are more commonly written using u_* , thus the terms for humidity, potential temperature and momentum become $u_* q_*$, $u_* \theta_*$, u_*^2 . The term ρ is density of air, c_p is heat capacity of moist air, L_v is latent heat of vaporisation of water, U_{ref} is wind speed at reference height (commonly taken as 10m), U_0 is the velocity at ocean surface (which is equal to zero in case of still surface), θ_{ref} is potential temperature of air at reference height, T_s is temperature of sea surface, q_{ref} is specific humidity of air at reference height and q_s is saturation specific humidity of air at temperature T_s .

Using Monin-Obukhov (MO) similarity theory (Monin and Obukhov, 1954), the terms q_* , u_* , θ_* in the eq. (3.8) can be written as:

$$\begin{aligned}u_* &= \frac{\kappa U}{\ln\left(\frac{z_{ref}}{z_0}\right) - \psi_m\left(\frac{z_{ref}}{L_0}\right)} \\ \theta_* &= \frac{\kappa(\theta_{ref} - T_s)}{\ln\left(\frac{z_{ref}}{z_T}\right) - \psi_h\left(\frac{z_{ref}}{L_0}\right)} \\ q_* &= \frac{\kappa(q_{ref} - q_s)}{\ln\left(\frac{z_{ref}}{z_Q}\right) - \psi_h\left(\frac{z_{ref}}{L_0}\right)}\end{aligned}\tag{3.9}$$

where z_0 , z_T and z_Q are the roughness length for momentum (wind stress), temperature (sensible heat) and humidity (latent heat) respectively. The terms ψ_m, ψ_h are the functions for stability correction for momentum and heat, and L_0 is the Monin-

Obukhov length scale (see eq. (2.13)). Under neutral stratification, the aforementioned functions for stability correction are:

$$\psi_m \left(\frac{z_{ref}}{L_0} \right) = \psi_h \left(\frac{z_{ref}}{L_0} \right) = 0 \quad (3.10)$$

Using eqs. (3.8) and (3.9), the expressions for the coefficients of drag, heat and humidity are as follows:

$$\begin{aligned} C_d^{1/2} &= \frac{\kappa}{\ln \left(\frac{z_{ref}}{z_0} \right) - \psi_m \left(\frac{z_{ref}}{L_0} \right)} \\ C_h^{1/2} &= \frac{\kappa C_d^{1/2}}{\ln \left(\frac{z_{ref}}{z_T} \right) - \psi_h \left(\frac{z_{ref}}{L_0} \right)} \\ C_q^{1/2} &= \frac{\kappa C_d^{1/2}}{\ln \left(\frac{z_{ref}}{z_Q} \right) - \psi_h \left(\frac{z_{ref}}{L_0} \right)} \end{aligned} \quad (3.11)$$

Furthermore, under neutral stratification, using eqs. (3.10) and (3.11), the coefficient terms are written as:

$$\begin{aligned} C_{d,N} &= \left(\frac{\kappa}{\ln \left(\frac{z_{ref}}{z_0} \right)} \right)^2 \\ C_{h,N} &= \frac{\kappa C_{d,N}^{1/2}}{\ln \left(\frac{z_{ref}}{z_T} \right)} \\ C_{q,N} &= \frac{\kappa C_{d,N}^{1/2}}{\ln \left(\frac{z_{ref}}{z_Q} \right)} \end{aligned} \quad (3.12)$$

Within the WRF model, the different bulk formulas modulate the exchange of fluxes by affecting the roughness length terms z_0 , z_T , z_Q as given in eq. (3.12). The three different bulk formulas available in WRF version 3.4.1 are:

1. Default Option

In the default option, calculation of momentum roughness length over ocean is based on Charnock (1955), where an additional viscous stress term due to molecular level interactions is added. This additional viscous stress terms also prevent roughness length becoming zero in the absence of wind. The roughness

length z_0 is written as:

$$z_0 = \frac{z_{ch}u_*^2}{g} + \frac{0.11\nu}{u_*} \quad (3.13)$$

where z_{ch} is Charnock parameter, ν is kinematic viscosity and g is acceleration due to gravity.

The thermal roughness length z_T for this option is equal to the momentum roughness length z_0 . Therefore, from eq. (3.12) it can be noted that $C_{d,N} = C_{h,N}$. Also, for the cases beside neutral stratification, an additional constraint (*i.e.* $\ln\left(\frac{z_{ref}}{z_0}\right) - \psi_m\left(\frac{z_{ref}}{L_0}\right) \geq 2$) is applied here. This constraint acts as an upper-limit on the computation of coefficient of drag C_d (and coefficient of heat C_h), such that, in the case of neutral stratification, $z_0 \leq 1.35m$.

For this **default** option, the roughness length for humidity z_Q is computed using following expression:

$$z_Q = \frac{z_0}{\left(1 + \frac{\kappa z_0 u_*}{K_a}\right)} \quad (3.14)$$

where, molecular diffusivity $K_a = 2.4 \times 10^{-5} m^2/s$. In eq. (3.14), as the second term is always positive, therefore, roughness length for humidity is less than or equal to the momentum roughness length (*i.e.* $z_Q \leq z_0$).

2. Donelan et al. (2004)

In WRF model, as option was implemented following the results of Powell et al. (2003); Donelan et al. (2004). The former study utilizing the dropsonde data collected during TC, while the latter study using laboratory scale measurements in which C_d levels off (see Figure 2.5) at wind speeds ($\geq 33m/s$) prevalent during hurricane. This is in contrast to the previous **default** option where C_d monotonically increases with wind speed.

The momentum roughness length z_0 here is computed as follows:

$$\begin{aligned}
 z_0 &= \max\{1.27 \times 10^{-7}, \min[z_w z_2 + (1 - z_w) z_1, 2.85 \times 10^{-3}]\} \\
 z_w &= \min\left(1, \left[\frac{u_*}{1.06}\right]^{0.3}\right) \\
 z_1 &= \frac{0.011 u_*^2}{g} + 1.59 \times 10^{-5} \\
 z_2 &= \frac{10}{\exp(9.5 u_*^{-0.333})} + \frac{1.65 \times 10^{-6}}{\max(u_*, 0.01)}
 \end{aligned} \tag{3.15}$$

In eq. (3.15), the momentum roughness length z_0 is bounded (*i.e.* $1.27 \times 10^{-7} \leq z_0 \leq 2.85 \times 10^{-3}$), where the upper and lower bounds are obtained from Davis et al. (2008). The roughness length for temperature and humidity are set equal to a constant value of $0.95 \times 10^{-4} m$ following Large and Pond (1982). The roughness lengths z_T, z_Q despite being set to a constant value, the coefficient terms C_h, C_q will still vary with wind speed, due to their dependence on the coefficient of drag (see eq. (3.12)).

3. Garratt (1992)

In the third option, momentum roughness length z_0 is calculated using eq. (3.15), thus giving same values of z_0 as the previous formulation. However, the roughness lengths z_T, z_Q here are calculated following Brutsaert (1975), and can be written as:

$$\begin{aligned}
 z_T &= z_0 \exp\left(-\kappa(7.3 Re_*^{1/4} Pr^{1/2} - 5.0)\right) \\
 z_Q &= z_0 \exp\left(-\kappa(7.3 Re_*^{1/4} Sc^{1/2} - 5.0)\right)
 \end{aligned} \tag{3.16}$$

where roughness Reynolds Number $Re_* = z_0 u_* / \nu$. The terms Pr, Sc are Prandtl number and Schmidt number, which following Garratt (1992) are set to 0.71, 0.6 respectively. Furthermore, the constants 7.3, 5.0 are experimental values, which can be tuned to improve model results.

3.2 Ocean Model

MIKE3 (DHI, 2012b) is a three dimensional hydrodynamic model developed for the applications in oceanographic and coastal environment. It is an unstructured grid model which solves Reynolds Averaged Navier-Stokes (RANS) equations using hydrostatic and Boussinesq assumptions. MIKE3 model uses finite volume approximation for spatial discretization, and has the capability to use two different vertical grids *i.e.* terrain following sigma layers, and hybrid sigma-z layers, where z-layers are used for higher depths, while sigma layers for shallower depth and free surface. The sigma layers allow for increased resolution in areas of particular interest, for example in the areas of strong mixing, and near the surface. The structure of vertical levels is user controlled where, it is possible to use variable or fixed spacing between both the sigma and z-layers.

The governing equations used in MIKE3 model consist of continuity equation, momentum equations, equation of state, and scalar transport equation for temperature and salinity. The continuity equation is written as:

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = S \quad (3.17)$$

and the momentum equations in two horizontal direction are:

$$\begin{aligned} \frac{\partial u}{\partial t} + \vec{v} \cdot \nabla u &= f v - g \frac{\partial \eta}{\partial x} - \frac{1}{\rho_0} \frac{\partial p_a}{\partial x} \\ &\quad - \frac{g}{\rho_0} \int_z^\eta \frac{\partial \rho}{\partial x} dz + F_u + \frac{\partial}{\partial z} \left(v_t \frac{\partial u}{\partial z} \right) + u_s S \\ \frac{\partial v}{\partial t} + \vec{v} \cdot \nabla v &= -f u - g \frac{\partial \eta}{\partial y} - \frac{1}{\rho_0} \frac{\partial p_a}{\partial y} \\ &\quad - \frac{g}{\rho_0} \int_z^\eta \frac{\partial \rho}{\partial y} dz + F_v + \frac{\partial}{\partial z} \left(v_t \frac{\partial v}{\partial z} \right) + v_s S \end{aligned} \quad (3.18)$$

the scalar transport equations for temperature and salinity as:

$$\begin{aligned} \frac{\partial T}{\partial t} + \vec{v} \cdot \nabla T &= F_T + \frac{\partial}{\partial z} \left(D_v \frac{\partial T}{\partial z} \right) + \hat{H} + T_s S \\ \frac{\partial s}{\partial t} + \vec{v} \cdot \nabla s &= F_s + \frac{\partial}{\partial z} \left(D_v \frac{\partial s}{\partial z} \right) + s_s S \end{aligned} \quad (3.19)$$

where, the horizontal diffusion terms $F_{T,s}$ are defined as:

$$(F_T, F_s) = \left[\frac{\partial}{\partial x} \left(D_h \frac{\partial}{\partial x} \right) + \frac{\partial}{\partial y} \left(D_h \frac{\partial}{\partial y} \right) \right] (T, s) \quad (3.20)$$

and horizontal stress terms $F_{u,v}$ in momentum equations (eq. (3.18)) have expression of:

$$\begin{aligned} F_u &= \left[\frac{\partial}{\partial x} \left(2A \frac{\partial u}{\partial x} \right) + \frac{\partial}{\partial y} \left(A \left(\frac{\partial u}{\partial y} + \frac{\partial v}{\partial x} \right) \right) \right] \\ F_v &= \left[\frac{\partial}{\partial x} \left(A \left(\frac{\partial u}{\partial y} + \frac{\partial v}{\partial x} \right) \right) + \frac{\partial}{\partial y} \left(2A \frac{\partial v}{\partial y} \right) \right] \end{aligned} \quad (3.21)$$

Here, $f = 2\Omega \sin(\phi)$, where Ω is earth rotational speed, ϕ is geographical longitude, ρ_0 is reference density of water, p_a is atmospheric pressure, ν_t is vertical eddy viscosity, η is surface elevation and (u_s, v_s) are the river discharge velocity in ambient water of the point source with magnitude of discharge given by S , while, T_s, s_s are the temperature and salinity of the point source. The horizontal momentum equation eq. (3.18) given above are written without the radiation stress terms, which can be applied by inclusion of wave model (MIKE21 SW) coupling with the MIKE3. The radiation stress term in zonal direction (*i.e.* west-east) is:

$$- \frac{1}{\rho_0 h} \left(\frac{\partial s_{xx}}{\partial x} + \frac{\partial s_{xy}}{\partial y} \right) \quad (3.22)$$

whereas in meridional direction (*i.e.* south-north) is:

$$- \frac{1}{\rho_0 h} \left(\frac{\partial s_{yx}}{\partial x} + \frac{\partial s_{yy}}{\partial y} \right) \quad (3.23)$$

where, $s_{xx}, s_{xy}, s_{yx}, s_{yy}$ are the components of the radiation stress tensor. The total water depth h in eqs. (3.22) and (3.23), is defined as the algebraic sum of surface elevation η and still water depth d (*i.e.* $h = \eta + d$) and can be obtained by vertically integrating the local continuity equation:

$$\frac{\partial h}{\partial t} + \frac{\partial h \bar{u}}{\partial x} + \frac{\partial h \bar{v}}{\partial y} = hS + \hat{P} - \hat{E} \quad (3.24)$$

where $\hat{P}, \hat{E}$ are the rates of precipitation and evaporation and $\bar{u}, \bar{v}$ are the depth averaged

velocity in zonal and meridional direction respectively. Lastly, in MIKE3 model, with the assumption of incompressibility, the density ρ only depends on the temperature T and salinity s , and is defined using equation of state:

$$\rho = \rho(T, s) \quad (3.25)$$

where the UNESCO equation of state (UNESCO, 1981) is used.

MIKE3 model provides several modules for modelling processes like oil spill, sand and mud transport, as well as temperature and salinity. It is also possible to include the effects of tides and wave radiation stresses. These modules and effects can be incorporated within a simulation, however the inclusion of these modules depends on the process to be studied. For the model physics, MIKE3 model provides two different numerical schemes, namely "Lower order" and "Higher order", where lower order is a first order scheme while the higher order scheme is a second order Runge-Kutta method for time integration. For vertical mixing with MIKE3 model, $k - \epsilon$ model can be used for turbulence closure. This turbulence closure model was chosen because it is the only option available in the MIKE3 model.

3.2.1 Surface Heat Exchange

Here, we briefly describe the processes by which exchange of fluxes of momentum, mass and heat between atmosphere and MIKE3 takes place. The primary processes by which heat exchange are calculated as follows:

1. net short wave radiation SW_{net}
2. net long wave radiation LW_{net}
3. sensible heat flux H_s
4. latent heat flux H_l

Here, the fluxes of sensible heat, latent heat and long wave radiation occur at the surface, where the atmosphere-surface interface is assumed to be very thin. Moreover, the absorption profile of the short wave radiation in MIKE3 model is approximated

using Beer's law

$$I(d) = (1 - \beta)I_0 \exp^{-\lambda d} \quad (3.26)$$

where, d is depth below the surface, λ is light extinction constant, I is light intensity (subscript "0" represents the value just below the ocean surface) and β quantifies the fraction of light energy absorbed near the surface. The terms β , λ are usually tuned to improve model results, where the typical values of β are 0.2 – 0.6, whereas that of λ are $0.5 - 1.4m^{-1}$. Furthermore, from eq. (3.26), it can be noted that the light absorbed near the surface is βI_0 , which decreases exponentially with the increase in depth d .

At the surface, the net heat flux can be written as:

$$H_{net} = H_s + H_l + \beta \cdot SW_{net} + LW_{net} \quad (3.27)$$

and, the source term $\hat{H}$ given in scalar transport equation for temperature eq. (3.19) in two dimensional calculation is:

$$\hat{H} = \frac{H_s + H_l + \beta \cdot SW_{net} + LW_{net}}{\rho c_p} \quad (3.28)$$

while, the source term $\hat{H}$ for three dimensional calculation is:

$$\begin{aligned} \hat{H} &= \frac{\partial}{\partial z} \left(\frac{SW_{net}(1 - \beta) \exp^{-\lambda(\eta - z)}}{\rho c_p} \right) \\ &= \frac{SW_{net}(1 - \beta) \exp^{-\frac{\lambda(\eta - z)}{\lambda}}}{\rho c_p} \end{aligned} \quad (3.29)$$

where vertical coordinate z in MIKE3 goes from η at the surface to $-d$ at the bottom boundary.

Sensible Heat Flux H_s

The convective heat flux in MIKE3 model is computed using following relation:

$$H_s = \begin{cases} \rho_a c_p C_{h,w} U_{ref} (T_{ref} - T_s) & T_{ref} \geq T_s \\ \rho_a c_p C_{h,c} U_{ref} (T_{ref} - T_s) & T_{ref} < T_s \end{cases} \quad (3.30)$$

where $C_{h,w}$, $C_{h,c}$ are coefficients for heat for warming and cooling respectively and set equal to 0.0011. As the convective heat flux transfer takes place both by means of free and forced convection, therefore to make sure that there is free convection in the absence of wind, wind speed is bounded as $U_{ref} = \max(U_{ref}, 2)$.

Latent Heat Flux H_l

In MIKE3 model, the latent heat flux is parametrised as:

$$H_l = L_v C_q (a_1 + b_1 U_2) (q_s - q_{ref}) \quad (3.31)$$

where U_2 is wind speed at 2 meter above sea surface, coefficient of humidity $C_e = 1.32 \times 10^{-3}$ and a_1, b_1 are tunable constants with typical values of 0.5, 0.9 respectively. In MIKE3 model, currently, it is not possible to directly provide the specific humidity terms q_s, q_{ref} , however, they can be computed from the temperature of water and air using following expressions:

$$\begin{aligned} q_i &= \frac{0.2167}{T_i + T_k} e_i \\ e_s &= 6.11 \exp^{5418} \left(\frac{1}{T_k} - \frac{1}{T_s + T_k} \right) \\ e_{ref} &= RH 6.11 \exp^K \left(\frac{1}{T_k} - \frac{1}{T_{ref} + T_k} \right) \end{aligned} \quad (3.32)$$

where subscript i refers to both air (ref) and water (s), reference temperature $T_k = 273.15K$, e_s refers to saturation vapour pressure, while e_{ref} is vapour pressure of air and RH is relative humidity. Lastly, as the wind speed from atmosphere model is traditionally provided at 10 meter above the sea surface, therefore U_2 is obtained using eq. (2.6).

Finally, it is necessary to mention that, in the present study, a coupled model is developed, where the net radiation fluxes (*i.e.* long wave and short wave) are directly provided from the atmosphere model to the ocean model and no further computation are carried out in MIKE3 model, therefore it is not necessary to provide details of radiation flux computation in MIKE3 model. However, interested readers are referred to DHI (2012b) for further details.

3.3 Wave Model

MIKE21 spectral wave model (DHI, 2012a) is a third generation wave model used to simulate wind generated wave fields in water depths ranging from open ocean to coastal areas and lakes. It solves the wave action balance equation using unstructured grid with finite volume discretisation in space. The model is capable of taking into account the effect of white-capping, directional variation of wind, depth induced breaking, energy dissipation due to bottom friction, non-linear energy transfer due to quadruplet and triad wave-wave interaction.

In Cartesian coordinates, two dimensional wave action balance equation can be written as:

$$\frac{\partial N}{\partial t} + \frac{\partial c_x N}{\partial x} + \frac{\partial c_y N}{\partial y} + \frac{\partial c_\sigma N}{\partial \sigma} + \frac{\partial c_\theta N}{\partial \theta} = \frac{\mathbf{S}}{\sigma} \quad (3.33)$$

here, $c_{x,y,\sigma,\theta}$ are wave propagation velocity in x, y directions, in frequency and in directional space respectively. The first term in eq. (3.33) represents time rate of change of wave action density, second and third terms represents the propagation of action density in x, y direction, whereas fourth term gives the shift in relative frequency and fifth term represents the diffraction and refraction processes. The term $\mathbf{S}$ on RHS represents the source terms, which include the effects of wave generation due to wind, and dissipation due to wave breaking, bottom friction, white-capping as well as the energy transfer due to non-linear processes.

Similar to ocean model, spectral wave model also provides two different numerical schemes *i.e.* "Lower order" and "Higher order" scheme for both space and time.

3.3.1 Source terms

As mentioned earlier, the term $\mathbf{S}$ in eq. (3.33) represents the superposition of various physical processes as described earlier. The source terms can be written as:

$$\mathbf{S} = S_{in} + S_{nl} + S_{ds} + S_{bot} + S_{surf} \quad (3.34)$$

Here, S_{in} represents the energy input by wind, S_{nl} represents the non-linear wave-wave interaction which redistributes the energy between different scales (wave frequency),

S_{ds} is the source term for wave energy dissipation due to white-capping, S_{bot} is the dissipation of wave energy due to bottom friction and S_{surf} represents the wave energy dissipation due to depth induced wave breaking.

For the purpose of this study, energy input by wind and energy dissipation by white-capping are the two dominant physical processes, as we aim to study the effects of interaction between atmosphere and ocean. Since the wind input source term S_{in} has been described in considerable details in Section 2.2, therefore, for the sake of brevity we will limit our description to the source term for the wave energy dissipation by white-capping.

Combining the effects of pressure induced decay, short wave attenuation due to large white-caps and wave steepness dependent white-cap coverage, Komen et al. (1984) introduced a dissipation function in terms of mean frequency. Group, The Wamdi (1988) subsequently modified the dissipation source function to be in terms of wave number. The source function can be written as:

$$S_{ds}(f, \theta) = -C'_{ds} \left(\frac{\hat{\alpha}}{\hat{\alpha}_{pm}} \right)^m \frac{k}{\bar{k}} \bar{\sigma} E(f, \theta) \quad (3.35)$$

here, C'_{ds} and m are the fitting constants, $\hat{\alpha}$ is the overall wave steepness, $\hat{\alpha}_{pm}$ is the value of $\hat{\alpha}$ for the Pierson-Moskowitz spectrum (Pierson and Moskowitz, 1964), $\bar{\sigma}$ is the mean relative angular frequency and $\bar{k}$ is the mean wave number. The overall wave steepness is defined as:

$$\hat{\alpha} = \bar{k} \sqrt{E_{tot}} \quad (3.36)$$

After the introduction of wind input source term S_{in} following Janssen (1989), the source term for white-cap dissipation S_{ds} was adjusted to obtain a proper balance between wind input and dissipation source term at high frequencies (Komen et al., 1996). The modified source term is given as:

$$S_{ds}(f, \theta) = -C_{ds} \left(\frac{\hat{\alpha}}{\hat{\alpha}_{pm}} \right)^m \left[(1 - \delta) \frac{k}{\bar{k}} + \delta \left(\frac{k}{\bar{k}} \right)^2 \right] \bar{\sigma} E(f, \theta) \quad (3.37)$$

where C_{ds} , δ , m are constant terms with $m = 4$ and $\delta = 0.5$. Additionally, the first term in eq. (3.37) is modified by defining a tunable constant $C_{ds}^* = C_{ds} / (\hat{\alpha}_{pm})^4$ and is set to

4.5. The mean relative angular frequency $\overline{\sigma}$ and mean wave number $\overline{k}$ in eqs. (3.35) to (3.38) are defined as:

$$\begin{aligned}\overline{\sigma} = 2\pi\overline{f} &= 2\pi \left(\frac{\int_0^{2\pi} \int_0^\infty E(f, \theta) (f)^{p_\sigma} df d\theta}{\int_0^{2\pi} \int_0^\infty E(f, \theta) df d\theta} \right)^{p_\sigma} \\ \overline{k} &= \left(\frac{\int_0^{2\pi} \int_0^\infty E(f, \theta) (\overline{k})^{2p_k} df d\theta}{\int_0^{2\pi} \int_0^\infty E(f, \theta) df d\theta} \right)^{2p_k}\end{aligned}\tag{3.38}$$

where $p_\sigma = p_k = -1$. The integrals in eq. (3.38) are computed by splitting wave spectrum into modelled (prognostic) and unresolved part (diagnostic). For further information, readers are referred to DHI (2012b). Lastly, in order to improve wave model results in mixed seas (*i.e.* when wind sea and swells are present), Bidlot et al. (2007) proposed a new set of values for the tunable constants used in eqs. (3.37) and (3.38). In the present study, we employed the values given by Bidlot et al. (2007), where $p_\sigma = p_k = 1$, and $C_{ds}^* = 2.1$ and $\delta = 0.6$.

3.4 Coupled Model Framework

In this section, we briefly describe the atmosphere and ocean model coupling. This is accompanied by the coupled-model grid configuration, variables exchanged between models, initial and lateral boundary conditions used in the respective models.

3.4.1 Model Coupling Procedure

The coupling of respective atmosphere and ocean models require the use of third party libraries. In recent coupled-model studies, experiments have been conducted using a wide array of model coupling libraries, where WRF-ARW (Skamarock et al., 2008) has been coupled with Regional Ocean Model System (ROMS) (Haidvogel et al., 2008) using Model Coupling Toolkit (MCT) to study TC-like vortex generation

Figure 3.1: The atmosphere-ocean-wave model coupling framework

(Warner et al., 2010). Other recent studies such as (Chen and Curcic, 2016) used WRF-ARW and HYCOM model within University of Miami Coupled Model developed with ESMF, whereas (Smith et al., 2013) employed ESMF for developing the TC version of coupled atmosphere/ ocean mesoscale prediction system (COAMPS).

In all the above studies, a substantial modification to the source codes of different model components is needed in order to develop a coupled model whereas studies like (Bi et al., 2013) used CERFACS OASIS3.2-5 coupler (Valcke et al., 2012) for developing Australian Community Climate and Earth System Simulator coupled model (ACCESS-CM) which couples UK Met Office Unified model with Modular Ocean Model (MOM). At run time, OASIS3 in the coupled model remains as a separate executable which receives, interpolates and sends the coupled variables to respective sub-components which are also run as separate executable. This technique requires minimal modification to source code of the model components, in order to establish a coupled model.

In current study, the coupled model is developed using a technique similar to Bi et al. (2013), and the coupling is carried out using synchronous coupling algorithm (see ??). The outline of the complete model is shown in Figure 3.1, while the coupler

Figure 3.2: The detailed illustration of model coupling interface

is presented in a bit more detail in Figure 3.2. From Figure 3.1, we can see the different variables being passed between the respective models as indicated by the respective colors. The coupled model proceeds as follow:

1. The atmosphere model integrated from $t = t_0$ to $t = t_0 + \Delta t$ where t is the model time, t_0 is initial time and Δt is the coupling time step. This yields the atmospheric variables which are required to force the ocean (or wave) model.
2. The atmosphere variable required by the respective model (*e.g.* 10 meter wind speed (U_{10}, V_{10}), 2-meter air temperature (T_{2m}), 2 meter mixing ratio (Q_{2m}), the net downward long wave radiation (LW), short wave radiation flux (SW), Evaporation rate $\hat{E}$, Precipitation rate $\hat{P}$) are interpolated to the ocean (or wave) model grid by calling module for **Regridding**.
3. The wind speed components U_{10}, V_{10} are adjusted to neutral stratification by calling module for **Stability Correction**.
4. The ocean (or wave) model then integrates from $t = t_0$ to $t = t_0 + \Delta t$ and

calculates the ocean (or wave) state at the next coupling time step.

5. The ocean (or wave) state calculated by the ocean (or wave) model is then interpolated to the atmospheric model grid by calling module for **Regridding**.
6. Any component (*i.e.* **Sea Spray model**, **Wave Boundary layer model**) dependent on either of the model components (*i.e.* atmosphere, ocean or wave) are invoked before calling the atmosphere model.

Within this coupling, there are number of options implemented, where if the ocean model grid dimensions are smaller than the atmosphere model grid, then at the atmosphere grid with no ocean model cells, respective variables (if available from the global datasets are applied).

3.5 Coupled Model Setup

The atmospheric component of the coupled model is configured using the nesting approach with Lambert Conformal grid projection available within WRF model. The outer (coarse) domain "domain 01" in the model covers the western north Atlantic ocean together with the parts of Caribbean sea and Gulf of Mexico (see Figure 3.3). Within this coarse domain, a higher resolution domain "domain 02" with 223×256 cells is nested which covers the east coast of USA. The "domain 01" consists of 170×170 cells and spans from 100°W to 55°W in longitudinal and from 13°N to 45°N in latitudinal direction. This outer domain has a horizontal resolution of 21.6km and uses 41 vertical η levels stretching from surface to 100 hPa¹ at the model top. The nested domain "domain 02" has a horizontal resolution of 7.2km and also uses 41 vertical η levels. The 41 η levels² are prescribed such that there are at least 10 levels within 1km from ground level with lowest level at height 8m.

The initial and boundary conditions for the atmosphere simulations were taken from Modern-Era Retrospective analysis for Research and Applications, version 2

¹ Hectopascal

² 1.000, 0.998, 0.996, 0.992, 0.989, 0.983, 0.977, 0.969, 0.962, 0.954, 0.945, 0.934, 0.909, 0.880, 0.845, 0.807, 0.765, 0.719, 0.672, 0.622, 0.571, 0.520, 0.468, 0.420, 0.376, 0.335, 0.298, 0.263, 0.231, 0.202, 0.175, 0.150, 0.127, 0.106, 0.088, 0.070, 0.055, 0.040, 0.026, 0.013, 0.000

Figure 3.3: The atmosphere model domain, where shading represents the terrain height. The polygon outlines represent the extent of parent domain (domain 01) and child domain (domain 02)

(MERRA-2) (Bosilovich et al., 2015) dataset, which has a horizontal resolution of $0.5^\circ \times 0.625^\circ$ (approximately 50km in latitudinal direction). Due to the computational constraints and the need to perform multiple simulations, we chose to use the one-way nesting approach, where the feedback from nested domain (*i.e.* "domain 02") to the parent domain (*i.e.* "domain 01") was turned off. Moreover, the lateral boundary conditions from the MERRA-2 to the parent domain were supplied at every 6h interval, while that from the parent domain to the inner (or nested) domain at every 1h interval. Furthermore, grid nudging was applied to the outer domain, where the model results were nudged towards the gridded data (*i.e.* MERRA-2) in order to minimize the model drift from the large scale flow. In the atmosphere model, the static fields like terrain (see Figure 3.3) data was Global Multi Resolution Terrain Elevation Data (30 arc second USGS³ GMTED 2010, Danielson and Gesch (2011)), while the vegetation data was Moderate Resolution Imaging Spectroradiometer (MODIS) Fraction of Photosynthetically Active Radiation (FPAR) product (Myneni et al., 2002)

³ United States Geological Survey

and the land use data was 21-class International Geosphere Biosphere Programme (IGBP) MODIS product (Friedl et al., 2010).

Table 3.1: Summary of the physics options used in atmosphere model simulations

Physics	Schemes
Microphysics	Ferrier scheme (Rogers et al., 2001)
Cumulus Parametrisation	Simplified Arakawa-Schubert (SAS) scheme (Han and Pan, 2011)
Planetary Boundary layer	Yonsei University (YSU) scheme (Hong et al., 2006)
Surface layer	MM5 surface layer scheme (Beljaars, 1995)
Land Surface	Unified NOAH land surface model (Tewari et al., 2004)
Long wave radiation	RRTM scheme (Mlawer et al., 1997)
Short wave radiation	Dudhia scheme (Dudhia, 1989)

The various atmospheric processes within WRF are modelled using different physics schemes. These schemes can be combined in a number of ways to model the atmospheric event both effectively and efficiently. The aforementioned schemes vary from simple (*i.e.* computationally inexpensive) to rather comprehensive (and computationally expensive) representation of physical processes. In the present study, we used Ferrier scheme (Rogers et al., 2001) for microphysics, while Simplified Arakawa-Schubert (SAS) scheme (Han and Pan, 2011) for cumulus parametrisation, where the cumulus scheme is used to parametrise sub-grid scale effects of clouds such as updrafts, downdrafts within the clouds which are not modelled by the microphysics scheme. The long wave radiation in present study was parametrised using Rapid Radiative Transfer Model (RRTM) scheme (Mlawer et al., 1997), while the short wave radiation was parametrised using Dudhia Scheme (Dudhia, 1989) and the land surface was modelled using Unified "Noah" (Tewari et al., 2004) scheme. The Planetary Boundary layer (PBL) was parametrised using Yonsei University Hong et al. (2006) scheme together with the accompanying Monin-Obukhov theory based surface layer scheme. These physics schemes are summarised in Table 3.1.

The domain for ocean and wave model consists of unstructured triangular mesh with varying grid resolution (see Figure 3.4). It can be clearly noticed from Figure 3.4 (right) that, coarser resolution is used in outer domain while higher resolution is used in areas near the coastlines (and shallow areas). The model grid for wave model

Figure 3.4: The unstructured grid used in ocean (or wave) model with varying grid resolution, where high resolution grid in coastal (and shallow) areas, while coarse resolution elsewhere. A zoomed in view of Chesapeake Bay is shown on left.

consists of 34405 triangular cells, while that of ocean model consists of 1174346 prismatic cells. Due to the high resolution of mesh in coastal areas, it is rather difficult to visualize the grid, therefore a zoomed in view of region close to Chesapeake Bay and Cape Hatteras is shown in Figure 3.4 (left). From Figure 3.4, it can also be noticed that the unstructured grid is able to resolve the features of coastline very well, which would not be otherwise possible using a structured without substantial increase in computational cost (Pallares et al., 2017). Furthermore, unstructured grid reduces the number of telescopic grids (a traditional compromise is to use nested structured grid) as compared to a structured grid (see (Pallares et al., 2017)).

Figure 3.5 shows the bathymetry data used in the ocean (or wave) model, together with the outline of the inner domain ("domain 02") used in the atmosphere model. It should be pointed out that, in the present study, only the inner domain of the atmosphere model is used in the coupled model. Furthermore, from Figure 3.5, it can be noticed that the ocean (or wave) model grid doesn't span the whole bottom surface of the atmosphere model. In the areas beyond the ocean (or wave) model grid, the values of sea surface temperature (or roughness length) were obtained from

Figure 3.5: The ocean (and wave) model domain with bathymetry (shading). The extent of "domain 02" are represented by the polygon outline.

global datasets (or using default WRF parametrisation). The bathymetry data for the ocean (or wave) model was obtained from General Bathymetric Chart of the Oceans (GEBCO⁴) 30 arc second dataset. Due to the low resolution of GEBCO data (≈ 1 km horizontal resolution) in the coastal areas, data from 3 arc second (≈ 100 m horizontal resolution) U.S. Coastal Relief model (CRM⁵) was also used.

As described in Section 3.2, within MIKE3 model, there are multiple options for specifying the vertical grid. In the present study, a hybrid z - σ vertical grid with 44 levels was used, as it can provide higher vertical resolution in the mixed layer while coarser resolution in the deeper ocean. The aforementioned vertical grid used 20 high resolution terrain following σ layers (Haidvogel et al., 2008) within the 50 m depth from the ocean surface while the 24 z -layers in the rest of the water column. The spectral grid in wave model consists of 24 equally spaced directional bins with bin width of 15° and 31 logarithmically spaced frequency bins (ranging from 0.04-0.7 Hz).

⁴ <http://www.gebco.net>

⁵ <https://www.ngdc.noaa.gov/coastal/crm.html>

Table 3.2: Details of tidal and three dimension ocean (spinup) model setup

Parameters	Tidal model	Ocean model (spinup)
2D/3D	2D	3D (44 levels ^a)
Atmospheric Forcing	None	WRF model
River Forcing	None	None
Baroclinic Forcing	None	MyOcean data ^b
Tidal Forcing	DTU tidal model	BC ^c from tidal run

^a 20 σ -level and 24 z-level

^b Initial and Boundary Condition

^c Boundary Condition

As described in Section 3.3, the wave model (*i.e.* MIKE 21 SW) uses source terms for modelling different physical processes (see eq. (3.34)). Besides the different physics options, the wave model also provides two different model formulations (DHI, 2012a), which are a directionally decoupled parametric formulation and a fully spectral formulation. In the present research, we used the fully coupled formulation, where the directional-frequency spectrum is a dependent variable (see eq. (3.33)). As for the physics options, the wind input source term S_{in} used Janssen's formulation (Janssen, 1991), while the whitecap dissipation term S_{ds} used modified version of Komen's formulation (Komen et al., 1996) with the new set of coefficients obtained by Bidlot et al. (2007) (see Section 3.3.1). The source term for wave energy dissipation due to depth induced breaking S_{surf} was modelled using (Battjes and Janssen, 1978), the source term for energy dissipation due to bottom friction S_{bot} was based on Johnson and Kofoed-Hansen (2000); Weber (1991), and the non-linear energy transfer due to four wave interaction S_{nl} was represented by the discrete interaction approximation (Sell and Hasselmann, 1972). Lastly, the lateral boundary conditions for the wave model were obtained from the well validated Integrated Ocean Waves for Geophysical and Other Applications (IOWAGA) (Stopa et al., 2016) global wave hindcast, which is prepared using WAVEWATCH-III wave model (Tolman et al., 2002). The foregoing wave hindcast model uses winds from Climate Forecast System Reanalysis (CFSR) data (Saha et al., 2014).

For the ocean model (MIKE3), the initial conditions *i.e.* ocean temperature, salinity and lateral boundary conditions which include two dimensional vertical distribution of zonal and meridional velocity, ocean temperature, salinity were obtained from

Table 3.3: Summary of spectral wave and three dimensional ocean model setup

Parameters	Ocean model	Wave model
Vertical grid	44 layers ^a	N/A
Timestep	300 seconds (baroclinic) 60 seconds (barotropic)	90 seconds
Boundary Forcing	MIKE3 spinup (IC ^b) MyOcean data (BC)	Coldstart (IC) Wavewatch III (BC)
Tidal Forcing	BC from tidal run	N/A
Turbulence Model	$k - \epsilon$ scheme	N/A
Spectral grid	N/A	24 (directions ^c) 31 (frequencies ^d)

^a Same as Table 3.2

^b Initial Condition

^c Equally spaced direction bin width 15°

^d Logarithmically spaced frequency bins with minimum frequency of 0.04Hz and frequency factor 0.01

the operational Mercator global ocean analysis and forecast system disseminated via MyOcean service⁶, product: (GLOBAL_ANALYSIS_FORECAST_PHY_001_024). It is a global product on a 1/2° horizontal resolution and 50 vertical levels. It is prepared using an implementation of Nucleus of European Modelling of the Ocean (NEMO⁷) model (Madec, 2008), with a variational data assimilation for profiles of salinity and temperature together with the sea surface elevation obtained from along track satellite altimetry.

Furthermore, the ocean component included both the tidal and baroclinic boundary conditions, where the tidal boundary conditions were generated using an implementation of MIKE21 HD model. The tidal model domain covered an area larger than the grid used for the ocean component of the coupled model. This is to avoid shallow water depths when placing the open boundaries in tidal model, and to avoid any unforeseen errors which might arise when obtaining the tidal boundary conditions for the ocean model simulations. This tidal model was run using only the tidal elevations at the open boundaries, where the tidal elevations were generated using the DTU10 Tidal Model⁸ (Cheng and Anderson, 2010). Following Bolaños et al. (2014), the tidal elevation and depth averaged velocities thus obtained from the tidal model were linearly combined

⁶ <http://www.myocean.eu.org>

⁷ <https://www.nemo-ocean.eu/>

⁸ ftp.space.dtu.dk/pub/DTU10/DTU10_TIDEMODEL

with the surface elevation and two dimensional (in vertical) zonal and meridional velocities obtained from MyOcean dataset, and were applied to the ocean model using Flather boundary condition (Flather, 1976). The ocean model used Smagorinsky formulation for the horizontal eddy viscosity, while $k - \epsilon$ turbulence model for the vertical, and a constant value of roughness height for bottom friction. Also, tidal potential (Pugh, 1996) was applied in all the ocean model simulations. The model setup used in tidal model run and ocean model spinup is summarised in Table 3.2. Lastly, in the present study, the ocean model was initialized from rest, therefore, it (*i.e.* ocean model) was run for a period of 20 days (spinup period) before the start date of coupled model simulations. During this spinup period, the atmosphere forcing for the ocean model was obtained by running uncoupled atmosphere model for the same duration. This was done to avoid any biases that might arise due to differences in atmospheric forcing during spinup period and coupled model simulations.

Figure 3.6: An illustration of grid mask used for remapping between atmosphere and ocean (or wave) model, with zoomed in view of Chesapeake Bay (on left) where unstructured grid overlaid showing the disparity between grids.

As alluded above and in preceding section (Section 3.4), the various model components (*i.e.* atmosphere, ocean and wave) are dynamically coupled using a model coupler shown in Figures 3.1 and 3.2. The various components within the coupled model setup maintained their own timesteps, while the model coupler provided the

model coupling timestep. For the various coupled model simulations performed in the present research, the model coupling time step was set to 1800 seconds (*i.e.* when the variables between the model components were exchanged). The model setup used in wave and ocean model during coupled model experiments is described in Table 3.3. The model time step for atmospheric component was set to 30 seconds (*i.e.* variables were exchange after 60 atmospheric model time steps). The model time step for wave and ocean model components was 1800 seconds respectively. Additionally, the time step for solving shallow water equations was set to 60 seconds, while the time step for transport equations was set to 300 seconds. Lastly, every 30 minutes, the output from each of the model components were stored on their native grids.

Using Figure 3.6, the differences between the grid used in ocean (and wave) model compared to that in atmosphere model are further illustrated. Figure 3.6 (right) shows grid structure used for exchanging variables between the different model components, where the "dark" color shows the exchange grid. The variable remapping between the model components facilitated by this exchange grid used bilinear interpolation (Jones, 1998). In the zoomed in view of area close to Chesapeake Bay and Cape Hatteras (Figure 3.6 (left)), the complexity of the coupled model system build using components with different grid structure is further highlighted. It must also be mentioned that when the variables from ocean (or wave) model grid were remapped onto the atmosphere model grid, the land/sea mask in atmosphere model was preserved. However, when the variables from atmosphere model were remapped onto the ocean (or wave) model grid, preserving the land/sea mask in atmosphere model resulted in an additional constraint, *i.e.* the output from the exchange grid did not cover all the grid cells in ocean (or wave) model domain (see Figure 3.6 (left)). Therefore, to avoid any inconsistencies, an additional remapping step using nearest neighbour interpolation (Watson, 1999) was also applied.

3.6 Model Application

3.6.1 Synopsis of Hurricane Arthur (2014)

Figure 3.7 taken from Berg (2015) shows the best track data for the Hurricane Arthur, the first named storm of 2014 Atlantic hurricane season. Figure 3.7 (a) displays the best track positions of Hurricane Arthur, with the minimum sea level pressure (Figure 3.7 (b)) and maximum surface wind speed (Figure 3.7 (c)) attained by the hurricane during the course of its lifetime.

Hurricane Arthur was first identified as a tropical depression on 0000 UTC⁹ July 1, 2014 by National Hurricane Centre (NHC) while it was located 70 nautical miles north of Freeport, Bahamas (Berg, 2015). It subsequently upgraded to tropical storm on 1200 UTC July 1, 2014. By then the depression has drifted westward to 60 nautical miles east of Ft. Pierce, Florida. Arthur while located in a weak mid-level steering meandered east of Florida till July 2. On July 2, a mid-level anticyclone developing over the western Atlantic caused Arthur to track northward, whereby encountering low upper level winds and warmer ocean temperature ($> 28^{\circ}$), Arthur strengthened while located east of Florida. It subsequently upgraded to a hurricane on 0000 UTC July 3, 2014 (thick red line in Figure 3.7 (a)) located 125 nautical miles east-southeast of Savannah, Georgia.

Later that day, Arthur turned north-northeastward, accelerating while moving between a ridge over western Atlantic and a mid to upper level trough over the east United States (Berg, 2015). It continued to strengthen and reached its peak intensity of 85 knots (see Figure 3.7 (c)) at 0000 UTC July 4, 2014 just off the coast of North Carolina. At 0315 UTC on July 4, 2014 it made landfall just west of Cape Lookout, North Carolina. After landfall, and crossing Outer Banks, Arthur accelerated northeastward over the western Atlantic on July 4 and early July 5. It subsequently transitioned to an extra-TC (thin black line in Figure 3.7 (a)) at 1200 UTC July 5 just west of Nova Scotia.

⁹ Coordinated Universal Time

(a)

(b)

(c)

Figure 3.7: (a) The best track positions of hurricane centre, (b) Selected pressure observations overlayed with the best track minimum sea level pressure and (c) Selected wind observations plotted with the maximum surface wind speed from best track for Hurricane Arthur from 1-5 July 2014. The dashed vertical line represent 0000 UTC, while the solid vertical lines represent landfall. (Source: Berg (2015))

3.7 Measurement Datasets

Here, we briefly describe the different observational datasets used to evaluate the model results obtained from the model simulations carried out over the period of this study.

3.7.1 Satellite Measurements

As described in Cavaleri and Sclavo (2006), several altimeters have been deployed onboard satellites starting from Geosat in 1985. These altimeters provide data wind speed and wave height at an interval of 6-7 km along the satellite track. These datasets are essentially an instantaneous spatial average of the data within the footprint of the altimeter. This footprint can vary from a diameter of 3km in small wave conditions to 10km in large wave conditions (Durrant et al., 2009). The wind speed data from the altimeter data is not reliable outside the range 2-20 m/s because of the limitations of the retrieval algorithms (Cavaleri and Sclavo, 2006). Furthermore, the wave height data obtained from satellite altimetry loses its validity at wave height greater than 20m (Cavaleri and Sclavo, 2006).

Figure 3.8: The ground track of three altimeters: JASON-2 (red), CRYOSAT-2 (blue) and SARAL/AltiKa (red) over a period of one day.

The altimeter data used in this study was obtained from French Research Institute for Exploitation of the Sea (IFREMER). This dataset was a blended product of

10 different satellite altimeter (Queffeuilou and Croizé, 2017), however, due to the differences in the operational history of aforementioned satellites, only three satellite altimeters were operational during the simulation period considered here. These three satellite altimeters were onboard JASON-2, CRYOSAT-2 and SARAL/AltiKa. The acquisition and processing of JASON-2 data are described in Picot et al. (2008); Abdalla et al. (2010), while that of CRYOSAT-2 in Francis (2007); Queffeuilou and Croizé (2017) and of SARAL/AltiKa in Sepulveda et al. (2015). Figure 3.8 presents the typical coverage by respective satellite, where the ground track of all three satellite are overlaid together. Lastly, we would like to highlight that the altimeter data facilitates both spatial and temporal comparisons despite the scantness of the observations.

3.7.2 Station Measurements

In addition to the satellite measurements, there are number of buoys moored with the model domain used in the present study. These measurements are generously made available by National Buoy Data Centre (NDBC) through their website¹⁰. The data obtained from buoys is traditionally considered to be of very high quality, with good temporal resolution.

Figure 3.9 shows the location of the NDBC buoys considered here with the details of the station location and water depth described in Table 3.4. Additionally, Figure 3.9 also illustrates the location of buoys relative to the track of the hurricane. The detailed description of the data acquisition and validation is provided in Earle et al. (1999). These buoys besides providing data for atmospheric variables like wind speed, wind

Table 3.4: Details of the NDBC buoys used for model comparison

Buoy Name	Latitude (°N)	Longitude (°W)	Water Depth (m)
41002	31.470	74.840	3980
41004	32.501	79.099	38.4
41008	31.400	80.868	18.288
44014	36.611	74.843	44.5

direction, air temperature, pressure and dew point temperature, also measure wave height, wave period and directional wave spectrum which can be used to compare the

¹⁰ www.ndbc.noaa.gov

Figure 3.9: The various observation data used for model comparison. The buoys described in Table 3.4 are marked with black squares. The flight pattern of two aircraft missions executed at 1637 UTC 2 July (red) and 1422 UTC 4 July (green) 2014 are shown together with the best track of Hurricane Arthur (2014).

output of wave model. Additionally, these buoys provide the water temperature just below the surface of the ocean.

3.7.3 Airborne Measurements

Besides the satellite and buoy data described in preceding sections, in the present research, the observation dataset collected by the aircraft missions conducted by Hurricane Research Division (HRD) at Atlantic Oceanographic and Meteorological Laboratory (AOML), National Oceanic and Atmospheric Administration (NOAA) was also used. These aircraft missions are flown under the Hurricane Field Program of the Hurricane Research Division, and the data is disseminated through their website¹¹.

The aircraft missions are routinely carried out using NOAA WP-3D and United

¹¹ http://www.aoml.noaa.gov/hrd/programs_sub/HFP.html

Figure 3.10: An artistic illustration of the aircraft flying through a hurricane, where the stepped frequency microwave radiometer (SFMR) instrument is mounted under the aircraft wing. A plot of wind speed measured by SFMR during an aircraft mission flown in Hurricane Frances (2004) is shown on the left, where high wind speed region in eyewall with relatively calm eye can be noticed. (Source: NOAA HRD http://www.aoml.noaa.gov/hrd/about_hrd/HRD-P3_sfmr.html)

States Air Force (USAF) C-130 aircrafts. The in-situ measurements provided by these aircraft missions include ground and flight level winds from SFMR (Uhlhorn and Black, 2003; Uhlhorn et al., 2007), ocean temperature measurements from Airborne eXpendable BathyThermograph (AXBT) and vertical profiles of several atmospheric variables from GPS dropsondes. In addition to these measurements, NOAA WP-3D also measures the vertical slices through hurricanes using downward looking doppler radar. Figure 3.9 indicates the typical flight paths during these aircraft missions, where different patterns are due to the landfall of hurricane. Lastly, a typical wind speed observation made by SFMR is displayed in Figure 3.10, clearly showing the high wind speeds in eyewall region, with a relatively calm eye.

It is also necessary to point out that the observation data made available by HRD

is quality controlled and post-processed (*e.g.* remapping on radar data onto a gridded space in earth coordinates), therefore no further quality checks were carried out during the course of this study.

To assess the validity of the model results, typical error metrics of root mean square error (*RMSE*), scatter index (*SI*), bias and Pearson correlation coefficient (*R*) are defined as follows:

$$\begin{aligned}
 Bias &= \frac{1}{N} \sum_{i=1}^N (A_i - B_i) \\
 RMSE &= \sqrt{\frac{1}{N} \sum_{i=1}^N (A_i - B_i)^2} \\
 R &= \frac{\sum_{i=1}^N (A_i - \bar{A}) (B_i - \bar{B})}{\sqrt{\sum_{i=1}^N (A_i - \bar{A})^2 (B_i - \bar{B})^2}} \\
 SI &= \frac{1}{\bar{B}} \sqrt{\frac{1}{N} \sum_{i=1}^N [(A_i - \bar{A}) - (B_i - \bar{B})]^2}
 \end{aligned} \tag{3.39}$$

where model estimates are expressed as B_i while the observed data is expressed as A_i . The overbar denotes the mean value and N is the number of observations.

Chapter 4

Impact of Waves on Air-Sea Momentum Exchange

4.1 Preamble

Cavaleri et al. (2012) provide a schematic of generation and growth of waves due to wind, together with the motivation to include the effects of wind waves in a coupled climate model system. The two systems namely atmosphere (air) and ocean (water) are coupled by a thin viscous layer where the waves generated by the air blowing across a fully flat surface (*i.e.* interface), results in a two fluid instability known as Kelvin Helmholtz instability. This instability grows with the wind, causing differential surface pressure which enhances the transfer of energy and momentum from wind to waves. This enhanced transfer of energy results in growth of waves, where waves evolve in length and height and eventually get modulated by processes like wave breaking and non-linear interactions.

In an uncoupled atmosphere model, to account for the effects of waves in air-sea interaction process, the effects of waves are parametrised in terms of localized wind aloft the surface as evident from the empirical parametrisation (see section 2.2). From Chapter 2 and the simple schematic briefly described above, it can be construed that there is an obvious correlation between wind and waves. However, a contradiction to the above assumption, are the locations where waves are present when wind is not blowing (see section 2.3). The second limitation to the above assumption is the lack

of a unique correlation between wind and waves thereby limiting the accuracy and applicability of localized wind based parametrisation.

Furthermore, in an uncoupled ocean model traditionally a wind-to-current momentum transfer type approach is adopted, where it is assumed that the momentum transfer from atmosphere to waves is immediately transferred to ocean. As mentioned in Cavaleri et al. (2012), if wind and waves were always in equilibrium, with constant wave conditions, the assumption of immediate transfer of momentum from atmosphere to ocean currents (*i.e.* overlooking the waves in a coupled climate system) would have been valid. However, the wind input to waves and wave energy loss due to wave breaking are two separate well defined quantities.

Due to these considerations, wave models have been recognized as an essential element of a coupled model framework, as was shown by Fairall et al. (2011), where it was observed that including wave information for the processes that depends on waves, results in better agreement. The earliest implementation of a coupled wind-wave model was carried out by Janssen (1989, 1991) in the European Centre for Medium Weather Forecasts (ECMWF) model, where using results from the model implemented over the whole globe, it was shown that including wave effects in the model, affected the climate throughout the troposphere as well as in both the hemispheres. Furthermore, the sea state dependent momentum exchange resulted in reduction in wind speeds in the southern hemisphere.

It is widely accepted that the interaction between the waves and the lowest atmospheric layer is stronger in growing seas (*i.e.* young sea), where the waves are steep, thence a higher surface drag. The influence of this enhanced surface drag due to the surface waves results in obvious reduction in wind speeds. Additionally, the frequent breaking of these steeper waves air may aid in further increase in surface drag by upto a factor of 30, Sullivan et al. (2004). Furthermore, at the air-sea interface, processes such as white-capping, sea-spray and gas (aerosols) exchange also depend on wave breaking. Some of the important parameters affecting these aforementioned processes are the length of white cap, lifetime of white cap, and the amount of air entrained by white cap resulting in bubbles (sea spray).

In the tropical and extra-tropical zones, with the exception of tropical cyclones

and typhoons, sea surface is usually dominated by swells generated by distant storms for example cyclones. These swell waves affect the local conditions in two different ways, where on one hand, they affect the atmosphere above by transferring momentum resulting in wave driven wind as shown in Hanley and Belcher (2008); Song et al. (2015), while on the other hand, these swell waves induce currents in ocean by means of stokes drift. These swell dominated processes are independent of local winds, so their effects can't be parametrised in terms of wind speeds.

From the above discussion, it can be concluded that wave breaking at the air-sea interface, resulting in momentum transfer from atmosphere to ocean. Furthermore, the momentum transfer from atmosphere to ocean (and wave breaking) affects the upper ocean layer by injecting turbulence which enhances the mixing in the upper ocean layer, thus affecting the stratification in the mixed layer.

Besides the momentum exchange, at the air-sea interface, SEF¹ *i.e.* latent heat and sensible heat flux, provides another crucial link between atmosphere and ocean. Contrary to the momentum exchange parametrisation, besides the local wind speed, enthalpy fluxes parametrisation depend on ASTD² (sensible heat) and ASMD³ (latent heat). In the conventional parametrisations, it is implicitly assumed that the ocean surface is flat. However, in the presence of surface waves, it can be argued that a large extent of ocean surface would be exposed, thus facilitating the exchange of SEF. Also, when the surface waves are breaking, they result in production of sea spray, which further enhances the sea surface in contact with air, thus increased SEF.

In summary, SEF at the air-sea interface is the primary source of energy for tropical cyclone (TC) while momentum exchange is the primary source of cyclone dissipation. From the earlier discussion, we can construe that the surface waves play a large role in modulating the flux exchange (both heat and momentum) at the air-sea interface, therefore in order to improve the prediction and modelling of TC, it is imperative to include the effects of surface waves in a coupled model framework.

¹ surface enthalpy fluxes

² air-sea temperature difference

³ air-sea moisture difference

4.2 Numerical Experiments

To investigate the sensitivity of modelled hurricane to the different roughness length parametrisations discussed in Section 3.1.1, a set of numerical experiments (see Table 4.1) were carried out. These model simulations were initialised on 0000 UTC June 30, 2014 and were integrated for 120 hours.

Figure 4.1: The wind speed (U_{10}) dependence of (a) roughness length for temperature z_T (thick) and humidity z_Q (thin) - note the logarithmically scaled y axis; (b) coefficients of heat C_h (thick) and humidity C_q (thin). The roughness length z_0 is calculated using wind input source function S_{in} and a parametric JONSWAP spectrum. The option 0, 1 and 2 are marked with red $\square$, green Δ and blue ∇ with solid, dashed and dotted line.

To illustrate the differences between the different roughness length parametrisations, z_T, z_Q obtained from respective formulations are shown in Figure 4.1. As discussed in Chapter 3, in this study, momentum roughness length z_0 depends on the sea state (*i.e.* computed using wave model), therefore, z_T, z_Q in Figure 4.1, depend on the wind input source function S_{in} used in the wave model component. In order to use the wind input source function S_{in} , a two dimensional JONSWAP spectrum with cosine squared directional spreading was used. To discern the effects of coupling wave model, an uncoupled atmosphere model simulation (Run 1) was also carried out. From Section 3.1.1, the default option in WRF model sets $z_H = z_0$, therefore the thick solid line in Figure 4.1 (a) shows the wind speed dependence of both momentum roughness length and the roughness length for heat. For the different parametrisations besides the uncoupled simulations, momentum roughness length was obtained from the wave model, therefore, the thick solid line in Figure 4.1 represents z_0 (Figure 4.1

(a)) and coefficient of drag C_d (Figure 4.1 (b)) for Runs 2-4 summarised in Table 4.1.

Table 4.1: Summary of the different numerical experiments utilizing the flux options in a coupled atmosphere-wave model.

Runs	Isftcflx	Wave Coupling	Note
1	1	No	Uncoupled run with roughness lengths z_0, z_T, z_Q (uncapped) using Donelan et al. (2004) option
2	0	Yes	Coupled run with roughness length z_0 from wave model and z_T, z_Q (uncapped) using default option
3	1	Yes	Coupled run with roughness length z_T, z_Q (uncapped) using Donelan et al. (2004) option
4	2	Yes	Coupled simulation with z_T, z_Q (uncapped) using Garratt (1992); Brutsaert (1975)

From Figure 4.1, it can be noted that z_T, z_Q are constant for the option Donelan et al. (2004), however, C_h, C_q increase with the increase in wind speed, due to their dependence on the C_d (see eqs. (3.11) and (3.12)). When parametrising z_Q and C_q , it can be noted that the default option gives lower value compared to Donelan et al. (2004). To put it in perspective, for the same ASMD, the latent heat flux due to the default option will be lower than Donelan et al. (2004) (see eq. (3.8)). Finally, when Garratt (1992) is used in the coupled model simulations, it can be clearly noted that, contrary to previously discussed options, the value of roughness lengths and exchange coefficients decrease with increase of wind speed. Furthermore, it is also obvious that for the same value of ASTD and ASMD, this option will give much lower latent and sensible heat flux, thus potentially affecting the hurricane intensity and structure.

4.3 Results and Discussion

4.3.1 Temporal Evolution of TC

The track and intensity (*i.e.* minimum sea level pressure (MSLP) and maximum surface wind speed (V_{max}) obtained from the different numerical experiment (see Table 4.1)

are shown in Figure 4.2, with their respective position and intensity bias in Figure 4.3. The error metrics were computed using the best track data for Arthur (2014).

Even though the model simulations were started at 0000 UTC June 30, 2014 and integrated for a duration of 120 hours, the results are shown starting from 0000 UTC July 2, 2014. This was done to make sure that all the simulations have attained a $V_{max} \geq 17ms^{-1}$.

Figure 4.2: (a) Simulated tracks of Hurricane Arthur using Run 1 (solid), Run 2 (dashed), Run 3 (dash dotted) and Run 4 (dotted) starting from 0000 UTC July 2 to 1200 UTC July 4, 2014. Markers indicate hurricane centres at 6h interval; (b) Minimum sea level pressure at hurricane centre for the runs as in (a); As in (b) but for maximum wind speed at 10m

TC track

The track of the TC obtained from the numerical experiments show limited sensitive to both the different parametrisations and the coupling of wave model (Figure 4.2 (a)). Despite the limited sensitivity, it is worth noting that the weaker intensity TC (*i.e.* Run 1 and Run 4) traverse to the west of the relatively stronger TC (*i.e.* Run 2 and Run 3). It is worth mentioning that the numerical simulations being discussed here use same source of SST, therefore, the track differences shown here must be due to the differences in the roughness length parametrisations.

Figure 4.3 further illustrates the bias in TC centre location (see Figure 4.3 (upper)),

Figure 4.3: The bias in hurricane centre location (upper) and maximum wind speed (lower) for the simulation runs summarised in Table 4.1

where the bias is the distance between the modelled TC centre locations compared to the TC centre location in best track data. All the numerical experiments performed show a track bias of less than 100nm, with the largest track bias for uncoupled (Run 1) model, while the smallest bias for Run 3. It is noteworthy that the stronger TCs (Run 2 and 3) have a lower position bias, implying that the best track was to the east of the modelled track, with the Run 2 giving the best results. Furthermore, it can also be said that the coupling of wave model results in improved TC track, however only at the later stage of the model simulations.

TC intensity

Figure 4.2 (b) shows the MSLP, while Figure 4.2 (c) shows the V_{max} obtained from the different numerical experiments. Firstly, it can be clearly noticed that MSLP and V_{max}

are very sensitive to the parametrisations used. The strongest TC was obtained from Run 3, while the weakest from Run 1 (*i.e.* uncoupled). However, it must be pointed out that V_{max} obtained from Run 3 is quite similar to the uncoupled model, though MSLP is lower in Run 3.

Unlike the improvements in TC track, Figure 4.3 (lower) shows that the stronger TCs (Runs 2 and 3) have higher intensity bias of upto 20 knots in the earlier stage of the TC lifetime, where the modelled TCs were over-intensified. Also, it is noticeable that the uncoupled (Run 1) model simulation gives the lowest intensity. Comparing Figures 3.7 and 4.3, it can be realized that coupling wave model results in a stronger TC, when the TC (as observed in best track) is weaker, while lower intensity bias, when the TC is stronger. It should also be mentioned that the weaker TCs have a westward bias *i.e.* weaker TCs traverse over the land surface, which have relatively higher roughness length, therefore dissipates more momentum and weakens the TC. This can be attributed to the lower V_{max} that persists in the modelled TCs after 0000 UTC July 4, 2014.

Wind and Wave Measurements

To further examine the sensitivity of different parametrisations and using a sea state dependent momentum roughness lengths, a comparison of wind speed, wind direction, significant wave height H_s and average wave period T_{02} at various buoy locations described in Table 3.2 are given in Figures 4.4 to 4.7.

Also, from Figure 3.9, it is evident that the NDBC buoy 44014 was closest to the hurricane track. From the buoys considered here, only NDBC buoy 41002 was located to the east of TC track, while the rest were located to the west. Furthermore, compared to NDBC 41002, the other three buoys were located in the shallower water depths. The NDBC buoy 41002 was included to highlight the performance of model setup in areas not affected by TC. Figures 4.4 and 4.6 clearly show that the model performed very well at NDBC 41002, where a good correlation (see Table 4.2) was obtained for modelled wind speeds and wave height. When comparing the modelled wave heights and wind speeds obtained at NDBC 41004 and 41008, the effects of westward bias in TC track are quite evident. The model results suggest that NDBC

Figure 4.4: Time series of wind speed (m/s) at buoys 41002 (top left), 41008 (top right), 44014 (bottom left) and 41004 (bottom right) for the time period from 0000 UTC July 2 through to 1200 UTC July 4, 2014.

Figure 4.5: As in Figure 4.4 but for wind direction

buoy 41004 experiences the passage of TC eye, and the wind speed and wave height (Figures 4.4 and 4.6) associated with the eye wall region, which is not present in

the observed value. Additionally, it can also be seen that the modelled peak wind speeds (and wave heights) are observed 6h before the peak wind speeds in observed data. When comparing the wind direction (Figure 4.5) at buoys 41004 and 41008, the bias due to storm translation speed can also be noticed, where the change of wind direction due to arrival and departure of TC (or storm) takes place 6h prior to that in observed results. Table 4.2 summarises the performance of four model experiments described in Table 4.1. At buoy 44014, the modelled wind speed and wave heights

Figure 4.6: As in Figure 4.4 but for significant wave height H_s (m)

show the best correlation for Run 3. This can be attributed to the track bias described in earlier section, where Run 3 gives the least track bias. Also, the bias in wind speed and wave heights due to storm translation speed can also be noticed at buoy 44014. Finally, comparing the modelled and observed wave period (Figure 4.7) at the buoy locations, it was found out that besides the model results at buoy 41008, the wave model performed well. The anomalous passage of eye wall at 41004 can also be noticed in the results of wave period, where besides the foregoing anomaly, model results were very well correlated with the observations. However, similar to the wind speed and wave height results at 44014, the best correlation was obtained from Run 3.

Figure 4.7: As in Figure 4.4 but for average wave period T_{02} (sec)

4.3.2 Exchange Coefficients

Following the potential intensity theory of Emanuel (1986), the maximum intensity that a steady state hurricane, in gradient and hydrostatic balance can achieve is given as:

$$V_{max} = \frac{C_k}{C_d} \left(\frac{1 - 0.25\gamma r_0^2}{1 - \frac{1}{2} \frac{C_k}{C_d} \gamma} \right) \quad (4.1)$$

where

$$\gamma = A \left(\frac{1 - RH}{1 - A \cdot RH} \right) \quad (4.2)$$

$$A = \frac{T_s - T_0}{T_s} + \frac{\chi_s}{R_d T_s (1 - RH)}$$

Here, r_0 is the normalised outer radius of the TC where the surface wind speeds vanish, RH is the ambient relative humidity, T_s and T_0 are the surface and outflow temperature, χ_s is the subcloud-entropy variable and R_d is the universal gas constant for dry air. The eq. (4.1) underscores the role C_k/C_d plays within the context of the intensification model briefly described in section 2.5.1.

Figure 4.8 compares the coefficient of drag, coefficient of enthalpy and the ratio

Table 4.2: Comparison of modelled wind speed U_{10} , significant wave height H_s and wave period T_{02} from different runs summarised in Table 4.1 with NDBC buoys 41002, 41004 and 44014

Buoy 41002						
Runs	$U_{10} [ms^{-1}]$		$H_s[m]$		$T_{02}[sec]$	
	RMSE	R	RMSE	R	RMSE	R
1	0.87	0.94	0.41	0.95	0.35	0.89
2	0.76	0.96	0.42	0.94	0.37	0.87
3	0.87	0.96	0.39	0.96	0.38	0.89
4	0.92	0.95	0.47	0.95	0.44	0.91
Buoy 41004						
1	6.21	0.76	1.44	0.87	0.52	0.90
2	6.95	0.74	1.60	0.87	0.65	0.91
3	6.86	0.73	1.40	0.88	0.57	0.90
4	5.75	0.75	1.30	0.86	0.57	0.91
Buoy 44014						
1	4.44	0.81	1.18	0.81	0.72	0.84
2	3.84	0.76	0.86	0.89	0.70	0.90
3	3.11	0.93	0.50	0.97	0.51	0.95
4	3.29	0.84	0.88	0.84	0.68	0.87

C_k/C_d obtained from the different numerical experiments to illustrate the influence of coupling wave model (see Figure 4.8 (a,c,e)) and the different flux parametrisations (see Figure 4.8 (b,d,f)). The results presented in Figure 4.8 are for the whole domain (only grid points used in coupled model), and correspond to the peak intensity (see Figure 4.2) attained by the TC in the respective numerical experiment.

The key difference between the uncoupled (Run 1) and coupled model (Run 3) simulation was the method used to compute momentum roughness length, where in the former case z_0 was parametrised using eq. (3.15), while in the latter case was obtained from wave model. Furthermore, in the present study, the upper limit for z_0 given in eq. (3.15) was removed, thus allowing a consistent comparison between Runs 1 and 3. Following Green and Zhang (2013), the total enthalpy flux using eq. (3.8) can be defined as:

$$E = \rho C_k U \left(c_p(\theta_{ref} - T_s) + L_v(q_{ref} - q_s) \right) \quad (4.3)$$

where $C_k \approx C_h \approx C_q$. The coefficient of enthalpy C_k in Figure 4.8 follows eq. (4.3),

Figure 4.8: The effects of coupling wave model on (a) coefficient of drag C_d , (c) coefficient of enthalpy C_k and (e) the ratio of exchange coefficients C_k/C_d for uncoupled (Run 1) in black (+) and coupled (Run 3) runs in green ($\circ$). The effects of different flux options on (b) coefficient of drag C_d , (d) coefficient of enthalpy C_k and (f) the ratio of exchange coefficients C_k/C_d . The red Δ indicate results from Run 2, whereas green $\circ$ indicate results from Run 3, while the blue ∇ indicate results from Run 4. The abscissa in Figures (a-f) refer to wind speed U_{10} , and the results are shown at 1800 UTC July 3, 2014.

and the underlying assumption that $C_k \approx C_h$, where C_h can be obtained from eq. (3.8) by plugging in the sensible heat flux H_s computed by atmosphere model. It is important to keep in mind that the results shown in Figure 4.8 depend both on the flux parametrisations and the model response to these parametrisations. The coupling of wave model results in higher coefficient of drag C_d compared to the uncoupled model simulations. The C_d values obtained from coupled model simulations Figure 4.8 (a,b) exhibit saturation at wind speed values greater than 32 m s^{-1} , which is consistent with the observation data of Powell et al. (2003); Donelan et al. (2004). However, it is worth mentioning this saturation exhibited in the coupled model simulations was not a result of an upper bound normally applied in uncoupled atmosphere models. The higher values of C_d suggest increased dissipation due to friction, while also alluding

to the increased frictional convergence in the regions of higher wind speeds.

To better understand the results of C_k shown in Figure 4.8 (d), it is important to keep in mind, how z_T is parametrised in the various flux options. The roughness length z_T in Run 2 is set equal to z_0 , while that in Run 3 is set equal a constant value, and in Run 4 is computed from eq. (3.17). Now, from eq. (3.8), for the same $\Delta\theta$, it can be said that Run 2 will have the highest sensible heat flux, while Run 4 will have the lowest H_s . In Runs 2-4, z_0 depend on the sea state, which was modelled using a wave model, thus implying the same C_d for the different coupled model simulations. Now comparing the ratio C_k/C_d , the effects of different parametrisations can be seen (Figure 4.8 (f)), where similar to C_k , Run 2 has the highest ratio C_k/C_d , with the lowest for Run 4. Moreover, the ratio C_k/C_d obtained from Runs 3-4 are consistent with the measurements of Black et al. (2007), while that in Run 2 are substantially higher. The ratio term can be taken as a proxy for the ratio of energy gain (sensible heat flux) to the energy dissipated (surface friction), where the decrease in ratio with the wind speed implies an increase in energy dissipation compared to the enthalpy gain, with vice-versa for the increase in ratio.

4.3.3 Surface Level Variability

The effects of both the sea state mediated momentum flux and the different parametrisations schemes on the structure and evolution of the TC are explored by using radius-time Hovmöller diagrams of the azimuthally averaged surface level momentum flux (u_*), SEF (*i.e.* H_s and H_l), wind speeds (*i.e.* tangential V_t and radial V_r velocity component) and equivalent potential temperature (θ_e) obtained from the different numerical experiments summarised in Table 4.1. These radius-time diagrams provide insights on both temporal and spatial (radial direction) evolution, thereby allowing a comparison both within a model simulation and between different model simulations.

Figure 4.9: Time-radius cross-section of azimuthally averaged (a) friction velocity u_* ($m s^{-1}$), (b) sensible heat flux H_s ($W m^{-2}$), (c) latent heat flux H_l ($W m^{-2}$), (d) tangential velocity V_t ($m s^{-1}$), (e) radial velocity V_r ($m s^{-1}$), and (f) equivalent potential temperature θ_e (K) for uncoupled model (Run 1). The contours of tangential velocity (d) and radial velocity (e) were plotted at $4 m s^{-1}$ interval while that of equivalent potential temperature θ_e at $5 K$ interval. The equivalent potential temperature θ_e was calculated following Bolton (1980). The radial distance (abscissa) is from TC centre and time (ordinate) from 0000 UTC July 2 through to 1200 UTC July 4, 2014.

Surface fluxes

From Figure 4.9 (a) and Figure 4.10, it is evident that sea state mediated momentum flux is higher compared to the uncoupled model for all the flux parametrisations considered in this study. From the radius-time diagrams, it can also be noted that higher surface flux region are closer to the TC eye wall with the peak values at approximately 50 km from the TC centre. It is needless to say that these high surface

Figure 4.10: Hovmöller diagrams of azimuthally averaged friction velocity u_* (m s^{-1}) for (a) Run 2, (b) Run 3 and (c) Run 4. The radial distance (abscissa) is from TC centre and time (ordinate) from 0000 UTC July 2 through to 1200 UTC July 4, 2014.

Figure 4.11: As in Figure 4.10 but for sensible heat flux H_s (W m^{-2})

flux regions coincide with the regions of higher wind speeds.

It is also worth mentioning that the analysis period considered here spans the entire intensification stage of the TC till the landfall, as well as some duration after the TC emerge over the Atlantic ocean after the landfall. Now, when investigating the effects of different numerical experiments on the SEF (*i.e.* sensible and latent heat flux), some interesting observations can be made. Firstly, it is observed that the latent heat flux H_l

Figure 4.12: As in Figure 4.10 but for latent heat flux H_l (Wm^{-2})

Figure 4.13: As in Figure 4.10 but for tangential velocity V_t (ms^{-1}). The contour interval for V_t is $4 ms^{-1}$

(Figure 4.9 (c) and Figure 4.12) is one order of magnitude greater than the sensible heat flux H_s (Figure 4.9 (b) and Figure 4.11). Secondly, it is evident that coupling wave model not only affects the momentum flux, but also the surface enthalpy fluxes, where the coupled model runs (Runs 2-4) have consistently higher fluxes except for Run 4, which is attributed to the significantly lower C_q (see Figure 4.1 (b)) compared to Run 1. Moreover, as expected the sensible heat flux is substantially higher for Run 2 as compared to the Run 3, which is believed to be due to higher C_h . However,

Figure 4.14: As in Figure 4.10 but for radial velocity V_r (ms^{-1}). The countour interval for V_r is $4 ms^{-1}$.

Figure 4.15: As in Figure 4.10 but for equivalent potential temperature θ_e (K). The contour interval for θ_e is 5 K.

contrary to the sensible heat flux, the latent heat flux is substantially higher in Run 3 with a broader spread as compared to Run 2, which can also be explained by the different flux parametrisation used in Runs 2 and 3.

This aforementioned dichotomy in the modelled values of the two components of SEF due to the different flux parametrisation alludes to the varying potential of H_s and H_l in affecting the storm intensity. Also, from Figure 4.2, it is known that the Run

3 gives the stronger TC as compared to Run 2.

Surface levels winds and θ_e

The structural differences in the surface fluxes noted in preceding section warrants an analysis of the both dynamical and thermodynamical fields obtained from the model simulations. The dynamical quantities presented here are the tangential and radial components of surface level winds as opposed to the zonal and meridional components. This is because the tangential wind component represents the primary circulation in the TC, while the radial component represents the wind component directed inwards towards the TC centre. For the thermodynamical field, the equivalent potential temperature θ_e obtained from the different model simulations is compared, where higher θ_e implies warmer and moister air (*i.e.* includes the effect of both temperature and moisture).

It is observed that the low level wind components (*i.e.* V_t and V_r) obtained from Run 1 (Figure 4.9 (d,e)) and Run 3 (Figure 4.13 (c) and Figure 4.14 (c)) are somewhat similar, with one noticeable difference, where TC intensification initiates approximately 5h earlier in Run 4 as compared to that in Run 1. When comparing the θ_e obtained from the aforementioned runs, similar to the heat fluxes, the effect of flux parametrisation is quite evident, where the region with warmer and moister air is much broader with higher θ_e in the TC inner region for Run 1 as compared to Run 4 owing to the higher values of flux coefficients used in Run 1.

The surface level winds obtained from Runs 2 and 3 show that the tangential winds are much stronger for Run 3 (Figure 4.12 (b)) compared to that in Run 2 (Figure 4.12 (a)), with much stronger inflow (Figure 4.14 (a,b)). It is necessary to keep in mind that these differences are primarily due to the flux exchange coefficients (or SEF) discussed in preceding section, where higher SEF leads to stronger TC. This is further illustrated in the θ_e plots shown in Figure 4.15, where Run 3 has comparatively warmer and moister air in the inner TC region compared to Run 2. It is worth noting that the fields obtained from Runs 2 and 3 also exhibit some structural differences, whereas Run 2 has a slightly wider inner region compared to Run 3, where the wider inner region are traditionally associated with weaker TC.

Figure 4.16: Radius-height plots valid at 1200 UTC July 3, 2014 of azimuthally averaged (a) tangential velocity V_t (shaded) with contours at an interval of 5 m s^{-1} , (b) radial velocity V_r (shaded) with contours at an interval of 2 m s^{-1} , (c) radar reflectivity (shaded) with contours of vertical velocity at an interval of 0.3 m s^{-1} , and (d) absolute angular momentum AAM with contours at an interval of $6 \times 10^5 \text{ m}^2 \text{ s}^{-1}$. The fields are plotted for uncoupled model run (Run 1).

4.3.4 TC vertical structure

In addition to the radial variation of the surface fields, the effects of coupling wave model and the different flux parametrisations on the vertical structure of TC are explored by means of Radius-Height plots of azimuthally averaged wind velocity components (*i.e.* tangential and radial components), radar reflectivity and vertical velocity are shown in Figures 4.16 to 4.19. The Radius-Height plots for uncoupled run (Run 1) are valid at 1200 UTC July 3, 2014 while those for the coupled model runs (Runs 2-4) are shown at 1200 UTC and 2100 UTC July 3, 2014. The latter time instant is just before the landfall. In addition to the aforementioned quantities, a plot of absolute angular momentum AAM (Figure 4.16 (d)) is also shown, where AAM is defined as:

$$AAM = rV_t + \frac{1}{2}fr^2 \quad (4.4)$$

Figure 4.17: Radius-height plots of tangential velocity V_t with contours at an interval of 5 ms^{-1} for (a,b) Run 2, (c,d) Run 3, and (e,f) Run 4, where column on left is valid at 1200 UTC July 3, 2014 while on the right is valid at 2100 UTC July 3, 2014.

here r is the radial distance from the TC centre, and f is the Coriolis parameter. From the comparison of primary circulation (*i.e.* tangential wind component) done using Figure 4.16 (a) and Figure 4.17, it can be said that the Runs 1 and 4 give a structurally similar TC, with slightly higher peak tangential winds in Run 4. One of the prominent features is that the maximum tangential wind speeds (Figures 4.16 and 4.17) occur within the 3km above the sea surface. This region of maximum V_t occur just above the top of strong inflow layer induced by the friction (Figures 4.16 and 4.18). It should also be noticed that in all the coupled model runs, the coefficient of drag C_d depends

Figure 4.18: As in Figure 4.17 but for radial velocity V_r with contours at an interval of 2 m s^{-1} .

on the sea state modelled using wave model, therefore the structural differences noted in Figures 4.17 and 4.18 are primarily due to the response of TC to the applied SEF (or the flux parametrisations). These applied SEF increase the TC intensity, which in turn increases the friction.

The structural similarities between the primary circulation obtained from Run 1 and Run 4 noted above are not as pronounced in the secondary circulation (*i.e.* radial wind component) shown in Figure 4.16 (b) and Figure 4.18. A physical explanation for this behaviour would be the increased gradient wind imbalance in the boundary layer⁴ due to the increased friction (from wave model). Smith et al. (2014) argued

⁴ boundary layer here refers to a shallow layer with strong inflow winds and spans upto 500m or

Figure 4.19: As in Figure 4.17 but for radar reflectivity (shaded) with contours of vertical velocity at an interval of 0.4 m s^{-1} .

that the increased friction would apply larger agradient force to the inflow rings of boundary layer, and would cause them to move further inwards before they rise out of the boundary layer and ascend into the eyewall updraft (Figure 4.19). Figure 4.18 show that a deeper inflow layer was present in Runs 2 and 3 compared to Run 4. It is also revealed that Run 3 exhibits a much stronger return outflow above the inflow layer, with deeper and stronger upper level outflow. From Figures 4.18 and 4.19, it is of note that the stronger TC results in a narrower, more upright eyewall (more vertical updraft) with strong vertical motion above the aforementioned inflow layer. Lastly,

1km above the surface

using Figures 4.16 to 4.19, it can be concluded that the different flux parametrisations result in a somewhat structurally similar TC, albeit with some key differences, where the largest differences are limited to the inner core of the TC in respective model simulations.

4.4 Summary

Over the last two decades significant progress have been made towards improving TC track prediction, however, despite considerable efforts towards the improvement of TC intensity forecast, the predictions have plateaued. The previous research using a combination of measurements, modelling and theory have argued that TC intensity *inter alia*, is very sensitive to the fluxes of momentum and enthalpy exchange at the air-sea interface. As described in preceding chapters, traditionally the numerical weather prediction (NWP) models approximate the flux exchange using bulk parameterisations.

In the present chapter, the atmosphere model was applied to study the effects of these aforementioned flux parameterisations on the intensity and track prediction of a TC. However, to allay the deficiencies of the momentum flux parameterisations used in the NWP models, the WRF model was coupled with a spectral wave model, thereby accounting for the effects of sea state on the momentum flux exchange. It was found out that the different flux parameterisations don't have a significant effects on TC track, they affect the MSLP and V_{max} , *i.e.* MSLP and V_{max} varied for different flux parameterisations. From the model simulations performed here, it was found that the coupling of wave model has noticeable effects on the model results. It was also observed that none of the simulations exhibit unequivocal improvements in all areas, rather the different flux parameterisations improve different properties of model simulations.

From the coupled atmosphere-wave model simulations, it was found that wave model dependent surface roughness results in increased drag coefficient, which affects atmosphere in two very different ways, where the increased surface drag would act to reduce wind speeds in high wind areas, while at the same time, this increased drag results in higher coefficient of heat and moisture, which causes the TC to intensify.

However, such an intensification process could be modified by other processes such as sea spray, which will be discussed in Chapter 5. Besides affecting the general characteristics of the TC, different bulk parameterisations also affected the vertical structure of the TC. As asserted earlier that coupling wave model results in higher heat and moisture fluxes at the surface, which results in an intense storm with a well developed secondary circulation *i.e.* relatively stronger inflow from boundary layer to mid troposphere and stronger outflow in the upper troposphere. It was also seen that the coupling wave model although improves the model results when compared to buoy measurements, causes TC to over intensify.

Lastly, it must be pointed out that this chapter in no way tries to solve the on-going problem of the forecast skill of TC intensity, but merely tries to point that there are other processes besides parameters such as surface roughness that can affect the model results due to their rather overlooked dependence on the said variables (or parameters) even when coupled modelling approach is employed. Furthermore, it behoves us to mention that, the numerical experiments used in present chapter only consider some aspects of the physical processes at the air-sea interface, while there are other physical processes (*e.g.* sea spray) and parameters (*e.g.* sea surface temperature, ocean currents) that can also affect the simulated TCs.

Chapter 5

The Role of Sea Spray on Modelled Tropical Cyclone

5.1 Preamble

The strong winds prevalent during severe storms like TCs cause intense breaking of ocean surface waves, resulting in generation of sea spray. This sea spray is believed to play a role in modulating air-sea flux exchanges as speculated by Ling and Kao (1976). In the last two decades, there have been numerous studies (see Section 2.5.4) investigating the role of sea spray within the context of atmosphere-ocean interaction. These studies have focussed mainly on the effects of sea spray on momentum and heat exchange between atmosphere and ocean. Despite considerable effort, there hasn't been an unequivocal answer on the role of sea spray due to *inter alia*, the uncertainty in modelling sea spray and the lack of measurements during hurricanes. The lack of measurements at high wind speeds is due to the extreme difficulty in carrying out direct measurement of air-sea thermodynamic fluxes in such conditions, thus limiting the measurements of sensible heat and moisture flux to wind speeds of upto 20ms^{-1} . To elaborate further on the aforementioned assertion, there are three sets of observational datasets which have been used to obtain the surface flux parameterizations where the first dataset was collected under the HEXOS campaign (DeCosmo, 1991; Smith et al., 1992), the second dataset was collected under the FASTEX campaign (Fairall et al., 2003; Persson et al., 2005), and the last dataset was collected under the CBLAST

campaign (Black et al., 2007). In the former two campaigns, the dataset collected was limited to the wind speeds of 20ms^{-1} , whereas in the CBLAST campaign, there were some observations made at wind speeds $20\text{-}30\text{ms}^{-1}$, however, these measurements were made using GPS dropsondes, which have a lower reliability in 6-8m near the sea surface as compared to the dataset collected (in the former two campaigns) using eddy covariance method within the 10-20m height from the sea surface. An ensemble of the different measurement datasets can be seen in Drennan et al. (2007, p. 1104), where it is clearly noticeable that the measurements are limited to wind speeds of 20ms^{-1} . Although, Drennan et al. (2007, p. 1112) shows that the CBLAST campaign extended the measurements upto the wind speed of 30ms^{-1} , the increased scatter and reduced reliability of dropsonde measurements limits the applicability of these CBLAST measurements. However, despite such limited observations, Andreas and DeCosmo (1999) showed that the physically based parametrisations for sensible and latent heat flux are unable to reproduce the measured fluxes. Figure 5.1 shows a comparison of measured and parametrised sensible and latent heat flux, with a noticeable bias towards the measured fluxes, with higher bias at higher wind speeds. They postulated that this bias between the modelled and measured fluxes is due to the presence of sea spray.

Keprt et al. (1999); Andreas and Emanuel (2001) using numerical models have shown that sea spray, when included in the model simulations can have dramatic effects on air-sea flux exchange thereby affecting storm intensity. This is because when sea spray is present, it provides an additional mechanism for exchange of heat fluxes *i.e.* sensible and latent heat, while also affecting the momentum exchange between the atmosphere and ocean.

When sea spray droplets are lofted in air, they increase the effective areal contact between atmosphere and ocean. Furthermore, in conditions prevalent during hurricanes, the ocean surface is warmer than air, hence this enhanced areal contact between atmosphere and ocean, results in additional heat flux exchange from ocean to atmosphere. Besides the heat flux (convective) transfer from ocean to atmosphere, these sea spray droplets also extract latent heat from atmosphere so as to evaporate, thus causing some cooling in the near surface atmospheric layer. Anthes (1982) argued

Figure 5.1: A scatter plot comparison of measured and parametrised sensible heat flux (left) and latent heat flux (right) using HEXOS (DeCosmo, 1991) dataset. The markers are coloured by mean wind speed. The parametrised fluxes are computed using COARE2.6 (Liu et al., 1979; Fairall et al., 1996).

that during a hurricane, this strong cooling caused by the evaporating spray droplets on one hand would make the air (in boundary layer) more stable thus reducing the storm intensity, while on other hand would enhance sensible heat transfer from ocean to atmosphere, resulting in intensification of the storm. Moreover, the evolution and impact of these sea spray droplets depends on the rate at which spray droplets are being generated (*i.e.* mass/volume flux). Due to lack of complete knowledge of spray generation process over the wide range of droplet sizes, the flux (mass/volume) of spray droplets is usually represented with spray source generation function (SSGF), see Figure 2.9 for an exemplary SSGF. As described in (Andreas, 1998; Wu et al., 2015; Richter and Veron, 2016), there exist a number of sea spray generation functions based on the field observations, albeit limited by both the number of reliable observation as well as the range of wind speed, with no observations in hurricane intensity wind speeds. Due to the aforementioned difficulty in conducting field measurements, studies like Fairall et al. (2009) carried out measurements in the laboratory. Even with the relative ease in conducting investigations in laboratory, the results for production rate obtained from these studies showed wide divergence thus underlining the lack of understanding and ways to fully characterise the spray generation at the wide range of droplet sizes.

5.2 Sea Spray Parametrisation

5.2.1 Thermal effects of sea spray

From the above description, the sea spray droplets when lofted into air from a relatively warmer ocean surface compared to air, they can exchange both heat and moisture. Also, as the sea spray are saline droplets, therefore when evaporated, would either result in saline crystals or as suggested by Andreas (1995), attain temperature and radius, where they are in quasi-equilibrium state with their environment.

To model such a dynamic process, a microphysical model similar to the one suggested by Pruppacher and Klett (2010) is needed. However, due to the complexity in implementing such a model within a large scale atmosphere model and the large computational overhead, Andreas (1989, 1996, 2005); Kepert (1996) devised approximations to compute the equilibrium temperature and radius of the evaporating sea spray droplets. With the effects of sea spray evaporation included, the total sensible and latent heat fluxes can be written as

$$\begin{aligned} H_{L,T} &= H_{L,I} + \alpha \overline{Q_L} \\ H_{S,T} &= H_{S,I} + \beta \overline{Q_S} - (\alpha - \gamma) \overline{Q_L} \end{aligned} \quad (5.1)$$

Here, $\overline{Q_S}$, $\overline{Q_L}$ in eq. (5.1) are the spray mediated "nominal" sensible and latent heat flux obtained from the microphysical calculation devised by Andreas (2005), and $H_{L,I}$, $H_{S,I}$ are the interfacial latent and sensible heat fluxes representing the interaction at air-sea interface. The terms α , β , γ in eq. (5.1) are small non-negative constants obtained by statistical fitting "nominal" fluxes to the field observations. The "nominal" fluxes in eq. (5.1) are obtained by integrating sensible and latent flux contribution for all the spray droplets with radius r :

$$\overline{Q_S} = \int_{r_1}^{r_2} Q_S(r) dr, \quad \overline{Q_L} = \int_{r_1}^{r_2} Q_L(r) dr \quad (5.2)$$

where r varies from minimum radius r_1 to maximum radius r_2 of the spray droplets considered in the microphysical computation. The terms Q_L , Q_S in the eq. (5.2) are calculated for each spray droplet with radius r , where spray mediated latent heat flux

Q_L is given as

$$Q_L(r) = \begin{cases} -\rho_s L_v \left(1 - \left(\frac{r(\tau_f)}{r_0}\right)^3\right) \left(\frac{4\pi r^3}{3} \frac{dF}{dr}\right), & \text{if } \tau_f \leq \tau_r \\ -\rho_s L_v \left(1 - \left(\frac{r_{eq}}{r_0}\right)^3\right) \left(\frac{4\pi r^3}{3} \frac{dF}{dr}\right), & \text{otherwise} \end{cases} \quad (5.3)$$

while the sensible heat flux Q_s as

$$Q_s(r) = \rho_s c_{ps} (T_s - T_{eq}) \left[1 - \exp\left(\frac{-\tau_f}{\tau_T}\right)\right] \left(\frac{4\pi r^3}{3} \frac{dF}{dr}\right) \quad (5.4)$$

In eqs. (5.3) and (5.4), the quantities τ_f, τ_T, τ_r are the three different time scales associated with the different stages over the lifetime of a sea spray droplet. Following Andreas (1990), τ_f is the time spray droplet remain aloft in air, while τ_T is the time taken by droplet to reach e^{-1} of its equilibrium temperature and lastly, τ_r is the time taken by droplet to evaporate to e^{-1} of its equilibrium radius. T_s, T_{eq} are the temperature of sea surface and the equilibrium temperature of evaporating droplets, ρ_s is density of sea water, c_{ps} is specific heat of sea water at constant pressure, L_v is the latent heat of vaporisation, and r_{eq} is the radius of droplets at when it reaches equilibrium with its environment. The term dF/dr in eqs. (5.3) and (5.4) represents SSGF, which is the rate at which droplet with initial radius r_0 are generated at the sea surface, and $(4\pi r^3/3)dF/dr$ is the total volume flux of spray generated at the sea surface.

From Andreas (1989, 1990, 2005), the instantaneous temperature and radius of spray droplet in eqs. (5.3) and (5.4) are obtained using an exponential decay model following Pruppacher and Klett (2010), where the temperature evolution of spray droplet is approximated as

$$\frac{T(t) - T_{eq}}{T_s - T_{eq}} = \exp(-t/\tau_T) \quad (5.5)$$

and the radius evolution as

$$\frac{r(t) - r_{eq}}{r_0 - r_{eq}} = \exp(-t/\tau_r) \quad (5.6)$$

These microphysical quantities, depend not only on the initial droplet radius and air-sea temperature difference, but also on the relative humidity near the sea surface, water salinity as well as the atmospheric pressure at sea level. An example output from the sea spray model used in present study is shown in Figure 5.2. From Figure 5.2, it is seen that the spray droplet cools much faster as compared to the change in droplet size due to evaporation, where the former took $< 1 \text{ sec}$ to loose (or drop) 3°C , while the latter took ≈ 50 seconds for any noticeable change in droplet size and about 1000 seconds for droplet radius to reduce by $40 \mu\text{m}$. This noticeable decoupling between the temperature and radius evolution is used as a justification for the usage of approximate formulas for modelling sea spray droplets. For the sake of brevity, readers are referred to Andreas (1989); Andreas (1992); Andreas (2005) for details regarding the computation of these terms.

Figure 5.2: The temperature and radius evolution of spray droplet with radius $100 \mu\text{m}$, air temperature of 18°C , water temperature of 20°C and relative humidity of 90 %. The solid line represent the output of (approximate) microphysical model used in present study and dashed line represent the (full microphysical) model output from Andreas (2005).

5.2.2 Momentum effects of sea spray

When sea spray is present in the DEL, total surface stress τ_t near the atmosphere-ocean interface can be partitioned into ocean wave induced surface stress τ_w , surface stress supported by sea spray droplets τ_{sp} and the viscous stress τ_v at the sea surface. The total stress τ_t in eq. (2.5) can therefore be written as:

$$\tau_t = \tau_w + \tau_v + \tau_{sp} \quad (5.7)$$

To obtain the sea spray induced stress τ_{sp} , we follow the approach similar to the one applied for computing spray induced thermal fluxes, where we compute the contribution of individual droplets and then integrate them over the whole range of droplet radius considered, thus obtaining the total spray induced stress. Following Andreas and Emanuel (2001), the spray induced stress can be written as:

$$\tau_{sp} = \frac{4\pi}{3} \rho_s \int_{r_1}^{r_2} u_{sp}(r) r_0^3 \frac{dF}{dr} dr \quad (5.8)$$

where u_{sp} is the horizontal velocity of the spray droplet before it falls back into the ocean.

5.2.3 Sea spray generation function

The sea spray droplets present in the near surface layer can be classified in two broad categories, film and jet droplets that generate by means of bursting of bubbles formed due to the air trapped by the breaking of waves, and spume droplets that generate by means of tearing off of the wave crest. As mentioned in Andreas (1998), spume droplets are generated at higher wind speed ($> 9ms^{-1}$) and are usually of radius greater than $30\mu m$ and can be as large as $500\mu m$, whereas the film and jet droplets have radius less than $30\mu m$. Furthermore, when the spray droplets are generated, they are either at the ocean wave propagation speed or at rest. These droplets when lofted in air, gets accelerated by the wind speed, thereby affecting the momentum exchange at the air-sea interface. Few possible explanation on the effects of sea spray on the momentum exchange have been provided in literature, where Andreas (2004)

argued that the spray droplets when return to ocean surface, will result in sheltering of short waves, responsible for carrying much of the wave stress, while Bye and Jenkins (2006); Kepert et al. (1999) suggested that the presence of spray droplets will cause suppression of turbulence, due to spray droplet mass loading, and will increase the stability of the boundary layer.

From eqs. (5.3), (5.4) and (5.8), it is evident that the volume flux (or SSGF) of spray generated has a direct influence on the spray mediated thermal and momentum fluxes. Mueller and Veron (2014); Andreas (2004) showed that the spray effects on momentum and thermal fluxes increase at higher wind speeds, where this increase in spray effects at higher wind speeds is attributed to the increased number of spume droplets generated at higher wind speed. However, most of SSGF in the literature are only valid at wind speeds below 20ms^{-1} and droplet radius below $30\mu\text{m}$. Following Fairall et al. (1990); Andreas (1992); Fairall et al. (1994), a typical spectral distribution of sea spray droplets $S_n(r)$ can be written as

$$S_n(r) = W_f(U)f_n(r) \quad (5.9)$$

where, U is wind speed, usually taken at 10m height, W_f is the fraction of sea surface covered by whitecaps, and f_n is the distribution of droplet spectrum. Whitecaps at the sea surface are produced when breaking waves trap air in form of bubbles, thus resulting in white foamy appearance on the sea surface. Therefore, these whitecaps can be taken as representation of wave energy dissipation. Recent studies Anguelova and Hwang (2016); Scanlon et al. (2016) have described two different approaches to obtain whitecap fraction, where the former follows the approach described by Phillips (1985), while latter follows the method suggested in Kraan et al. (1996), hereafter referred as "Kraan96". For the purpose of this study, we will limit our discussion to the latter approach. Following Janssen (2012); Breivik et al. (2015), the turbulent kinetic energy (TKE) flux ϕ_{oc} from breaking waves (whitecap dissipation) to ocean is related to the dissipation source function S_{ds} of the spectral wave model and can be given as:

$$\phi_{oc} = \rho_w g \int_0^{2\pi} \int_0^\infty S_{ds} d\omega d\theta \quad (5.10)$$

where θ is wave direction. Furthermore, following Kraan et al. (1996), it can be further assumed that the TKE flux ϕ_{oc} in eq. (5.10) is linearly proportional to whitecap fraction W_f as

$$\phi_{oc} = \epsilon \rho_w g W_f \omega_p E \quad (5.11)$$

Here, ω_p is angular frequency corresponding to the peak wave energy density, E is the total wave energy density, ϵ is the average fraction of wave energy dissipated per whitecap event, and is set to 0.01. In applying eqs. (5.10) and (5.11), it is implicitly assumed that in the equilibrium range (Phillips, 1985) the energy source S_{in} and sink terms S_{ds} in the wave action density equation (Komen et al., 1984) are in balance (Hanson and Phillips, 1999):

$$S_{in} + S_{nl} - S_{ds} = 0 \quad (5.12)$$

Here, S_{nl} is the non-linear wave-wave interaction which represents the redistribution of wave energy from large scales to smaller scales. Hence, in the integrated action balance equation, wind input S_{in} and energy dissipation S_{ds} are in balance, thus permitting the usage of dissipation source function for estimating the spray mediated stress term given in eq. (5.12).

5.3 Wave dependent spray parametrisation

Following the arguments of present approach in section 5.2, the sea spray fluxes are modelled using the wave energy dissipation. Here, we compare the whitecap fraction obtained from wave energy spectrum with the model used in present study to that of the model derived in recent study of Anguelova and Hwang (2016). We also obtain the constant terms needed to obtain "effective" sea spray fluxes from the "nominal" spray fluxes calculated using the microphysical model described in the section 5.2.

5.3.1 Whitecap fraction

As described earlier (section 5.2), in the present study we estimate SSGF using eq. (5.9), where the whitecap fraction is obtained via eqs. (5.10) and (5.11). Anguelova

Figure 5.3: Wind speed dependence of whitecap fraction (on log scale) obtained from (a) Kraan96 (Kraan et al., 1996), AH2016 (Anguelova and Hwang, 2016) and MOM80 (Monahan and Muircheartaigh, 1980), (b) As in (a), but data are binned in wind speed bins of $1 ms^{-1}$ using data from buoy 46001

and Hwang (2016) developed an analytical model for estimating whitecap fraction, and applied their model to wave energy spectrum observations from NDBC buoy 46001 moored in Gulf of Alaska at ($56.3^\circ N$, $147.9^\circ W$). Additionally, they compared their results with the photographic measurements of whitecap fraction obtained in Gulf Alaska. It was shown that the whitecap fractions obtained from their model were comparable to the photographic observation. Furthermore, they also compared their results with the whitecap fraction model from Monahan and Muircheartaigh (1980), hereafter referred as MOM80.

Figure 5.3 shows the comparison of whitecap fraction obtained using Monahan and Muircheartaigh (1980); Anguelova and Hwang (2016) and "Kraan96", where the results obtained from eqs. (5.10) and (5.11) are comparable to Anguelova and

Hwang (2016). For details on Anguelova and Hwang (2016) model hereafter referred as AH2016, processing of wave energy spectrum from buoy measurements and their validation procedure, readers are referred to Anguelova and Hwang (2016). Figure 5.3 (b) shows the same data (Figure 5.3 (a)) binned by wind speed. From Figure 5.3, it can be inferred that the whitecap fraction obtained from eqs. (5.10) and (5.11) are slightly higher as compared to AH2016 model. It should however be noted that both methods show similar wind speed dependence where at higher wind speed, both methods give lower whitecap fraction compared to MOM80 model.

5.3.2 Estimation of sea spray fluxes

One implication of using wave state dependent SSGF is the need to obtain the coefficients α , β , γ given in eq. (5.1). As described in section 5.2.1, these coefficients are obtained from statistical fit of total fluxes (*i.e.* spray mediated and interfacial) to the fluxes obtained from field observations. From eqs. (5.3), (5.4) and (5.8), it is also known that the sea spray fluxes depend on the volume flux of sea spray ejected in the DEL. For the purpose of this study, a procedure similar to the one described in Andreas and DeCosmo (1999) was followed, and the HEXOS dataset (DeCosmo, 1991) was used. Andreas et al. (2008) utilised the same HEXOS dataset together with FASTEX dataset (Persson et al., 2005) to obtain the constant terms for the spray flux algorithm.

Using microphysical spray model with the COARE2.6 bulk flux algorithm we obtained $\alpha = 7.7036$, $\beta = 0.0$ and $\gamma = 8.3202$. We also calculated the correlation coefficients of modelled fluxes to the observed fluxes, where the correlation coefficient for sensible heat was 0.93 and for latent heat was 0.89. Figure 5.4 shows the ratio of measured fluxes *i.e.* latent and sensible heat flux to the corresponding interfacial and total fluxes *i.e.* interfacial fluxes with spray contribution. It is evident that the addition of sea spray fluxes is able to account for the bias noticed between the measured and interfacial fluxes, however, it is necessary to keep in mind that this improvement is due to both the addition of spray fluxes and the statistical fit used to achieve improved correlation. Therefore, it is advised that extreme precaution must be exercised when using these results. Using eq. (5.1), total enthalpy flux above the DEL can be written

Figure 5.4: The ratios of HEXOS measurements of (a) sensible, (b) latent heat flux to the corresponding interfacial fluxes (●) and total fluxes *i.e.* interfacial fluxes with spray contribution (▲). The interfacial fluxes are computed using COARE2.6 algorithm (Fairall et al., 1996) and the HEXOS data is obtained from DeCosmo (1991). The abscissa is the neutral stability wind speed at 10m height above surface.

as:

$$H_{S,T} + H_{L,T} = H_{L,I} + H_{S,I} + \beta \overline{Q_S} + \gamma \overline{Q_L} \quad (5.13)$$

When viewed in context of enthalpy flux, it can be said that only β, γ have an

Table 5.1: Summary of numerical model experiments with different coupling configurations

Experiment No.	Wave Coupling	Spray Heat Fluxes	Spray Stress
1	No	No	No
2	Yes	No	No
3	Yes	Yes	No
4	Yes	Yes	Yes

effect on heat flux transfer from ocean to atmosphere. The values for the constants obtained in present study imply that only spray mediated latent flux have a role on the heat flux transfer. This is in contrast to the results obtained in Andreas et al. (2008) and Andreas and DeCosmo (1999) where they found β to be positive non-zero value. Also, it contradicts with the conclusion that spray sensible heat flux is the primary route by which spray affects the storm energy as stated in Andreas and Emanuel (2001). However, we want to stress that further investigation with more observation data is needed to support our earlier statement.

5.4 Numerical Experiments

To evaluate the effects of ocean waves dependent sea spray on a TC, two additional numerical experiments (Table 5.1) were conducted. The Expt. 1 and 2 have already been summarised in Chapter 4, therefore, we will limit the discussion to the additional experiments conducted for investigating the effects of spray mediated fluxes. In Expt. 3 and 4, the sea spray fluxes were added to the coupled model runs, where in Expt. 3 only the spray mediated heat fluxes were applied, while in Expt. 4, both the spray mediated heat and momentum fluxes were applied. Similar to Chapter 4, the model experiments summarised in Table 5.1 do include data assimilation. To include the effects of sea spray fluxes on the atmosphere model, the surface layer scheme in atmosphere model was modified. Also, the wave energy dissipation and whitecap fraction from the wave model output was computed using the modelled wave energy spectra over the whole coupled model domain. The sea spray fluxes thus obtained were remapped within the coupler following the procedure described in Chapter 3.

5.5 Results and Discussion

Figure 5.5: Comparison of (a) modelled tracks of hurricane Arthur (2014) (b) Time series of minimum sea level pressure (MSLP), and (c) Time series of maximum wind speed (V_{max}) for observed and four model experiments

5.5.1 Storm track and intensity

Storm track

A comparison of simulated tracks of hurricane Arthur (2014), obtained using the minimum sea level pressure, with the location of storm centre from best track and model results is presented in Figure 5.5 (a). The simulated storm tracks are generally consistent with the best track, where the modelled storms first track southwestward and thereafter turning and move northward before making landfall. We can notice that all the modelled storms are westward of the best track, however when coupled with wave model the storm tracks improved. We can also see that the storm in uncoupled model (Expt. 1) has a higher translation speed compared to coupled atmosphere-wave model (Expts. 2-4) and best track data. Additionally, Figure 5.6 illustrates anomalies due to sea spray mediated heat and momentum flux in TC track (Figure 5.6 (a)) and intensity (Figure 5.6 (b)), where the results are only presented for Expts. 3 and 4.

Although, sea spray coupling (Expts. 3-4) do not have an appreciable effects on the model track, it does affect the translation speed of storm, where in Expt. 3 (*i.e.* sea

Figure 5.6: As in Figure 4.3, but for the numerical experiments described in Table 5.1. The error in TC centre location and intensity are only presented for Expt. 3 and Expt. 4, whereas for Expt. 1 and 2 are shown in Figure 4.3.

spray coupling with only heat fluxes) storm moves faster compared to Expt. 4 (*i.e.* sea spray coupling with heat and momentum fluxes). Also both the storms moved faster as compared to Expt. 2.

Minimum sea level pressure (MSLP)

The time series of minimum sea level pressure (MSLP) are compared with the MSLP from best track data in Figure 5.5 (b). Comparing the results of different numerical experiments (Expts. 1-4) indicate that uncoupled model underestimates the storm intensity while the coupled model overestimates the storm intensity. The effect of both sea spray heat and momentum flux (Expt. 4) have little effect compared to atmosphere-

Figure 5.7: Comparison of coupled model results with NDBC buoys (a) 41002 and (b) 44014 for mean wind speed U_{10} [ms^{-1}] at 10m elevation

wave coupled model (Expt. 2). If only sea spray heat fluxes are considered (Expt. 3), the storm intensity is closer to the best track data. Including the wave and sea spray coupling (Expts. 2-4), the storm intensifies earlier compared to uncoupled model (Expt. 1).

Maximum wind speed (V_{max})

The temporal development of maximum wind speed at 10 meters for the 4 experiments (Expts. 1-4) is shown in Figure 5.5 (c). The effects of different model couplings on maximum velocity V_{max} are similar to that on MSLP. We can see that hurricane under intensifies by upto $\approx 10\text{ms}^{-1}$ in uncoupled (Expt. 1) model compared to best track data. Also, we can note that the storm intensity improves when the atmosphere is coupled with waves (Expts. 2-4). However, in contrast to section 5.5.1, from

Figure 5.5 it is evident that the V_{max} are better modelled in Expt. 2 and Expt. 4.

Figure 5.8: As in Figure 5.7, but for wind direction

5.5.2 Model Validation

Wind observations

To further investigate the effects of ocean waves and sea spray, similar to chapter 4 we compare the computed wind speeds with the winds measured at NDBC buoys. Both coupled and uncoupled model in the present study over predict the intensity of the storm at 44014 while performing well at buoy 41002 (Figure 5.7 (a)), however we do note that the wave coupling on atmosphere results in improved timing of the storm at buoy 44014 (Figure 5.7 (b)). We also see that the coupling sea spray (Expts. 3-4) results in lower V_{max} at the buoy location, though when spray mediated momentum flux is applied together with spray mediated heat fluxes (Expt. 4), it improves both the timing as well as the buoy location. Additionally, Figure 5.8 compares the wind direction

at buoy 41002 and 44014 obtained from the 4 numerical experiments. Furthermore, the computed statistics for both wind speed and wave parameters (i.e. wave height H_s and wave period T_{02}) are given in Table 5.2. The coupling of ocean surface wave reduces (increase) the error in wind speed at buoy 44014 (41002), however when sea spray fluxes are applied, this reduction in error diminishes. Moreover, when comparing both the RMSE and correlation between wave measurements and data obtained from various numerical experiments, it can be said that the coupling of wave model improves the model results. It can also be construed from Table 5.2 that it is necessary to account for both spray mediated heat and momentum flux, as when only spray mediated heat fluxes are accounted, there is a noticeable reduction in correlation coefficient with increase in RMSE.

Table 5.2: Root mean square error (RMSE) and correlation coefficient (R) for mean wind speed U_{10} at 10m elevation, significant wave height H_s and wave period T_{02} between model experiments and buoys 41002 and 44014

Runs	Buoy 41002					
	$U_{10}[ms^{-1}]$		$H_s[m]$		$T_{02}[sec]$	
	RMSE	R	RMSE	R	RMSE	R
Expt. 1	1.17	0.88	0.39	0.93	0.37	0.87
Expt. 2	1.35	0.88	0.39	0.92	0.37	0.87
Expt. 3	1.21	0.88	0.43	0.88	0.37	0.84
Expt. 4	1.13	0.90	0.39	0.91	0.34	0.88
	Buoy 44014					
	$U_{10}[ms^{-1}]$		$H_s[m]$		$T_{02}[sec]$	
	RMSE	R	RMSE	R	RMSE	R
Expt. 1	3.87	0.80	0.99	0.83	0.68	0.87
Expt. 2	2.98	0.92	0.52	0.96	0.50	0.95
Expt. 3	3.61	0.79	0.72	0.89	0.63	0.91
Expt. 4	3.62	0.81	0.64	0.94	0.51	0.94

Wave observations

During a hurricane, the waves are not only affected by the winds but also by the hurricane intensity and translation speed among other factors. Figures 5.9 and 5.10 show the comparison of significant wave height H_s and mean wave period T_{02} for the four different model experiments with the buoy measurements.

The significant wave height H_s and wave period T_{02} were rather well estimated when atmosphere model was coupled with waves (Expts. 2-4) compared to uncoupled

model (Expt. 1). Although the size of storm induced wave fields in coupled model (Expt. 2) indicates that the storm size is bigger compared to uncoupled model and buoy measurements. Also, when sea spray effects are included in the coupled model,

Figure 5.9: As in Figure 5.7, but for significant wave height H_s [m]

the modelled peak H_s are similar to that of measurements, hurricane passes the buoy locations earlier than observed. We attribute the bias in timing of the storm passage to the translation speed of the storm, where higher translation speed can result in increased effective fetch, thereby giving higher wave heights.

We also compared the collocated significant wave heights from model experiments to the satellite altimeter derived wave heights (Figure 5.11). This was done by computing the closest model data point (both in space and time) to each satellite observation point. This allows us to evaluate the spatial and temporal variation of wave field due to different processes investigated here. The computed statistics of the modelled wave heights to the altimeter data are given in Table 5.3. When the

Figure 5.10: As in Figure 5.7, but for wave period T_{02} [sec]

wave effects were included in the wave model, a higher correlation and lower scatter compared to altimeter data was observed. When sea spray was included in the coupled atmosphere-wave model runs, we can see that including the spray mediated momentum flux improves the model results.

It is noteworthy there are only minor differences between the output of uncoupled and coupled model experiments at lower wave heights, however only coupled models were able to capture the higher wave heights. It is also worth mentioning that apart from modelling uncertainties, the bias in model results can arise from the differences in temporal and spatial resolution of wave model and satellite altimeter. Furthermore, for the comparison presented in Figure 5.11, no spatial (or temporal) smoothing of altimeter data was carried out. This is due to the use of unstructured grid in the present study for the wave model setup.

Figure 5.11: Scatter plot of modelled versus observed H_s for (a) Expt. 1, (b) Expt. 2, (c) Expt. 3, and (d) Expt. 4 during June 30 to July 5 within the coupled model domain. The colorbar indicates the number of occurrence in each 0.1m bin on a logarithmic scale. The solid black represents the ideal fit of model and measurements while the dashed red line represents the linear fit of the data.

Figure 5.12: Computed whitecap fraction from Expt. 4 at 0000 UTC July 3, 2014 (a) Distribution of whitecap fraction with contours of mean wind speed U_{10} , (b) comparison of whitecap fraction (on log scale) versus mean wind speed U_{10}

Table 5.3: Statistical comparison of altimeter and wave model derived significant wave height H_s ; normalised bias (NBIAS), root mean square error (RMSE), Pearson correlation coefficient (R) and scatter index (SI)

Runs	NBIAS[m]	RMSE[m]	R	SI
Expt. 1	0.084	0.44	0.78	0.42
Expt. 2	0.093	0.43	0.82	0.40
Expt. 3	0.095	0.44	0.81	0.42
Expt. 4	0.087	0.43	0.82	0.41

5.5.3 Surface fluxes

Figure 5.12 shows the distribution of wave model dependent whitecap fraction computed from Expt. 4 at 0000 UTC July 3, 2014, where Figure 5.12 (a) presents the spatial distribution of whitecap fraction with wind speed contours while Figure 5.12 (b) shows a comparison of whitecap fraction obtained from wave energy spectrum to the widely used MOM80 formula. It should be noted that the results given in Figures 5.11 to 5.13 are presented on a storm centric grid with 2km spacing in radial and 1° spacing in azimuthal direction. Here, the storm relative grid were created using storm centres which corresponds to the location of minimum sea level pressure. The whitecap fraction (Figure 5.12) and surface heat fluxes (Figures 5.11 to 5.13) obtained from wave and atmosphere model were interpolated onto the aforementioned storm relative grid. Furthermore, only data points that are within 200km from the storm centre and over the ocean are presented. From Figures 5.3 and 5.12 (b), for wind speeds between $10 - 20\text{ms}^{-1}$, whitecap fraction obtained from MOM80 and eqs. (5.9) to (5.11) show similar wind speed dependence, however, when extended to wind speeds present during hurricanes, whitecap fraction obtained from MOM80 are substantially higher with whitecap fraction of 1.0 at 40ms^{-1} . This seems to show that at wind speed greater than 40ms^{-1} whole sea surface will be covered in whitecaps, whereas only 20% of sea surface is covered when whitecap fraction are computed from wave energy dissipation. Until now, we have only discussed the wind speed dependence of whitecap fraction, when looking at the spatial distribution of whitecap fraction in relation to the wind speed (see Figure 5.12), it can be noted that the peak of whitecap fraction is in the rear left quadrant of the hurricane translation direction (see black arrow in Figure 5.13 (d)), whereas the peak of wind intensity is in the front

Figure 5.13: Plan views of surface sensible heat flux for (a) Expt. 1, (b) Expt. 2, (c) Expt. 3 and (d) Expt. 4 at 0000 UTC July 3, 2014. The black arrow indicates the hurricane translation direction over three hour interval.

right quadrant of the hurricane. This shows that within the coupled model used in present study, not only the volume of sea spray droplets generated was altered, but also the spatial spread of sea spray volume flux was modified. This is noteworthy, as when sea spray production is parametrised using MOM80 (where whitecap fraction $W_f = 3.8e^{-6}U_{10}^{3.41}$), the peak of whitecap fraction will be collocated with the peak of wind speed. It is important to keep in mind that the results presented in Figure 5.12 (b) merely highlight the fact that in most studies MOM80 model is applied beyond the range of its validity, as is the case in Figure 5.12 (b).

The effects of including spray mediated heat fluxes as well as ocean surface waves on the enthalpy fluxes (*i.e.* sensible and latent heat flux) are shown in Figures 5.13 to 5.18. Here, we first compare the total sensible heat flux (Figure 5.13), then the total latent heat flux (Figure 5.14) and eventually their azimuthal averaged radial variation in Figure 5.15. The heat fluxes are presented at 0000 UTC July 3, 2014, because the storm centres were collocated (see Figure 5.5 (a)). It is evident from the comparison

Figure 5.14: As in Figure 5.13, but for surface latent heat flux

of sensible and latent heat flux obtained from uncoupled atmosphere model (Expt. 1) and ocean wave coupled atmosphere model (Expt. 2), that coupling ocean waves with atmosphere results in substantial increase in sensible and latent heat fluxes. Using Figures 5.13 and 5.14, it can be argued that the increase in heat fluxes is largely due to wave induced surface roughness, rather than any air-sea temperature difference that might arise due to the usage of fixed sea surface temperature (SST) field. These results are inline with the arguments given in Janssen et al. (2001), where it was suggested that the increased surface roughness will enhance the surface heat fluxes causing vortex stretching, thus intensifying the storm.

By comparing Figure 5.13 (b) and Figure 5.13 (c), it can be noticed that when only spray mediated heat fluxes (Expt. 3) are added, there is a reduction in sensible heat flux as well as broadening of storm core compared to Expt. 2. Furthermore, when both spray mediated heat and momentum flux are added (Expt. 4), it results in higher sensible heat flux (see Figure 5.13 (d)) compared to Expt. 3. Besides the differences in sensible heat flux, there also exist differences in location of maximum sensible heat flux with respect to storm centre. Contrary to assumption that applying spray mediated

Figure 5.15: Radial distribution of azimuthal averaged total (a) latent heat flux (b) sensible heat flux at 0000 UTC July 3, 2014

heat flux will intensify the hurricane, these results show that the interaction between sea spray and hurricane are rather more intricate, where both the thermodynamic and dynamic processes play different roles. For instance, coupling waves with atmosphere model increases the surface roughness, which results in intensification of hurricane. Increased surface roughness may however also decelerates the hurricane as seen in Figure 5.5, causing it to stay on warmer ocean for longer duration.

The radial distribution of azimuthally averaged total latent and sensible heat fluxes for the 4 model experiments are presented in Figure 5.15. It is clearly noticeable that in all the experiments, maximum value of heat fluxes (*i.e.* sensible and latent heat flux) are in high wind region (*i.e.* radius = 20 to 75 km). Also, maximum value of latent heat flux in Expt. 2-4 is twice that of Expt. 1 whereas maximum value of sensible heat flux in Expt. 2 and Expt. 4 is thrice, while in Expt. 3 is 2.5 times that of Expt. 1. Besides the effects of coupling wave model with atmosphere model (Expt. 2), the effects of sea spray on sensible and latent heat fluxes can also be noticed. In the case of sea spray heat fluxes (Expt. 3), there is a noticeable reduction in maximum value of sensible and latent heat fluxes compared to Expt. 2. However when both spray mediated momentum and heat fluxes are considered (Expt. 4), there is a reduction in maximum latent heat flux while an increase in the maximum sensible heat flux compared to Expt. 2. It should also be pointed out that these difference in heat fluxes (between Expts. 2-4) are only in high wind region, with negligible effects at higher radius. In addition to affecting the value of heat fluxes, sea spray (Expts. 3-4) also

Figure 5.16: Time-radius cross-section of azimuthally averaged sensible heat flux H_s anomaly due to spray mediated heat and momentum flux for (a) Expt. 3, (b) Expt. 4 as compared to Expt. 2. The contours of ΔH_s are from -60 W/m^2 to 60 W/m^2 with an interval of 20 W/m^2 , where dashed line represent negative anomaly and solid line represent positive anomaly. The radial distance (abscissa) is from TC centre and time (ordinate) from 0000 UTC July 2 through to 1200 UTC July 4, 2014.

cause slight broadening of core size compared to Expt. 2.

Furthermore, radius-time plot of sensible (Figure 5.16) and latent (Figure 5.17) heat flux anomalies for Expt. 3 and 4 show that spray mediated heat fluxes weakens the TC with a broadening of TC core. Also, it is noticed that the sensible heat flux increases just outside the eyewall while decreases within the eyewall when both spray mediated heat and momentum fluxes are included (see Figure 5.16 (b)). Figure 5.18 provide the total latent and sensible heat fluxes averaged over the 200 km radial distance from TC centre for the Expts. 1-4. These heat flux anomalies illustrate that the addition of spray fluxes not only modify the storm intensity but also the low level storm structure, with possible modification to the vertical structure as well.

To examine the effects of coupling wave model and sea spray on the vertical structure of hurricane, Figure 5.19 shows the azimuthally averaged radius-height cross-section of temperature (Figure 5.19 (b,d,f)) and mixing ratio q (Figure 5.19 (a,c,e)) anomaly. The radius-height cross-section utilises the storm centre located at

Figure 5.17: As in Figure 5.16, but for latent heat flux H_l anomaly. The contours of ΔH_l are from -300 W/m^2 to 300 W/m^2 with an interval of 75 W/m^2 .

the surface for all the vertical levels so as to construct a storm relative grid. The anomaly fields were calculated by subtracting azimuthally averaged fields for Expt. 1 from those for Expt. 2-4 respectively. For Expt. 2 (Figure 5.19 (b)) a strong positive anomaly extends from $z = 4$ to 16 km and from $r = 0$ to 140 km, whereas a weak negative anomaly in the near surface region, at radii greater than 40 km. The warming in the upper air region within the hurricane core (i.e. near the eye wall) in Expt. 2 can be attributed to the increase in surface heat fluxes. Comparing Figure 5.19 (b,d), we can see that, when sea spray mediated heat fluxes are added, there is a reduction in upper level warming in core region whereas enhanced cooling in near surface layers at radii greater than 4 km. Also, broadening of warm anomaly in the core can be noted, where the edge of warm anomaly in Figure 5.19 (b) has shifted rightwards compared to the Figure 5.19 (a). This broadening of warm anomaly and the increased cooling in near surface layers can be associated with the decreased storm intensity. Figure 5.19 (d) shows the temperature anomaly when the spray mediated momentum flux is added together with the spray heat fluxes. The first key effect is the enhancement of the warm anomaly in the upper levels compared to Expt. 3. Although the broadening of core is still present, there has been a slight reduction in cooler region compared to the

Figure 5.18: The area-averaged total (a) sensible $\langle H_s \rangle$, (b) latent $\langle H_l \rangle$ heat flux within the 200 km inner core region of hurricane Arthur (2014) from the different experiments described in Table 5.1.

Expt. 3.

Figure 5.19 (a) shows the mixing ratio anomaly for Expt. 2 relative to the Expt. 1 (*i.e.* uncoupled atmosphere model run). A broad region of positive anomaly can be noted extending from radius = 60 km to 160 km. Also, just above this region at radii greater than 120 km, a large region of negative anomaly can be noticed. The largest positive anomalies are in the near surface region in all the experiments, with the maximum occurring in the core region. When the spray heat fluxes are added (Figure 5.19 (b)), the negative q anomaly has shifted from radius = 120 km to 80 km, however with a considerable downward shift of the vertical extent from 12 km to 6 km. However, when both the spray heat and momentum fluxes are utilised (Figure 5.19

Figure 5.19: Height-radius cross sections of difference of azimuthally averaged mixing ratio (left) and temperature (right), for (a,b) Expt. 2, (c,d) Expt. 3 and (e,f) Expt. 4. The difference was calculated by subtracted azimuthal averaged quantity for Expt. 1 from respective coupled model result at 0000 UTC July 3, 2014

(c)), the extent of negative anomaly has shifted back to 120 km radial location from 80 km. Also, worth noting is the increase in negative q anomaly within the eye wall region, where the q values have decreased by -0.6 g kg^{-1} compared to -0.1 g kg^{-1} seen in Figure 5.19 (a,b).

5.6 Summary

There have been limited work in assessing the effects of sea spray mediated fluxes using a coupled atmosphere-wave model, where the sea spray generation function was modelled with wave energy dissipation. Furthermore, most of the previous studies have used bulk approximations of sea spray fluxes when used in conjunction with

atmosphere or a coupled atmosphere-wave model. The aforesaid bulk approximations were formulated as a function of surface wind speed or friction velocity. In the present study, a consistent approach for incorporating sea spray fluxes without relying on bulk approximations was demonstrated. Moreover, a comparison of whitecap fraction obtained from wave energy dissipation to the widely used MOM80 (Monahan and Muirheartaigh, 1980) and recently formulated AH2016 (Anguelova and Hwang, 2016) was presented. It was shown that the method adopted in this study results in whitecap fraction comparable to the results reported by Anguelova and Hwang (2016), while resulting in substantially lower whitecap fraction at higher wind speed compared to MOM80. Due to the limitations in the sea spray microphysical model, a new set of coefficients for incorporating nominal spray fluxes using the HEXOS dataset were obtained.

To investigate the role of surface waves and sea spray fluxes, a two-way coupled atmosphere-wave model was utilised. The coupled model was developed using a flexible coupler where, different processes (such as sea spray physics) were integrated at the coupler level. Within the coupled model, sea spray fluxes were incorporated as discrete contribution of a spectrum of spray droplets, where the spray droplet generation was modelled using the ocean wave energy dissipation due to whitecapping. The results from four different model simulations were analysed to elucidate the effects of wave induced surface roughness and spray mediated heat and momentum fluxes on the distribution of sensible and latent heat as well as the temperature and mixing ratio among different model coupling scenarios. Furthermore, the wave model results from different numerical experiments were compared with the measurements obtained from the floating offshore buoys and satellite altimeter.

As illustrated in Figure 5.5, the model developed in present study captures the lifecycle of simulated TC relatively well, where the uncoupled atmosphere model results in somewhat weaker TC while the coupled model results in relatively stronger TC compared to the best track data. Furthermore, all the simulated TCs traverse westward of the best track data, however when the atmosphere model is coupled with a surface wave model, the TC track shifts east of the uncoupled model track. Also, compared to uncoupled atmosphere model, the coupled model simulated storms are

able to attain the maximum velocity similar to that of best track data but all the simulated storms attain maximum intensity almost 12 hours before that observed in best track data. Despite all the foregoing differences, it behoves us to argue that the numerical experiments performed in the present study are adequate for conducting a preliminary investigation of the role of ocean waves and sea spray fluxes.

Moreover, in the recent literature, a number of reasons have been associated with the bias in the simulated TC tracks. These include the effects of dataset used for model initialisation, initialisation time of the model run, atmosphere model resolution, and physics scheme used for cumulus parametrization. All of these are the topics of active research, where a number of advanced techniques such as ensemble Kalman filter and advanced data assimilation have been developed to alleviate the effects of initial condition error, TC bogussing for reducing the bias due to initialisation time, and Large Eddy Simulation (LES) and super-parametrization are being used to improve the cumulus parametrisation in numerical models. It is arguable that running the same numerical experiments with different initial conditions (e.g. ERA-Interim or GFS Final analyses) would be useful to assess the validity of the results presented here, however such tests are beyond the scope of present study, but recommended for future work.

Including the sea state dependent surface roughness increases the sensible and latent heat flux exchange within the surface layer. This study doesn't couple an ocean model, therefore this increase in surface heat fluxes is presumably from the increased surface friction velocity. Moreover, here, the sea spray fluxes (or SSGF) depend on the wave energy dissipation due to whitecapping, therefore it is not possible to distinguish between the effects of coupling sea spray fluxes and ocean surface waves. However, the results discussed here do underscore the significance of the friction velocity in modulating the storm intensity. Furthermore, the results discussed here also alludes to the uncertainty associated with the inclusion of sea spray fluxes, where the limitations are due to the lack of observation data at higher wind speeds and the limited understanding of the underlying physical processes necessary for modelling sea spray fluxes.

Chapter 6

Effects of Ocean Coupling on Tropical Cyclone

6.1 Preamble

In the preceding chapters (chapters 1 and 2), discussion has been focussed on the addition of sensible and latent heat flux from ocean to atmosphere, as the ocean was assumed to be at rest. However the relationship between the ocean and atmosphere is not a one-way interaction (as briefly discussed in Chapters 1 and 2. The surface stress exerted by the atmosphere on the ocean and the negative pressure (atmospheric) anomaly associated with the TC induces acceleration in the ocean, and associated changes in the ocean circulation. The aforementioned negative pressure anomaly results in rise of mean sea level beneath the storm, where this risen water moves with the TC, and contributes to the storm surge at the time of landfall. An conceptual description of ocean state under the TC is shown in Figure 6.1. The illustration is shown along a west-east transect, where the tangential stress (wind induced) is directed into (out of) the plane to the right (left) of TC centre indicated by $R_{max} = 0$, where R_{max} the radius of maximum wind. The strong wind stress induced circulations in the ocean mixed layer (OML) results in enhanced mixing due to the wind stirring and shear production. Even though the density stratification in ocean continuous, for simplicity, Anthes (1982) ocean is divided into two layers, where the upper layer (*i.e.* OML) is assumed to have a constant temperature (due to vertical mixing), while

Figure 6.1: (a) A satellite image of TC, (b) An illustration of physical processes forced by hurricane winds such as shear-induced mixing, ocean mixed layer (OML) deepening, upwelling below the storm and surface heat fluxes from ocean to atmosphere, (from Shay (2010, p. 94))

the lower layer is where the temperature decreases with the increasing depth. The tangential stress due to the winds affect the ocean (more specifically OML) in two ways, where firstly, it increases the turbulence in the OML, thereby entraining cooler water from below the thermocline (see Figure 6.1), thus resulting in deepening of mixed layer. Secondly, the curl of stress induces region of upwelling where the curl is positive and downwelling where the curl is negative (Anthes, 1982). The region of positive stress curl is inside the radius of maximum wind (RMW), where this positive stress curl induces horizontal divergence of water in the OML (and the upwelling), while the regions of negative stress curl is outside the RMW, where convergence is induced by the anticyclonic relative vorticity causing downwelling.

Additionally, these strong winds generate surface waves with wave heights of

amplitude 20m or more, where the effects of surface waves on the atmosphere has been investigated in the preceding chapter (see Chapter 4). The most widely observed consequence of TC passage over ocean is the SST cooling. This SST cooling varies spatially with maximum cooling occurring to the right (left) of the TC track in northern (southern) hemisphere. From Anthes (1982), the various mechanisms for producing

Figure 6.2: Sea surface temperature cooling from before to after Hurricane Betsy (1965), (from Anthes (1982, p. 136))

these lower temperatures can be summarised as:

- upwelling of cooler water
- wind-incuded vertical mixing of deeper cooler water to surface
- horizontal advection of cooler water by ocean currents driven by TC
- loss of heat to the atmosphere (shown by Q_o in Figure 6.1)

The observed TC induced cooling during Hurricane Betsy (1965) is shown in Figure 6.2. These observations show the temperature change from before and after hurricane Betsy, where a temperature decrease of upto $4^{\circ}F$ to the right of TC track can be seen. It is also worth mentioning that Betsy was a moderate storm with a central pressure of 950 mb, reduced SST by several degrees over a large area. This TC induced cooling is usually observed as a cold wake behind the eye of the TC. Shay

(2010, p. 95), as the SST continue to decrease, the heat and moisture fluxes decrease accordingly (see eq. (3.8)), *i.e.* the more cooling of SST, the lesser H_s , H_l , and lower TC intensity. In addition to the TC intensity, this SST cooling can also affect the TC structure, where the asymmetric distribution of SST cooling around the storm eye, as well as weaker mechanical mixing at higher radial distance from the TC centre would have a considerable effect on the upper boundary layer structure. Within the two-layer

Figure 6.3: The current velocity field obtained from a simplified primitive equation model where the wind stress was moving from right to left with a translation speed of 5 m s^{-1} , (from Ginis (2002, p. 88))

fluid system described above, the wind at the upper surface excites two modes of response namely the barotropic and baroclinic. In the former mode both layers are affected in a similar way *i.e.* the current velocity in upper and lower layer will be same. This mode of response would exist in a fluid with no horizontal or vertical variation of density (*i.e.* unstratified). In the latter mode of response, a vertical shear develops (*i.e.* velocities in upper and lower layer are different), and the interface is distorted with respect to the free surface. This mode depends on the vertical density differences between the adjacent layers. In an idealised situation, both baroclinic and barotropic modes are produced when a TC passes over the quiescent ocean. Figure 6.3 shows the surface currents due to baroclinic mode of response obtained from a two layer ocean model forced with hurricane wind stress which was moving from right to left with a translation speed of 5 m s^{-1} . From Figure 6.3, the strong asymmetric distribution

of the baroclinic currents relative to the storm track can be noted. This rightward bias occurs because the inertially rotated currents are effectively accelerated on the right and decelerated on the left. This is due to the rotation of the wind stress vectors produced due to the moving storm. Besides the effects of TC translation speed, Price (1981) argued that this observed asymmetry is due to the resonance effects between the winds and currents in the higher wind speed region that exist to the right of TC track.

In summary, the ocean response to a TC has been shown to be arise from the interaction between the TC winds (tangential stress) and the upper ocean (OML), where the response of OML can be thought to depend on number of factors like upper ocean structure (*i.e.* mixed layer depth, temperature gradient across thermocline, storm translation speed), and latitude location. It is quite obvious that the initial structure of the upper ocean controls the ocean response, as it will govern how easily the cooler water present below the thermocline are entrained and mixes with the relatively warmer water in the OML. Schade and Emanuel (1999) used a simplified coupled atmosphere-ocean model to demonstrate the strong dependence of TC intensity on the translation speed, where in a coupled model, slowly moving TC were able to attain only 50% of the intensity attained in an uncoupled simulation. From the above discussion, these results can be deduced rather intuitively, as slow moving TC will likely result in more surface cooling at a location, and as it is moving slowly, it will spend longer at the respective location, where the cooler surface temperature would results in negative feedback with heat fluxes directed from atmosphere to ocean. The above discussion underscores the inherently coupled nature of the interaction between atmosphere and ocean, and direct consequences for the TC intensity.

6.2 Numerical Experiments

To assess the effects of ocean coupling on the TC, five numerical experiments (see Table 6.1) were carried out. These numerical experiments were designed to include increasing level of interaction among different model components. The first experiment coupled WRF and MIKE3 ocean model, where only SST from ocean model is

Table 6.1: Summary of the 5 numerical experiment detailing the level of ocean and wave interaction

Experiment	Ocean Coupling		Wave Coupling	Sea Spray	
	SST	Currents		Heat Flux	Momentum Flux
1	✓	×	×	×	×
2	✓	✓	×	×	×
3	✓	×	✓	×	×
4	✓	✓	✓	×	×
5	✓	×	✓	✓	×

provided to atmosphere model. In the second experiment, the both SST and ocean current were provided from ocean model to the atmosphere model. The third experiment utilizes the coupling between all three major model components *i.e.* atmosphere, ocean and wave, where the ocean model provides SST to atmosphere model; sea surface height and ocean currents to wave model, while the wave model provides friction velocity to atmosphere model; radiation stresses to ocean model. It should be noted that this experiment doesn't include ocean current interaction, which is included in the fourth experiment, in addition to the aforementioned interactions accounted for in Expt. 3. Lastly, in fifth experiment, effects of sea spray mediated heat fluxes are investigated together with the atmosphere-ocean-wave interaction. For the fifth experiment, the ocean currents were again not considered. In addition to the numerical experiments summarised Table 6.1, Run 1 and 2 described in Table 4.1 will also be used to study the effects of ocean coupling.

6.2.1 Observed Ocean Response to Hurricane Arthur (2014)

Figure 6.4 compares the daily average SST obtained from Advanced Microwave Scanning Radiometer AMSR2 (Wentz et al., 2014) before and after Hurricane Arthur (2014), where the TC track is overlaid for the sake of clarity. It is noticeable that Arthur traversed over moderately warm ocean with surface temperature $\approx 28\text{-}29^\circ\text{C}$. Additionally, Figure 6.5 shows the surface temperature anomaly computed from the SST before and after Arthur, where cooling of upto 1.5°C is seen to the right of TC track. The SST anomaly was computed using Operational Sea Surface Temperature and Sea Ice Analysis OSTIA (Donlon et al., 2012) data, which is an optimally in-

Figure 6.4: Spatial variation of SST (a) before, (b) after the passage of Hurricane Arthur (2014) from AMSR2 Wentz et al. (2014). The magenta line indicate the TC track.

terpolated product with a relatively coarse horizontal resolution, where some of the detailed features gets filtered out resulting in a smoothed dataset. In addition to the

Figure 6.5: Spatial variation of SST anomaly showing the TC induced cooling along the TC track. The anomaly were calculated from Operational Sea Surface Temperature and Sea Ice Analysis (Donlon et al., 2012) data before and after the passage of Hurricane Arthur.

SST measurements, observations of surface currents along the east coast of US are also available from several land-based high frequency HF coastal radars. Figure 6.6 shows the surface currents as observed by the HF radar network during the passage of Hurricane Arthur near Cape Hatteras, North Carolina, and Figure 6.7 shows the maximum current velocity across the Gulf Stream at 35.22 °N latitude (just north of Cape Hatteras). As the hurricane passed through the region, it affected the surface currents, with much weaker surface currents of maximum current speed 1.5 m s^{-1} from 0000 UTC to 0600 UTC on July 4, while stronger currents with maximum current speed greater than 2 m s^{-1} on 1200 UTC with clear anticyclonic rotation due to the surface stress vector. Traditionally, ocean currents are not included when

Figure 6.6: Surface currents from CODAR data along the east coast of US at (a) 1800 UTC July 3, (b) 0000 UTC July 4, (c) 0600 UTC July 4, (d) 1200 UTC July 4. The CODAR data used here was on a 6km resolution horizontal grid, (from <http://hfradar.ndbc.noaa.gov/>)

Figure 6.7: Time series of maximum current velocity (across the Gulf Stream) at 35.22 °N latitude

computing the surface stress in atmosphere model, as the maximum current velocity are considerably smaller as compared to the wind speeds prevalent in hurricanes. A recent study by Renault et al. (2016) have shown that the ocean currents can modulate the momentum flux from atmosphere to ocean, although Renault et al. (2016) did not investigate the effects of ocean currents on the hurricane. Therefore, in addition to the effects of SST, additional numerical experiments including the effects of ocean currents were carried out. In the next section (section 6.3), the potential effects of this noticeable increase in current velocity, and the associated change in current direction on the surface stress in atmosphere model are discussed. Additionally, section 6.3 also describes the modification to the atmosphere model necessary for properly including the effects of ocean currents.

6.3 Ocean Current Interaction with Atmosphere

In a recent study, Renault et al. (2016) argued that the effect of ocean currents is also relevant to the air-sea interaction, however, the response of atmosphere to the ocean currents have not been studied. Using a coupled ocean-atmosphere model, they showed that the current feedback modulates the energy transfer from atmosphere to

ocean, which then controls the ocean eddy kinetic energy (EKE). In their study, by including the effect of currents in atmosphere model, it was shown that the currents have a noticeable effect on surface stress, which also affects the winds aloft.

In the NWP models, the surface stress is usually parameterised as a function of surface wind speed, and the effects of ocean currents are largely ignored, even though ocean currents apply drag force on the atmosphere. Following Renault et al. (2016), ignoring the effects of ocean currents can lead to over estimation of energy input from atmosphere to ocean, thus the effects of ocean currents should be including when estimating surface stress. In this section, a brief description of the changes required in the atmosphere model to include the effects of ocean currents is provided. As mentioned in Lemarié (2015), the atmosphere model used in present study (*i.e.* WRF) requires modification both in surface layer (where surface stress is estimated) and PBL scheme (where tridiagonal matrix for vertical diffusion) to properly account for ocean currents. The latter is needed because in WRF model, the tridiagonal matrices are build locally in each PBL scheme, therefore the modification have to be made for all the different PBL schemes used. To account for the ocean currents in surface stress estimation, first consider following equations:

$$\begin{aligned}\frac{\partial}{\partial t}(\mathbf{u}_a) &= \frac{\partial}{\partial z} \left(K_v^a(z) \frac{\partial \mathbf{u}_a}{\partial z} \right) \\ \frac{\partial}{\partial t}(\mathbf{u}_o) &= \frac{\partial}{\partial z} \left(K_v^o(z) \frac{\partial \mathbf{u}_o}{\partial z} \right)\end{aligned}\tag{6.1}$$

where

$$\begin{aligned}K_v^a(z) \partial_z \mathbf{u}_a &= \overline{\mathbf{u}'_a w'}(z) \\ K_v^o(z) \partial_z \mathbf{u}_o &= \overline{\mathbf{u}'_o w'}(z)\end{aligned}\tag{6.2}$$

The subscripts (and superscript) a and o refer to atmosphere and ocean respectively, $\mathbf{u} = (\mathbf{u}, \mathbf{v})$ represent the velocity vectors, K_v is the coefficient of turbulent viscosity, $\overline{\mathbf{u}'_a w'}$ and $\overline{\mathbf{u}'_o w'}$ are the vertical turbulent fluxes, and z is the vertical coordinate in the respective models. Now, if both u_a and u_o are in the same coordinate system, then $\overline{u'_a w'}|_b$ will be the bottom boundary condition in atmosphere model, whereas $\overline{u'_o w'}|_s$ will be the boundary condition at the surface in ocean model. Therefore at the air-sea

interface, wind stress can be written as:

$$\tau = \rho_a \overline{\mathbf{u}'_a w'}|_b = \rho_o \overline{\mathbf{u}'_o w'}|_s \quad (6.3)$$

where τ is the wind stress vector, and can be written in modified form as:

$$\tau = \frac{u_*^2}{\|\mathbf{u}_a|_b - \mathbf{u}_o|_s\|} (\mathbf{u}_a|_b - \mathbf{u}_o|_s) \quad (6.4)$$

Here, it is clearly seen that the surface stress $\|\tau\| = \rho u_*^2$, and the velocity vector $\mathbf{u}_a$ is replaced with $\mathbf{u}_a|_b - \mathbf{u}_o|_s$. It is worth mentioning that this modification is only carried out in the atmosphere model, and the effect of ocean currents is implicitly included in both ocean and wave models when the wind speeds are provided from atmosphere model to the respective model component. In addition to modifying the expression for

Figure 6.8: Schematic of variable arrangement on the oceanic and atmospheric vertical grid on both sides of air-sea interface

surface stress estimation, the righthand side of tridiagonal matrix for vertical diffusion have to be modified. Section 6.3 illustrates a typical grid structure in vertical direction, where u_* is usually located at cell centers in horizontal direction, while velocity vectors (*i.e.* u_a, v_a, u_o, v_o) are located at cell edges. The ocean grid spans from $k = 1$ at the bottom surface to $k = N_o$ at the sea surface, while the atmospheric grid spans from $k = 1$ at the sea surface to $k = N_a$ at the model top, here N_o, N_a are the number of levels in ocean and atmosphere model respectively. Typically, backward Euler method is used to integrate eq. (6.1), where for given variable X ($X = u_a, v_a, u_o, v_o$), discretised

eq. (6.1) can be written as:

$$\frac{X_k^{n+1} - X_k^n}{\Delta t} = \frac{1}{\Delta z_k} \left[K_{k+\frac{1}{2}}^x \frac{X_{k+1}^{n+1} - X_k^{n+1}}{\Delta z_{k+\frac{1}{2}}} - K_{k-\frac{1}{2}}^x \frac{X_k^{n+1} - X_{k-1}^{n+1}}{\Delta z_{k-\frac{1}{2}}} \right] \quad (6.5)$$

which, after some rearrangement can be written as:

$$\begin{aligned} X_k^n = & -X_{k+1}^{n+1} \left[\frac{\Delta t}{\Delta z_k} \frac{K_{k+\frac{1}{2}}^x}{\Delta z_{k+\frac{1}{2}}} \right] - X_{k-1}^{n+1} \left[\frac{\Delta t}{\Delta z_k} \frac{K_{k-\frac{1}{2}}^x}{\Delta z_{k-\frac{1}{2}}} \right] \\ & + X_k^{n+1} \left[1 + \frac{\Delta t}{\Delta z_k} \frac{K_{k+\frac{1}{2}}^x}{\Delta z_{k+\frac{1}{2}}} + \frac{\Delta t}{\Delta z_k} \frac{K_{k-\frac{1}{2}}^x}{\Delta z_{k-\frac{1}{2}}} \right] \end{aligned} \quad (6.6)$$

where, on the RHS, first term is at upper level, second term at lower level while the third term is at the respective model level (*i.e.* where tridiagonal matrix is being constructed). The eq. (6.6) can be expressed in matrix form $\mathbf{A}\mathbf{X} = \mathbf{F}$, where $\mathbf{A}$:

$$al_k X_{k-1}^{n+1} + ad_k X_k^{n+1} + au_k X_{k+1}^{n+1} = X_k^n \quad (6.7)$$

and $\mathbf{X}$ as:

$$\mathbf{X} = (X_1^{n+1}, X_2^{n+1}, \dots, X_N^{n+1}) \quad (6.8)$$

where N is the number of vertical levels in respective model (*i.e.* ocean and atmosphere). In eq. (6.7), terms al_k , ad_k and au_k are

$$\begin{aligned} al_k &= -\frac{\Delta t}{\Delta z_k} \frac{K_{k-\frac{1}{2}}^x}{\Delta z_{k-\frac{1}{2}}} \\ au_k &= -\frac{\Delta t}{\Delta z_k} \frac{K_{k+\frac{1}{2}}^x}{\Delta z_{k+\frac{1}{2}}} \\ ad_k &= (1 - al_k - au_k) \end{aligned} \quad (6.9)$$

where al_k , ad_k , au_k are the terms at lower diagonal, diagonal and upper diagonal respectively. As the effect of ocean currents is only included at the bottom level (*i.e.* $k = 1$) in atmosphere model, therefore the bottom boundary condition for u_a using

eqs. (6.2) to (6.5) can be written as:

$$K_{\frac{1}{2}}^a \frac{u_a^{n+1}|_{k=1} - u_a^{n+1}|_{k=0}}{\Delta z_{\frac{1}{2}}} = \left[\frac{u_*^2}{\|u_a^n|_{k=0} - u_o^n|_{k=N_o}\|} \right] (u_a^{n+1}|_{k=1} - u_o^n|_{k=N_o}) \quad (6.10)$$

and eq. (6.5) can be expressed as:

$$\begin{aligned} u_a^{n+1}|_{k=1} - u_a^n|_{k=1} = & \frac{\Delta t}{\Delta z_1} \left[K_{\frac{3}{2}}^u \frac{u_a^{n+1}|_{k=2} - u_a^{n+1}|_{k=1}}{\Delta z_{\frac{3}{2}}} \right. \\ & \left. - \left(\frac{u_*^2}{\|u_a^n|_{k=0} - u_o^n|_{k=N_o}\|} \right) (u_a^{n+1}|_{k=1} - u_o^n|_{k=N_o}) \right] \end{aligned} \quad (6.11)$$

where following eq. (6.7)

$$\begin{aligned} a_1 &= 0 \\ a_{u_1} &= -\frac{K_{\frac{3}{2}}^u \Delta t}{\Delta z_{\frac{3}{2}} \Delta z_1} \\ a_{d_1} &= 1 - a_{u_1} + \left[\frac{u_*^2}{\|u_a^n|_{k=0} - u_o^n|_{k=N_o}\|} (u_a^{n+1}|_{k=1} - u_o^n|_{k=N_o}) \right] \end{aligned} \quad (6.12)$$

Finally, using eqs. (6.9) and (6.12), it can be noticed that the $\mathbf{F}$ (in traditional matrix form) for u_a should be modified as follows:

$$f_1 = u_a^n|_{k=1} + \left[\frac{u_*^2}{\|u_a^n|_{k=0} - u_o^n|_{k=N_o}\|} (u_a^{n+1}|_{k=1} - u_o^n|_{k=N_o}) \right] \quad (6.13)$$

Similar modifications have to be carried out for v_a . It can be said that the modification given in eq. (6.13) correctly applies the directional effects of ocean current on the computation of surface stress.

6.4 Ocean Model Spinup

As discussed in Section 3.2, in order to couple ocean model (*i.e.* MIKE3), in addition to the coupled model experiments, two additional simulations were carried out, where the first simulation was run to obtain the tidal boundary condition, while the second simulation was run to obtain initial conditions for ocean model during coupled model

simulation. To ascertain the validity of the tidal simulation and the tuning of surface parametrisations (see section 3.2.1) used in the ocean model, comparison of sea surface height (from tidal model) and sea skin temperature (just below sea surface) are compared with the observed sea surface height (SSH) and skin temperature are shown in Figures 6.10 and 6.11. The SSH obtained from the tidal model was compared

Figure 6.9: The unstructured grid used in ocean model with the location of four tidal gauge stations.

for four different tidal stations shown in Figure 6.9. These tidal stations were chosen for their diverse geographical location, thereby providing sufficient variability to examine the accuracy of tidal model. Furthermore, the SSH from tidal model was compared for a period of 4 days (see Figure 6.10) preceding the hurricane in the model domain. It can be noticed that the tidal model predictions were quite accurate with slight underestimation at times, however, it is very encouraging to note that the model was able to reproduce the tidal phase very well. As mentioned earlier, in addition to the tidal model simulation, ocean model was spunup so as to provide initial conditions for the coupled model setup. Figure 6.11 compared the ocean skin temperature at three buoy locations. In addition to the output of the ocean model used in present study, the data from global model used for initialising MIKE3 is also shown in Figure 6.11. As the global model dataset is only available at an interval of one day,

Figure 6.10: Comparison of surface elevation obtained from tidal model with the observed data at four tidal stations indicated in Figure 6.9.

therefore both observations and MIKE3 output were subsampled for the comparison. The model output shown in Figure 6.11 were compared for a period of three weeks, thus providing a sufficiently longer duration to ascertain the model performance. It behoves us to point out that no data assimilation is applied in MIKE3 model, and MIKE3 model was run with output from the atmosphere model used in the present study. From Figure 6.11, it is evident that the ocean model (MIKE3) performs quite well as compared to the global model (MyOcean), where the global model is warmer compared to the observed temperature. In addition to the temporal comparison of modelled skin temperature shown in Figure 6.11, spatial comparison of daily mean SST obtained from ocean model (*i.e.* MIKE3) was also conducted. Figure 6.12 compares the SST obtained from Optimally interpolated OSTIA (Figure 6.12 (b)), global model (Figure 6.12 (c)) and MIKE3 (Figure 6.12 (d)) to the observed SST obtained from AMSR2 (Figure 6.12 (a)). To facilitate an objective comparison, the SST data from AMSR2 was subsampled along east-west transect (Figure 6.12 (a)) spanning from $27^{\circ}N$ to $38^{\circ}N$ at an interval of 1° . From Figure 6.12, it is noticeable that all the products (and models) have a negative bias as compared to AMSR2,

Figure 6.11: Time series of observed and modelled surface temperature at NDBC buoys (a) 41002, (b) 41025 and (c) 44014 during the MIKE3 spinup period. The surface temperature from global ocean model data used for boundary conditions and initialising MIKE3 is also shown.

with MIKE3 exhibiting the highest bias (*i.e.* coolest sea surface temperature). It is also noted that neither models (*i.e.* MyOcean and MIKE3) nor OSTIA were able to reproduce observed SST values greater than 30 °C. This negative bias exhibited by MIKE3 can be attributed to the tuning coefficients used in MIKE3 model, as well the input data provided by the atmosphere model. Although, this negative bias in SST has direct implications for the surface fluxes used in atmosphere model, it can be argued that the MIKE3 model setup is appropriate for the purpose of this study.

Figure 6.12: Spatial comparison of observed (a) AMSR2 SST with that from different products, *i.e.* (b) Optimally Interpolated OSTIA product, (c) MyOcean, and (d) MIKE3. The scatted plot compare daily mean SST on 28 June 2014, sampled along the east-west transects (solid horizontal lines) spanning from 27 °N to 38 °N at interval of 1 °.

6.5 Results and Discussion

As described in Section 6.2, the effects of coupling ocean model on the TC are examined. To that end, five different numerical experiments described in Table 6.1 were carried out. The 60h modelled hindcast of Hurricane Arthur from the aforesaid numerical simulations is shown in Figure 6.13. These numerical simulations are initialised at 0000 UTC June 30, 2014 and integrate for a period of 120h, where only the modelled tracks from 0000 UTC July 2 to 1200 UTC July 4, 2014 are shown in Figure 6.13 (a). Along with the TC tracks Figure 6.13 (b) and Figure 6.13 (c) compare

the MSLP and V_{max} obtained from the various numerical experiments to the TC best track obtained from the National Hurricane Centre (NHC). From the TC tracks it can

Figure 6.13: The NHC best track and modelled tracks for five different numerical runs (a) summarised in Table 6.1, (b) Minimum sea level pressure, and (c) Maximum wind speed at 10m elevation.

be noticed that all the numerical experiments exhibit a westward bias as compared to the best track data, with the maximum bias exhibited by Run 2. Additionally, it is of note that the track bias reduces substantially during the landfall. Furthermore, it is seen that besides the Run 2, Run 1 has a more westward track, while in Runs 3 and 5, modelled hurricane traverses a track with the least westward bias. The V_{max} (Figure 6.13 (c)) that during the early stages, the storm strengthens to attain the TC intensity, with a very good agreement with the best track data, however, after 1200 UTC July 3, 2014, the TC fails to intensify any further, and starts to weaken after 1800 UTC July 3, 2014. Figure 6.14 (a) quantifies the anomaly in TC center location, while Figure 6.14 (b) shown the anomaly in V_{max} during the different stages TC lifetime for the different numerical experiments. These anomaly are computed in reference to the best track. Contrary to the previous assertions, it is seen that Run 5 has the least track anomaly out of the five numerical simulations, however Runs 1 and 4 have similar track anomaly with slightly higher track error. Similar to the track anomaly, for most part of the storm lifetime, Run 5 has the least intensity error as compared to the other numerical experiments.

Figure 6.14: Modelled TC (a) center location, and (b) intensity from 0000 UTC July 2 to 1200 UTC July 4, 2014 for five numerical runs, *i.e.* atmosphere-ocean (red), atmosphere-ocean with current (green), atmosphere-ocean-wave (blue), atmosphere-ocean-wave with current (yellow) and atmosphere-ocean-wave with spray heat flux (purple).

For various reasons, none of the numerical simulations were able to match the best track or TC intensity, it is surprising that the numerical experiments Runs 3 and 4 were able to attain comparable MSLP (see Figure 6.13 (b)), although the model simulations intensified much more rapidly as compared to the best track. Also, from the nearly identical features exhibited by the model tracks, it can be said that these anomalies arise solely due to the air-sea coupling, with the varied effects of the different physical processes considered (in the five numerical experiments) at the air-sea interface. For instance, in Run 3, besides the exchange of radiation stress from wave model to ocean model and exchange of surface elevation and ocean currents from ocean model to wave model, the wave induced surface roughness is applied in the atmosphere

model as compared to Run 1, where the roughness length is obtained from a bulk parametrisation, where the wave induced surface roughness has a significant effect on TC intensity as shown in Chapter 4. Although, similar to Chapter 4, it is seen that including wave induced surface roughness results in a stronger storm, here, it is also noticed that the coupling of ocean model causes the storm to under intensify. To examine the effects of surface heat fluxes on the storm intensity, Figure 6.15

Figure 6.15: Time-radius Hovmöller plots of azimuthally averaged surface latent heat H_l flux from (a) Run 1, (b) Run 3 from 0000 UTC July 2 to 1200 UTC July 4, 2014, with contours of H_l plotted at an interval of 100 W/m^2 .

shows the time-radius hovmöller plots of azimuthally averaged surface latent heat fluxes obtained from Run 1 (Figure 6.15 (a)) and Run 3 (Figure 6.15 (b)) for the 60h period starting from 0000 UTC July 2, 2014. Additionally, Figure 6.16 presents the time-radius plot of azimuthally averaged sensible heat fluxes for the Runs 1 and 3. Finally, Figure 6.17 compares the area averaged sensible heat flux H_s (Figure 6.17 (a)) and latent heat flux H_l (Figure 6.17 (b)) within the 200km radial distance from the TC center for the five numerical experiments described in Table 6.1. Examining the differences in the surface heat fluxes, it is seen that the sensible heat fluxes in Run 3 are significantly higher than Run 1, with sensible heat flux upto twice as much

Figure 6.16: As in Figure 6.15, but for H_s sensible heat flux with contours of H_s plotted at an interval of 50 W/m^2 .

in Run 1. Besides the substantial differences in H_s , there are modest increase in latent heat flux with values about 25% more than that in Run 1 within the 50km from TC centre. As expected, Run 2 has the smallest value of heat fluxes, similar to the TC intensity comparison. Although, it should be pointed out that there are two different reasons responsible for the lower intensity in Run 2. The first reason being the bulk formulation used for the roughness length, and the second reason is the inclusion of ocean currents in surface stress computation in atmosphere model. The inclusion of ocean currents results in lower surface stress which has a direct implication on the storm intensity. To further illustrate the effects of flux calculation (both momentum and enthalpy fluxes), Figure 6.18 compares the coefficient of drag C_d , coefficient of enthalpy C_k and ratio of the exchange coefficient C_k/C_d from the different numerical experiments at 1800 UTC July 3, 2014. Although, the TC is not at the peak intensity in any of the model simulations, it is seen that the Run 2 has the lowest values of both C_k and C_d . Also, the effects of coupling wave model is seen via the scatter exhibited by the results from Runs 3,4 and 5. The C_d values obtained

Figure 6.17: The area averaged total (a) sensible, (b) latent heat flux for the five numerical runs described in Table 6.1 within 200 km of TC centre over 60h of model integration from 0000 UTC July 2, 2014.

from Runs 1 and 2 increase monotonically with wind speed, and exhibit minimal scatter due to their direct dependence on wind speed. It is important to remind that in the present study, the upper limit of C_d value in the bulk formulation is relaxed thus allowing an objective comparison between the different models. When examining the C_d obtained from the Runs 3-5, it can again be noticed that the values increase with wind speed, however with no indication of saturation as argued by previous studies (discussed in Chapters 1 and 2). The C_d values are approximately same in Run 1 and 3 till wind speed of 25 m/s , after which the C_d is higher in Run 3 with values upto twice as much in Run 1. Despite being a diagnostic quantity, C_d is indicative of the transfer of horizontal momentum in vertical direction at a given wind speed value. It governs how much momentum atmosphere loses due to surface friction.

Figure 6.18: The model simulated (a) coefficient of drag C_d , (b) coefficient of enthalpy C_k and (c) ratio of exchange coefficient (C_k/C_d) as a function of mean wind speed U_{10} in the whole model domain from five numerical runs at 1800 UTC July 3, 2014.

A higher value of C_d in low and moderate winds may lead to decrease in winds in outer regions of storm, while a lower stress in high wind region may lead to further intensification. Besides the effects on momentum flux exchange, C_d also affects the transfer of enthalpy fluxes, where a higher value of C_d results in higher heat fluxes, resulting in opposite effects as compared to surface friction. As alluded in preceding sections, higher winds (or surface stress) enhance mixing in the ocean, resulting in a colder wake in areas where the upper ocean was more stratified. In the cases examined in present study, all these processes are occurring simultaneously, thereby making it

Figure 6.19: Comparison of SFMR obtained data, (top left) SFMR tracks with surface wind speed, (top right) Wavenumber 0+1 component of SFMR obtained wind speed, and mean wind speed obtained from model run (bottom left) Run 1, (bottom right) Run 3. The time for the model field are shown on lower right corner, and modelled and observed data are centered on storm location in earth coordinates.

rather difficult to ascertain the contribution of each process in modulating the final state of atmosphere and ocean. To evaluate the structure and intensity of TC from Runs 1 and 3, Figure 6.20 uses surface wind speed measurements from the stepped frequency microwave radiometer (SFMR) obtained from the flight surveys into Hurricane Arthur by National Oceanic and Atmospheric Administration (NOAA) and United States Air Force (USAF). Figure 6.19 (a) illustrates the flight track and wind speed obtained using SFMR on an earth relative coordinate system centered on TC location at 2100 UTC July 2, 2014. These tracks were post-processed using observations collected during different flight missions, and the best track data from NHC was used to locate the TC

Figure 6.20: Scatter comparison of SFMR wind speed from Figure 6.19 with modelled run (a) Run 1, (b) Run 3 along radial direction within 200km radial distance from storm centre.

centre. The measured data was decomposed into azimuthal wavenumber components following Uhlhorn et al. (2014), where the decomposed wind speed components are expressed as:

$$V(r, \theta) = V_0(r) + V_1(r) \cos(\theta - \theta_1(r)) + \epsilon(r, \theta) \quad (6.14)$$

where V is the measured wind speed, V_0 and V_1 are the amplitude of the wavenumber components 0 and 1 and θ_1 is the phase of wavenumber 1 (hereafter referred as WN1) and ϵ is the remainder containing the contribution of all the higher wavenumbers. The terms r and θ are the radius and azimuth of the polar coordinate system. As de-

scribed in Uhlhorn et al. (2014), the spatial density of SFMR is sufficient to represent the lowest 2 wavenumber components, therefore following Uhlhorn et al. (2014), the observed SFMR wind speeds were fitted with eq. (6.14). The terms thus obtained *i.e.* $V_0(r)$, $V_1(r)$ and $\theta_1(r)$ can be used to reconstruct a spatial wind speed field comprising of wavenumber 0+1. Figure 6.19 (b) shows the reconstructed wind speed field representing the WN0+WN1. Figure 6.19 (c and d) shows the mean wind speed at 10m elevation obtained from Runs 1 and 3 at the corresponding time instant. Unlike

Figure 6.21: Comparison of measured (left column) and modelled direction wave spectrum from Run 1 (middle column) and Run 3 (right column) at NDBC 44014 at 0000 UTC July 2 (top row), 1200 UTC July 3 (middle row) and 0000 UTC July 4 (bottom row). The directional spectrum are normalised by the peak value, and the colour shading starts at 10% of peak value, increasing linearly at 10% interval. The directional spectrum are plotted in nautical convention with wave direction (direction waves are coming from).

Uhlhorn et al. (2014), the results presented in Figure 6.19 aren't normalized using radius of maximum wind (RMW), so as to avoid any potential distortion that may be introduced by the asymmetric distribution of winds in the inner region. The size and intensity of the TC from Runs 1 and 3 vary as compared with the SFMR observations. Also, it is noted that there is small bias in the direction of maximum wind between the model simulations and measurements, where the direction of maximum wind is more to the southeast while that in model simulations is along the east. The radial distribution of wind speed is further illustrated in Figure 6.20, where Figure 6.20 (a) compares the SFMR wind speed with the mean wind speed obtained from Run 1 while Figure 6.20 (b) compares SFMR measurements with wind speed from Run 3. It is noticed that the model simulations have a good agreement with the measurements, where both the model simulations (*i.e.* Runs 1 and 3) show similar RMW, however, Run 3 has slightly higher wind speed in the inner region as compared to SFMR, while Run 1 wasn't able to attain the maximum wind speed observed in SFMR.

In addition to investigating the effects of ocean coupling on atmospheric fields, directional wave spectrum obtained from the coupled atmosphere-wave model (discussed in chapter 4), Run 3 and NDBC buoy 44014 are compared in Figure 6.21. The NDBC buoy used here is a pitch-and-roll buoy that has a 3-m discus hull and records time series of surface elevation and sea surface slope along two directions. The data obtained from the buoys are processed and disseminated by NDBC through their website. The directional wave spectrum shown in Figure 6.21 is further post-processed using Fourier Expansion Method (Longuet-Higgins et al., 1963), where the coefficients for reconstructing directional wave spectrum were obtained from NDBC. The directional spectrum thus obtained from buoy measurements are compared with the model at three different time instances. The measured directional spectrum is wider than that obtained from model simulations. This artifact in the observed wave spectrum is due to the limitations in the sampling capabilities of the pitch-and-roll buoys. The directional wave spectrum shown in Figure 6.21 is normalized using the peak value, thus allowing a comparison of key characteristics of modelled and measured wave energy spectrum. Furthermore, it permits an objective comparison as the effects of wind intensity on the wave energy are removed. Due to the dependence of surface

Figure 6.22: TC induced upper ocean cooling due to Hurricane Arthur for (a) Run 1 and (b) Run 3. The TC track for Run 1 and 3 are shown with solid white line during 60 model integration from 0000 UTC July 2, 2014.

stress in atmosphere model over the wave state, the spatial distribution of directional wave spectrum has direction implication on the skill of atmosphere model. It is seen that both the model simulations show a good comparison with the measurements, where both the wave direction and period of the peak wave energy were modelled reasonably well. Figure 6.22 shows the swath of minimum sea surface temperature during the 60 hour model integration period starting from 0000 UTC July 2, 2014 for Run 1 (Figure 6.22 (a)) and Run 3 (Figure 6.22 (b)). The SST swaths are shown along with the TC track for the respective model simulation. From the model results it can be seen that Arthur traversed over the Gulf Stream, with a westward shift as it moved further north thus crossing away from Gulf Stream before the landfall. From the SST swaths, it is worth noting that TC induced cooling is much stronger during the early stages for Run 1 as compared to Run 3, however as TC moved over the shallower depths and gained more intensity, the TC induced cooling is slightly higher for Run 3. This higher cooling can also be attributed to the wave radiation stresses, which would affect the ocean currents in shallower depths. Due to the dynamic nature of variable exchange between atmosphere and ocean model, it is somewhat difficult to explain the warming noticed in Run 3 as compared to Run 1. This warming could be a result of enhanced entrainment in horizontal direction in ocean model or convective

Figure 6.23: Vertical profile of (a) ocean temperature, (b) zonal, and (c) meridional velocity along TC track for Run 1 over 60h model integration starting from 0000 UTC July 2, 2014. The vertical profile are plotted for upper 75m ocean depth.

heat fluxes from atmosphere model directed towards the ocean model. Figure 6.23 reveals the along track sub-surface ocean temperature, zonal and meridional current velocities within the upper 75m water depth for Run 1, whereas Figure 6.24 shows the aforementioned variables for Run 3. Due to the lack of sub-surface measurements, it is not possible to gauge the accuracy of the model results. From the sub-surface results of ocean temperature, a mixed layer of depth 45m is noticed in Run 1, before the TC moves over shallower depths, whereas for Run 3, there is much stronger sub-surface

Figure 6.24: As in Figure 6.23, but for Run 3.

cooling, possibly due to the higher storm intensity. Also, worth noting that the zonal currents in Run 1 are somewhat weaker as compared to Run 3, again possibly a consequence of TC intensity. Besides the effects of TC on the sub-surface temperature and current velocity, these results are also affected by the location of TC centre with respect to the Gulf Stream. As the present study is motivated towards studying the role of air-sea interaction, therefore investigation of the effects of Gulf Stream on the sub-surface variables or TC for that matter is beyond the scope of present study. Also, the duration of model simulations and number of case studies performed are rather limited to fully ascertain the effects of such large scale processes. Figure 6.25 (a-d)

Figure 6.25: Composite of azimuthally averaged radar composites of (a) tangential velocity, (b) radial velocity, (c) radar reflectivity, (d) absolute angular momentum for Hurricane Arthur at 11:32 UTC July 3, 2014.

Figure 6.26: Radius-height cross section of azimuthally averaged tangential velocity (shaded) with contours (white) of tangential velocity at an interval of 5 m/s and superposed with azimuthally averaged vectors (V_r, W) for (a) Run 1, (b) Run 3 at 1200 UTC on July 3, 2014.

shows the radius-height cross-section of azimuthally averaged tangential velocity V_t , radial velocity V_r , radar reflectivity and absolute angular momentum (AAM) obtained from the aircraft mounted doppler radar. The radar data was collected onboard the aircraft missions flown into Hurricane Arthur, where the data shown in Figure 6.26

corresponds to the TC centre location at 11:32 UTC July 3, 2014. The radar data extends from surface to 20km with a horizontal and vertical grid spacing of 2km and 0.5km respectively. Lastly, the radar data shown is mainly observed above 500m, therefore the data is mostly above the boundary layer, and due to the increase in cone volume of radar with distance, has a rather coarse resolution near the surface. The

Figure 6.27: Radius-height cross section of azimuthally averaged absolute angular momentum (shaded) with contours (blue) of virtual potential temperature at an interval of 5 K and contour (black) of vertical velocity at an interval of 0.1 m/s for (a) Run 1, (b) Run 3 at 1200 UTC on July 3, 2014.

inner core region (and eyewall) is characterized by the high reflectivity from surface through to troposphere, representing the inner rainband and relatively weak reflectivity in the outer rainband. It is also seen that the storm develops a strong inflow in the boundary layer region while an outflow in the upper troposphere. In addition to the strong inflow in the boundary layer region, there also exist a weaker inflow in the mid-troposphere outside the eyewall region, albeit somewhat weak due to relatively weaker storm intensity. Although not presented here, there also exist updrafts in the inner rainband shown in Figure 6.25 (c). Furthermore, a strong primary circulation (Figure 6.25 (a)) with tangential velocity greater than 35 m/s is seen just above the boundary layer at a radius of 40km. The surface of AAM obtained from the radar composite are rather upright in eyewall in the middle-to-lower troposphere and tilt outward in the upper troposphere.

When comparing the radius-height plots of tangential wind speeds (Figure 6.26) obtained from Runs 1 and 3, it is seen that both model simulations result in similar vertical structure with some subtle differences. Consistent with the higher intensity

of the storm in Run 3, the tangential velocity is higher in Run 3 as compared to Run 1. Also, the radial inflow and outflow are much stronger in Run 3, with an intense updraft (Figure 6.27) and an upright eyewall region, whereas the eyewall region is more tilted in Run 1. Additionally, Run 3 also has a relatively warmer inner core region as exhibited by the contours of θ_v shown in Figure 6.27, superposed on the AAM, where the updraft follows the surface of constant AAM. The results shown here again highlight the importance of surface friction (or C_d) for the intensification of a TC, even though the model simulations discussed in present chapter grossly under intensified as compared to the best track data.

6.6 Summary

In this chapter, the fully coupled atmosphere-ocean-wave model developed over the course of this research was used to study the effects of ocean coupling. The two main processes under investigation were the SST and ocean currents. Similar to the previous Chapters 4 and 5, the coupled was applied to study the effects of coupling on hurricane Arthur (2014). Five numerical experiments were carried out to illustrate the effects of different physical processes on a TC, where more in-depth results were only presented for the experiments with and without wave coupling. Aside from the coupled model experiments, this chapter also described the procedure to include the effects of ocean currents on surface stress and vertical diffusion in atmosphere model. Furthermore, the validation and tuning of ocean model during spinup is described. It was found out that the current tuning resulted in a slight cold bias as compared to the satellite observations. In light of such results, it was therefore suggested that in addition to the effects of ocean coupling, this cold bias would have an additional influence on the model results.

The model experiments were compared with a wide range of measured datasets which include observations from airborne doppler radar, wind speeds from SFMR, directional wave spectrum and water temperature from moored buoys. The model results showed reasonable agreement with the measurements, where including wave model in the atmosphere-ocean coupled model results in better correlation with the

measurements.

For the numerical experiments carried out in this chapter, it was suggested that coupling ocean model results in decrease in TC intensity, which is consistent with the previous studies. It was also shown that when using wave coupling with atmosphere-ocean model, the TC intensity is higher as compared to without wave coupling, which is again inline with earlier findings. A key observation to be made here was that the TC over-intensification as noticed in Chapter 4 wasn't there when ocean coupling was included. Additionally, it was found out that by including the effects of ocean currents in Run 4, the TC intensity is further reduced compared to Run 3. This is another key observation, that the model skill can be improved by properly accounting for the relevant physical process in place of often unjustified (sometime vague) model tunings. Furthermore, this finding supports the development and usage of coupled model as opposed to uncoupled model.

Lastly, it is necessary to point out that the ocean model used in current form doesn't use the surface flux computed in atmosphere model, thus limiting the applicability of these results. Additionally, the flux parametrisations used in the ocean model are not similar to the one in atmosphere model, which shall result in an inconsistency within the current coupled model. Therefore, it is likely that the results (and skill) of present coupled model will change with the possible modification of ocean model to include the flux quantities computed in the atmosphere model. Furthermore, it is important to mention that the present study only uses a specific set of physics options in the atmosphere model, where the model results would likely change with the usage of different set of physics in the atmosphere model. As has been stated earlier, this study doesn't include data assimilation, further limiting the accuracy (skill) of the model results.

Chapter 7

Conclusions and Outlook

As discussed in the preceding chapters, TCs have the potential to cause catastrophic damage by inflicting significant economic loss, destruction of property and loss of life. To minimise the aforementioned undesirable consequences, it is therefore necessary to be able to accurately forecast the TC location and intensity with a sufficient lead time so as to take necessary evacuative measures. This requirement of accurate forecast has increasingly become important with the rising population and development of infrastructure in the coastal regions.

As stated in Chapters 1 and 2, despite the improvements in the model skill in forecasting TC locations, intensity forecasts have largely remained poor with large intensity error. This is especially true for the high-resolution regional NWP model, where despite the higher resolution, the models currently in use are unable to resolve the small scale processes, specifically the fluxes of momentum and enthalpy at the air-sea interface. Furthermore, the model skill is limited by the deficiencies of the bulk parameterisations that are used in these NWP models, due to the lack of the field measurements under hurricane strength conditions. Therefore, a large amount of effort has been directed towards improving model skill, where the modelling community has pursued two different approaches for improving the model results. The first approach has largely concentrated on improving parameterisations of air-sea fluxes by tuning the terms applied in the parameterisation schemes, conducting large measurement campaigns during hurricanes and using model tools like large eddy simulations. The second approach has concentrated on developing coupled models where the

atmosphere model is coupled with both an ocean and wave models, thus eliminating the usage of bulk parameterisations or minimising the dependency on such relations.

7.1 Overall summary

The research described in this thesis was spurred following the latter approach with the belief that the coupled modelling approach provides a consistent physically based framework for investigating the role of air-sea fluxes in modulating intensity and structure of a TC. Chapters 1 and 2 discuss the background literature on the air-sea interaction and coupling techniques. Chapter 3 describes the atmosphere, ocean and wave models and the necessary considerations to be made when developing a coupled model. The different physics parameterisations used in the individual model components are also discussed. Meanwhile, Chapter 4 tests the effects of different parameterisation schemes for exchange coefficients of momentum and enthalpy. It was found that the TC track largely remains unaffected for different surface flux parameterisation, while the TC intensity is extremely sensitive to the different schemes. It should be pointed out that Chapter 4 largely focuses on the effects of coefficient of enthalpy because the coefficient of momentum is computed using surface stress obtained from the wave model. From the numerical experiments carried out in Chapter 4, the invalidity of widely accepted notion, *i.e.*, the coupling of wave model with atmosphere model will result in reducing TC intensity. Furthermore, by comparing the field observations with model results, it was shown that the coupled model improves modelled wind speeds and wave heights, because it was able to better model the regions with higher wind speed.

Following the development of a coupled atmosphere-wave model in Chapter 4, Chapter 5 investigates the effects of sea spray heat and momentum fluxes on the TC intensity. In the past literature, it has been strongly argued that the contributions of sea spray fluxes may be important for the accurate modelling of TC. In order to include the sea spray fluxes, a sea state dependent spray model was developed and included with the coupled model. It is worth noting that instead of using any bulk estimation of spray contribution, the present study strived to include the spray effects

by modelling the contribution of a spectrum of spray droplets at each coupling time step. Additionally, Chapter 5 presents the coefficients used to transform the nominal fluxes computed from spray model to the enthalpy fluxes used in the atmosphere model. The numerical experiments have shown that including spray heat fluxes reduces the intensity of modelled TC. This reduction in TC intensity was attributed to the cooling of the near surface layer due to the heat extracted by the evaporating droplets in the DEL. It was also shown that it is necessary to include the effects of sea spray on heat and momentum fluxes, as the spray mediated momentum flux applies additional drag force on the near surface wind.

The different numerical experiments considered in Chapters 4 and 5 used the daily mean SST obtained from the global dataset, thus largely ignoring the effects of spatial and temporal evolution of SST. As discussed earlier, TC intensity is highly sensitive to SST, therefore, in Chapter 6, a fully coupled model was used to study the effects of SST on TC intensity and structure. Chapter 6 also describes the various steps necessary to include the effects of ocean currents on surface stress computation in atmosphere model and its subsequent effects on TC evolution. From the model experiments, it was shown that the coupling of ocean model reduces the TC intensity. It was likewise illustrated that compared to atmosphere-ocean model, the fully coupled model results in better agreement with the best track data. It was also shown that the inclusion of ocean current velocity in surface stress computation reduces the TC intensity by reducing the coefficient of drag and enthalpy. This was a significant observation as it demonstrated the relevance of important but largely ignored physical processes at the air-sea interface. Although the results presented in this thesis are encouraging, further coupled modelling studies using more TC cases are necessary to fully elucidate the complex interaction between wind, wave and ocean currents.

7.2 Future perspectives

From the usage of the model, a number of possible modifications (or future developments) concerning the atmosphere-ocean-wave coupled model arise, which can be carried out in future studies:

- In the present model, the variables from atmosphere to ocean model (and vice-versa) are exchanged using bilinear remapping. Although the current model system can exchange variables sufficiently well, methods like bicubic interpolation can be used in order to conserve "the higher order property, such as the curl of wind" (Valcke et al., 2012). Furthermore, other methods like "conservative remapping" (known for conservation of fluxes) can be tested.
- The interaction between the atmosphere and wave models is currently facilitated by using the surface roughness computed in wave model. Moon et al. (2004b) have presented an alternative approach, where a wave boundary layer model is applied to couple the atmosphere and wave models. Within the wave boundary layer model, special attention is paid to the separation of kinetic energy within the near surface layer. The wave boundary layer model attempts to account for the viscous stress, wave induced stress and turbulent stress in a consistent way. It is therefore suggested that a wave boundary layer model can be used in conjunction with the atmosphere model and wave model, where the surface roughness length rather than being obtained from wave model, is computed from the wave boundary layer model.
- The ocean and wave models used in this study interact through the exchange of radiation stress from wave model to ocean model while ocean currents and water level from ocean model to wave model. Recent studies such as those by Reichl et al. (2016) and Walsh et al. (2017) have shown that the ocean surface waves can enhance the turbulence in the upper ocean, increasing mixing, thus resulting in enhanced SST cooling and deepening of ocean mixed layer. This enhanced cooling can have direct consequences for the TC by reducing the enthalpy flux supplied from ocean to atmosphere.
- Studies like the one conducted by Chen et al. (2013) have utilised a very high resolution atmospheric model grid with a vortex following nest for modelling TCs, with the view that at such a high resolution, the numerical model is able to resolve convection reasonably well. One of the limitations of the study presented here was the inability to use a high resolution grid in atmosphere and

ocean models due to computational constraints. Therefore, it is recommended that in future studies, the model developed here be used with a high resolution atmosphere (and ocean) model grid.

- In the present study, the sea state dependent spray model was utilised, where a limited set of observation datasets was used to transform the nominal sea spray fluxes computed from the spray model into spray enthalpy fluxes that can be included in the atmosphere model. One potential avenue for investigation could be the usage of an expanded dataset collected from diverse meteorological and oceanic conditions for transforming the aforementioned nominal fluxes.

Finally, the current work only investigated a single case of a north Atlantic hurricane. However to fully realise the usability of the coupled model, it is suggested that more studies including cases in different ocean basins should be conducted. Furthermore, the TC studied here only attained a strength of category 2 on Saffir-Simpson scale. It would be very relevant to investigate the effects of coupling models on TCs with a strength higher than category 2.

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