Understanding the Effect of Atmosphere-Wave-Ocean Coupling on Tropical Cyclone Structure

Chia-Ying Lee
University of Miami, clee@rsmas.miami.edu

Follow this and additional works at: https://scholarlyrepository.miami.edu/oa_dissertations

Recommended Citation
https://scholarlyrepository.miami.edu/oa_dissertations/902
UNIVERSITY OF MIAMI

UNDERSTANDING THE EFFECTS OF ATMOSPHERE-WAVE-OCEAN COUPLING ON TROPICAL CYCLONE STRUCTURE

By

Chia-Ying Lee

A DISSERTATION

Submitted to the Faculty
of the University of Miami
in partial fulfillment of the requirements for
the degree of Doctor of Philosophy

Coral Gables, Florida

December 2012
UNIVERSITY OF MIAMI

A dissertation submitted in partial fulfillment of
the requirements for the degree of
Doctor of Philosophy

UNDERSTANDING THE EFFECTS OF ATMOSPHERE-WAVE-OCEAN
COUPLING ON TROPICAL CYCLONE STRUCTURE

Chia-Ying Lee

Approved:

Shuyi S. Chen, Ph.D.
Professor of Meteorology and
Physical Oceanography

M. Brian Blake, Ph.D.
Dean of the Graduate School

William M. Drennan, Ph.D.
Professor of Applied Marine Physics

Ralph Foster, Ph.D.
Senior Physicist
University of Washington

Mark A. Donelan, Ph.D.
Professor of Applied Marine Physics

James F. Price, Ph.D.
Senior Scientist
Woods Hole Oceanographic Institution

Jimy Dudhia, Ph.D.
Project Scientist
National Center for Atmospheric Research

Chun-Chieh Wu, Ph.D.
Professor of Atmospheric Science
National Taiwan University
Lee, Chia-Ying (Ph.D., Meteorology and Physical Oceanography)

Understanding the Effect of Atmosphere-Wave-Ocean Coupling on Tropical Cyclone Structure (December 2012)

Abstract of a dissertation at the University of Miami.

Dissertation supervised by Professor Shuyi S. Chen.

No. of pages in text. (173)

It is well-recognized that tropical cyclone (TC) intensity is strongly modulated by air-sea interactions. However, how and to what extent air-sea interactions affect TC structure remains an open question. The overall goal of this study is to better understand the physical processes of the atmosphere-wave-ocean couplings and their impact on TC structure. Because the boundary layer connects the air-sea interface to TC convection, it is also important to understand how the couplings modulate boundary layer structure. In this study, coupled atmosphere-(wave)-ocean models and observations from two field programs are used in this study: Coupled Boundary Air-Sea Transfer (CBLAST, 2003-04), and Impact of Typhoons on the Ocean in the Pacific (ITOP, 2010). High-resolution numerical experiments with and without ocean and/or wave couplings are conducted for Hurricane Frances (2004), Typhoon Choiwan (2009), and Typhoon Fanapi (2010). Results show that both ocean- and wave-couplings cause significant changes in TC and TC boundary layer structures. In particular, a stable boundary layer forms over the storm-induced cold wake. Tracer and trajectory analyses in a coupled-model simulation suggest that the stable boundary layer thermodynamically suppresses convection in and downstream of the cold wake, and dynamically causes the surface wind to turn further
inward. The stabilized air tends to stay in the boundary layer longer and penetrate further into the eyewall. This stabilized air then brings extra energy into the eyewall due to enhanced fluxes downstream of the cold wake.

The boundary layer in a TC has been seen as a passive layer that is driven by both the TC vortex above and by the ocean underneath. This study shows that the boundary layer, when in the presence of the storm-induced cold wake, can actively influence TC structure through the formation of an internal stable boundary layer. Although the stable boundary layer is a consequence of the TC-induced cold wake that has a negative impact on TC intensity, it appears counter-intuitive that the stable boundary layer has a positive impact on TC intensity via this separate mechanism.

In summary, we find that atmosphere-wave-ocean coupling affects boundary layer structure and the physical properties of the near-surface air flow in TCs, which in turn changes the convective organization and eventually affects TC structure, energetics and intensity. This indicates that atmosphere-wave-ocean coupling affects TC structure via complex physical processes. Hence it is difficult to parameterize the atmosphere-wave-ocean coupling processes in TCs without a fully coupled model.
Acknowledgments

I would like to gratefully acknowledge my advisor Dr. Shuyi Chen for her valuable advice and patient guidance. Her encouragement and support also contributed tremendously to this work. I would like to thank my committee members Dr. Chun-Chieh Wu, Dr. Jimy Dudhia, Dr. Ralph Foster, Dr. James Price, Dr. Will Drennan, and Dr. Mark Donelan for their insight to this work and valuable comments. I want to express my deep gratitude to Dr. Brandon Kerns for endless science discussion and technical support. I would like to thank my groupmates: Falko Judt, Milan Curcic, Patricia Sanchez-Rodriguez, Jie Ming, and Wei Zhao. I always get useful feedback from them in group meetings. I would like to thank Marc Michelsen for computer troubleshooting. Many thanks to MPO and RSMAS Graduate Study Office staff, especially Sandrine Apelbaum for her assistance in the past several years. I want to thank all my friends for their support and understanding in the past five and half years. Special thanks to Daniel Stern for many discussions about science, language and other topics, to Bjoern Lund for providing UM lyx template, to Cheryl Chan who is like my family in Miami, and to Jan-Huey Chen and other TDRCers for their endless support. Finally, I would like to express my appreciation to my family for their love and encouragement. This Ph.D. work was supported by ONR ITOP DRI grant (N000140810576).
# Contents

| List of Figures | viii |
| List of Tables | xxv |

## 1 Introduction

1.1 Background and motivation ...................................... 1
1.2 Science objectives .............................................. 9
1.3 Outline .......................................................... 9

## 2 Literature Review

2.1 Air-sea interaction in TCs ........................................ 10
  2.1.1 Theory of TC energetics ...................................... 10
  2.1.2 Feedback of TC induced ocean response ...................... 14
  2.1.3 Impact of surface waves on TC intensity and structure ... 17
2.2 Atmospheric boundary layer in TCs ............................. 22
  2.2.1 Conventional definitions and the characteristics of the boundary layer depth ......................................................... 22
  2.2.2 Boundary layer stability ........................................ 29

## 3 Methodology

3.1 University of Miami Coupled Model System (UMCM) ............ 32
3.1.1 UMCM-MWP ............................................. 33
3.1.2 UMCM-WP ............................................. 34
3.2 Data ....................................................... 36
  3.2.1 Airborne observations ................................. 36
  3.2.2 Satellite observations ................................ 40
3.3 Forward Lagrangian trajectory and tracer analysis ................. 41

4 Symmetric and Asymmetric Hurricane Boundary Layer Structure in TCs 44
  4.1 Hurricane Frances (2004) ............................... 45
  4.2 Numerical experiments ................................... 46
  4.3 Simulated track and intensity ............................. 46
  4.4 Surface winds and air-sea fluxes ......................... 48
    4.4.1 Spatial and temporal variations ....................... 48
    4.4.2 Symmetric and asymmetric structures ................. 54
  4.5 Hurricane boundary layer structure ...................... 56
    4.5.1 Vertical profiles of winds and temperature .......... 58
    4.5.2 Symmetric and asymmetric structures in HBL height .... 65
    4.5.3 Effects of storm motion and deep tropospheric inflow on HBL ... 69
  4.6 Conclusions ............................................. 72

5 Tropical Cyclone Structure and Stable Boundary Layer in Super Typhoon

  Choiwan (2009) ............................................. 74
  5.1 Super typhoon Choiwan (2009) .......................... 75
  5.2 Numerical experiments .................................... 77
  5.3 Forecasts of Choiwan in the coupled model ................ 78
  5.4 Stable boundary layer in TCs ............................. 80
  5.5 The impact of stable boundary layer on TC structure .......... 85
    5.5.1 The evolution of near surface airflow ................ 87
5.5.2 Thermodynamic and dynamic forcing associated with stable boundary layer ........................................... 95
5.6 Stable boundary layer and TC energetics ................................................................. 102
  5.6.1 Evolution of equivalent potential temperature along the airflow . . . 102
  5.6.2 TC efficiency ................................................................. 105
5.7 Conclusions and discussions ......................................................................................... 107

6 Coupled Observations and Modeling over the Cold Wake in Typhoon Fanapi (2010) 110
  6.1 Typhoon Fanapi (2010) .................................................................................... 111
  6.2 Numerical experiments ....................................................................................... 116
  6.3 Simulated track and intensity ............................................................................. 117
  6.4 Oceanic response ............................................................................................... 121
  6.5 Spatial distribution of surface winds and air-sea fluxes ........................................ 128
    6.5.1 Observed and simulated winds .................................................................. 128
    6.5.2 Observed and simulated air-sea fluxes ....................................................... 132
  6.6 Spatial variability in TC and oceanic structures across the cold wake . . . . . 134
    6.6.1 Thermodynamic structures and its relation to the distribution of convection ........................................... 137
    6.6.2 Simultaneous temperature profiles across the boundary layers . . . 140
    6.6.3 Boundary layer height and stability ....................................................... 144
    6.6.4 The dynamic characteristic of the surface airflow .................................... 146
  6.7 The modulation of TC energetics due to the cold wake ........................................ 151
  6.8 Conclusions ......................................................................................................... 154

7 Discussion: Potential Impacts of Boundary Layer Parameterization on Numerical Results 157

8 Overall Conclusions 160
List of Figures

1.1 (a) Terra/Moderate Resolution Imaging Spectroradiometer (MODIS) visible satellite image of Typhoon Fanapi at 0220 UTC 18 September, when Fanapi was approaching Taiwan. (b) Tropical Rainfall Measuring Mission Satellite (TRMM) rainfall data at 0653 UTC on 18 September. The red circle marks an approximate outer edge of the inner core. 

1.2 A schematic illustration of radar reflectivity in a Northern Hemisphere TC (Adapted from Willoughby 1988 and Houze 2009). The black arrow shows the direction of cyclonic wind, and the red arrow indicates the location of the schematic diagram in Fig. 1.4.

1.3 Example of TC structure from a numerically simulated TC: (a) Horizontal map of 10-m wind speeds. (b) Horizontal map of rain rate. (c) Vertical cross section of radar reflectivity (shading), equivalent potential temperature (red contour), and radial and vertical winds (black arrows) along the thick black arrow in (a) and (b).
1.4 A conceptual model of air-sea interaction in a TC. This is a vertical cross section from the storm center to the outer rainband, as indicated by the red arrow in Fig. 1.2. Black, purple, and orange arrows indicate the direction of airflow at the selected locations. The thick black arrows represent the secondary circulation, which includes the frictional inflow and the eyewall updrafts. $\theta_e$ along the frictional inflow increases with decreasing radius ($\theta_{e3} > \theta_{e2} > \theta_{e1}$) and then it remains constant along the eyewall updrafts. The thin black arrows indicate the convective updrafts and downdrafts in rainbands. The purple and orange arrows indicate the mid-level inflow and upper level outflow. The dashed green line represents the top of the HBL. Below the vertical cross section is the SST, where the colder color (bluish) represents colder SST; this is the location of storm induced SST cooling. At the air-sea interface, there are upward heat fluxes and downward momentum fluxes depicted by the red and blue arrows, respectively. The size of arrows indicates the magnitude of these fluxes.

2.1 The energy cycle of a mature hurricane. Air spirals inward close to the sea surface from point A to point B and acquires heat energy from the ocean. Air then ascends in the eyewall from B to C without gaining or losing energy. Between C to D, air loses the heat into the storm environments. Eventually, from D to A, air returns to its starting point. (Adapted from Emanuel, 1986)

2.2 (a) SST around the track of Hurricane Tess (1975). Tess moved toward the northwest at ~ 6 m s$^{-1}$. The SST minimum occurs 50 – 150 km to the right of the hurricane track. (b) Temperature along section AB in (a). The base of the ML is shown as a heavy dashed contour. The 200 km wide and 40 m amplitude upwelling beneath the track is the response to the positive stress curl of the hurricane. (From Price 1981)
2.3 (a) Spatial variation of SST after Hurricane France (2004) passage from Geostationary Satellite (GOES). Blue line shows the observed storm track. (b) Similar to (a) but from high resolution fully coupled model and magenta line shows the simulated storm track. (From Chen et al., 2012b)

2.4 Spatial variation of significant wave height ($H_s$) measure by NASA airborne scanning radar altimeter in Hurricane Bonnie on 24 August 1998. Contours for integer values of wave height (in meters) are solid and contours for integer value plus 0.5 m are dashed. (From Wright et al., 2001).

2.5 (a) Simulated significant wave height (color, m) and mean wave propagation direction (white vector) and (b) mean wavelength (color, m) and surface wind (black vectors) for Hurricane Frances (2004) from high resolution fully coupled model. The black “+” indicates the storm center and the arrow in the lower left corner indicates the direction of the storm motion. (From Chen et al., 2012a)

2.6 Laboratory measure of the neutral stability drag coefficient. The drag coefficient versus the equivalent wind speed measured at standard anemometer height of 10 m. (From Donelan et al., 2004).

2.7 The spatial variation of the simulated drag coefficient in Hurricane Frances from (a) fully coupled model (b) uncoupled model. The black lines indicate the four quadrants relative to the storm motion as shown by the black vectors. (c) and (d) are the corresponding scatter plots of the drag coefficient and enthalpy exchange coefficients. Data from the four quadrants are shown in different color as indicated at the top right corner. (From Chen et al., 2012a)
2.8 Vertical cross section of the composite differences in the virtual potential temperature (K) between each level and the near-surface value. The thick black lines denotes the constant contour with a value of 0.5 K. (From Zhang et al. 2011) .......................................................... 23

2.9 (a) Vertical cross section of the axisymmetric inflow layer. Arrows are wind vectors scaled in the same proportions as the axes. Heavy lines are the unsmoothed and smoothed tops of the inflow layer. The solid lines are the contours with constant absolute angular momentum ($\times 10^5$ m$^2$ s$^{-1}$). Dashed lines are for constant equivalent potential temperature, $\theta_e$. (From Frank 1984) (b) Vertical cross section of composite radial wind velocity. The white dashed line is the height of inflow layer while the back thick line is the height where radial wind speed is 10% of the peak inflow. (From Zhang et al. 2011) .......................................................... 24

2.10 Schematic diagram of the characteristics height scales of HBL based on the composite analysis of the dropsondes data. The $h_{inf}$ is the inflow layer depth (red dashed line), $z_i$ is the mixed layer depth (green dashed –dotted line) and $h_{vmax}$ is the height of the maximum wind speed. The solid black line represents the height where the bulk Richardson number is equal to 0.25. (From Zhang et al. 2011) .......................................................... 26

2.11 Mean profiles of storm-relative winds from dropsondes between 40- to 100-km radius for Hurricane George (1998). Each group contains averages over four quadrants and over the whole storm. (From Kepert 2006a) ........... 28

3.1 Schematic of University of Miami Coupled atmosphere-wave-ocean Model (UMCM) system and the options for atmospheric, surface wave, and ocean circulation models. The coupling parameters between each of the components are printed next to the black arrows. (From Chen et al., 2012a) ....... 33
3.2 Drawing of the various observations deployed into Hurricane Frances (2004) during CBLAST field campaign. (From Black et al., 2007) .......................... 36

3.3 Schematic illustrating of the resources of the ITOP (2010) field campaign. (From http://www.eol.ucar.edu/projects/itop/) ................................. 37

3.4 The Cold Wake flight module. .............................................. 38

3.5 The flight pattern for two of ITOP missions executed at 2200 UTC 16 (orange line) and 2100 UTC 17 September (green line). Symbols indicate the locations of various observations. Squares are GPS dropsondes, stars are AXBTs, triangles are the EM-APEX floats, and circles are the UCSD ADOS drifters. .......................................................... 39

4.1 (a) The NHC best-track (black) and three model simulated tracks of Hurricane Frances (UA-blue, AO-green, and AWO-red) from 27 August to 6 September 2004. (b) Observed and simulated maximum wind speeds from 1200 UTC 30 August to 0000 UTC 01 September during which when airborne observations are used in this study. .......................... 47

4.2 The model simulated surface wind speeds (a-c), enthalpy fluxes (d-f), and momentum fluxes (g-i) from UA, AO, and AWO, averaged over a 2-h period centered at 1800 UTC 31 August. The black arrows indicate the direction of storm motion. The white line in (a)-(c) marks a reconnaissance flight path of the NOAA WP-3D aircraft, where the aircraft measurement will be shown in Fig. 4.3. ......................................................... 49

4.3 Surface wind speeds from the Stepped Frequency Microwave Radiometer (SFMR) measurement (black) and three model simulations (UA-blue, AO-green and AWO-red) along the flight path indicated in Fig. 4.2. The SFMR data is collected during the time period from 1650 UTC to 1800 UTC 31 August while the model fields are sampled at 1700 UTC 31 August. .......................... 50
4.4 Time series of model simulated (a) sea surface temperature anomaly (°C), (b) azimuthally averaged peak surface wind speed at the radius of maximum wind (RMW), (c) latent heat fluxes (W m⁻²), (d) sensible heat fluxes (W m⁻²), (e) enthalpy fluxes (W m⁻²), and (f) momentum fluxes (N m⁻²) averaged over an annular area between radii of 0.5 and 5.0 times of RMW from 1200 UTC 31 August to 0000 UTC 01 September (UA-blue, AO-green, and AWO-red).

4.5 Azimuthally averaged surface tangential (a) and radial (b) wind speed at 1800 UTC 31 August from the model simulations.

4.6 Similar to Fig. 4.5, except averaged fields over each of the four quadrants divided based on the storm forward motion that points to the top (FL: front-left, FR: front-right, RL: rear-left, RR: rear-right).

4.7 Same as in Fig. 4.6, except for the drag coefficients (C_d).

4.8 Storm-relative locations of 34 dropsondes from Hurricane Frances research flights used in this study. The dropsonde data are collected from 30 August to 1 September. The numbers indicate the dropsondes that will be shown individually in Figs. 10 and 11, whereas “*” is dropsondes included only in the mean. The circles indicate radii of 50, 100, 150, and 200 km, respectively. The inner core region within the 50-km radius is enlarged shown at the right. The storm forward motion is pointed to the top.

4.9 The mean profiles of tangential (a-c) and radial (d-f) winds of all dropsondes shown in Fig. 4.8 at RMW (left), 2 times RMW (middle), and outer region greater than 5 times RMW (right) and azimuthally averaged profiles from UA (blue), AO (green), and AWO (red) simulations at 1800 UTC 31 August.
4.10 Observed profiles of tangential winds from 12 dropsondes inside of the 50-km radius and 9 dropsondes in the outer region. The layout of the dropsondes here are only proxy to what shown in Fig. 4.8. 61

4.11 Same as in Fig. 4.10, except for the radial winds (m s$^{-1}$). Negative (positive) values are inflow (outflow). 62

4.12 Same as in Fig. 4.9, except for virtual potential temperature. 63

4.13 Same as in Fig. 4.10, except for virtual potential temperature (K). The gray line indicated the THBL calculated from each sounding. 64

4.14 Mean $\theta_e$ profiles from dropsondes and model simulations. The left column (a, c, e, g) is from the dropsondes inside of 50-km radius inner-core region, whereas the right column (b, d, f, h) from the outer region. The solid lines are the mean profiles in the rear-right quadrant (RR), and the dashed lines are from all other three quadrants (OQ). Numbers of dropsondes in each group are indicated in the parentheses. The model fields are sampled according to the storm-relative locations and times of the dropsondes. 66

4.15 Azimuthally averaged radial winds (color) as a function of radius and height at 1800 UTC 31 August from (a) UA, (b) AO, and (c) AWO simulations. The heights of THBL and DHBL are shown in solid and dashed contours, respectively. The gray lines mark the surface RMW. 67

4.16 Similar to Fig. 4.15, except for averaged over each of the four quadrants. The storm forward motion points to the top. 68

4.17 The coupled model simulated the earth-relative DHBL (a-c) and storm-relative DHBL (d-f) heights in Hurricanes Frances (2004) and Floyd (199), and Typhoon Choiwan (2009). The inflow fields are averaged over a 2-h period in all three cases and the storm forward motion points to the top. 70
5.1 (a) The JMA and JTWC best-track (black) and the forecast tracks from AO (red) and UA (blue) of Choiwan from 0000 UTC 13 to 0000 UTC 17 September 2009. (b) Similar to (a), but for predicted maximum wind speed (MWS). The black dashed line is the best-track from JTWC while the black solid line is that from JMA, indicating the discrepancy between two operational agencies. 75

5.2 (a) The initial SST (°C) field. (b) The initial T100 (°C) field (calculated based on Price, 2009). (c) TMI/AMSRE SST (°C) swath from 13 to 17 September, which presents the minimum SST during the whole period. The black line is the JMA best-track and the black dots denote the observed storm center at 0000 UTC each day during the simulations. (d) Similar to (c) but for the forecast from AO. Because the ocean does not change in UA, the SST swath in UA is the initial SST. 76

5.3 Time series of model simulated surface (a) enthalpy fluxes, (b) latent heat fluxes, and (c) sensible heat fluxes averaged within 400 km in radius from storm center (AO-red, UA-blue). 79

5.4 Surface enthalpy fluxes (W m$^{-2}$) from (a) AO and (b) UA averaged over 12 hours from 0000 to 1200 UTC 16 September. The black contours in (a) and (b) are the SST isotherms. The storms move toward west-northwest as indicated by the black arrow. 80

5.5 Similar to Fig. 5.4 but for 10-m wind speed (m s$^{-1}$). 81

5.6 Similar to Fig. 5.4, but for BL depth (m). 83

5.7 The SBL (black) and stable surface layer in (a) AO and (b) UA averaged over 12 hours from 0000 UTC to 1200 UTC 16 September. The red contour denotes SST with an interval of 0.5 °C, and the thick red line is the 29.5 °C SST isotherm. The storm moves toward west-northwest direction as indicated by the black arrow. 84
5.8 $\theta_v$ profiles in the lowest 1 km from (a) AO and (b) UA at 1200 UTC 16 September. Profiles are grouped based on SST, as labeled at top of each column. For each group, we present only a subset of the sample to avoid unnecessary clutter. The gray lines indicate the profiles with unstable layers near the surface while the red lines indicate the profiles with stable layers only in the BL. The black line is the mean profile of each group. All profiles are from an annulus with an inner radius of 150 km and an outer radius of 400 km. For each figure, the x-axis ranges from 304 to 306 K.

5.9 The radar reflectivity at 80 m on 0000 UTC 16 September overlaid with the initial location of the tracers (blue contours) and trajectories (blue dots) in (a) AO, and (b) UA. In AO, the tracers released in the locations marked by dark-blue contours covers area of cold wake (cold-wake tracer), 90° downstream from cold wake (top one, downstream tracer), and 90° upstream from cold wake (lower one, upstream tracer). The tracer released in the light blue contours covers an annulus with inner radius of 150 km and outer radius of 350 km. Each dark-blue contour covers an area of 1/20 of the annulus. The trajectories are released over the cold wake. All the tracers and trajectories are released at 80 m in height. In UA, tracers and trajectories are released based on the same storm-relative location. The thick gray line denotes the 28.5 °C SST isotherm, which marked the location of the cold wake.

5.10 The 0.05 iso-surface (gray) of the cold-wake tracer in AO (top panels) and UA (bottom panels) at $t_{\text{tracer}} = 0, 20, 60, 120$ minutes. The color shading indicates the SST.
5.11 The azimuthally-integrated tracer concentration (shading) from 0300 UTC to 0600 UTC 16 September in AO (left) and UA (right). The contour shows the azimuthally averaged BL depth, and the dots indicate the RMW at each model level. 

5.12 Similar to Fig. 5.11, but for upstream tracer.

5.13 Similar to Fig. 5.11, but for downstream tracer.

5.14 The fraction of tracer originating from (a, b) cold wake, (c, d) upstream, (e, d) downstream, (g, h) annular area that stays in BL (diamond), enters the eyewall (diamond-line), and ends up in the area outside of BL and eyewall (solid line) (AO-red and UA-blue).

5.15 The simulated radar reflectivity at 1 km altitude in AO (left column) and UA (right column) every 3 hours from 0000 UTC to 0600 UTC 16 September. The magenta contours show the 28.5 and 29 °C SST isotherms. The dashed-black line encloses the area downwind adjacent to the cold wake in AO, and in the same storm-relative location in UA. The storm motion is indicated by black arrow.

5.16 (a) Perturbation field of 10-m winds (shading) and SST (contour) across the cold wake at 0000 UTC 16 September. It covers an area with an inner radius of 150 km and an outer radius of 350 km. (b) Same as (a) but for $V_r$. (c) Same as (a) but for $V_r$. The black diamonds in (a-c) indicate the center of a storm. (d) Perturbation field of 10-m winds in (a) as a function of SST. The white diamond indicates the medium value of the wind perturbations wind at given SST, while the edges of thick (thin) black bar represent 25% (10%) and 75% (90%) tile respectively.
5.17 The 10-meter wind vectors for AO (red) and UA (blue) at 0000 UTC 16 September. The gray shading shows the 1 °C SST cooling and the storm motion is roughly to west-northwest (black line). The cold wake and it adjacent downstream area is enlarged at the right. .......... 98

5.18 Time-averaged 10-m inflow angle (a-c) and radial wind (d-f) in AO (left column), UA (middle column), and the difference field (right column). The black contours in (a), (b), (d), and (e) are the 28.5 and 29 °C SST isotherm and the they indicate the 1 °C SST cooling in (c) and (f). The storm moves toward west-northwest direction as indicated by the black arrow. ........ 100

5.19 (a) The downward turbulent momentum fluxes averaged over the region across the cold wake as shown in Fig. 5.16. (b) The mean SST over the the same area. The x-axis is the distance related to the lowest mean SST, the center of the cold wake. From left to right is from the upstream region to downstream region. ............... 101

5.20 The equivalent potential temperature ($\theta_e$) along a subset of trajectories (16 out of 86) in (a) AO and (b) UA. ....................... 102

5.21 Vertical cross section of $\theta_e$ (shading) and 0.1 m s$^{-1}$ constant vertical velocity contour in AO (left) and UA (right) at 0000 UTC 16 September. The mean values of $\theta_e$ at surface and $\theta_e$ at RMW are showed on top of each figures. ....................... 103

5.22 The equivalent potential temperature along the trajectories as a function of height in (a) AO and (b) UA for the 6-hour trajectory calculation. They are separated into 2 groups: those entering the eyewall (red), and those entering the outer rainband convection (blue). The thick lines show the mean profiles respectively and the numbers indicate the data size of each group. The dots show the mean value of each group at the lowest level. . . 104
5.23 Schematic diagram of the airflow from the rear-right quadrant in (a) an atmospheric-ocean coupled model, and (b) an uncoupled atmospheric model. The dashed line represents a storm and it moves toward left. The white shading represents the tracer and the solid line indicates the Lagrangian trajectory paths. In (a) the air is stabilized by the storm induced strong cold wake; therefore it stays in BL and spirals inward to the storm center. Alternatively, the air in (b) remains unstable because there is no cold wake. Most of the air ends up in the rainband.

6.1 JTWC best-track (black) and simulated (red-AO, blue-UA) (a) tracks, (b) maximum wind speeds (MWS), and (c) minimum sea level pressures (MSLP). In (a), the circle indicates the center of storms at 1200 UTC 14 September, the beginning of the simulations; the diamonds are the center location at 0000 UTC each day. The initial location for the storms in UA and AO are the same. In (b) and (c), diamonds denote the storm intensity every 6 hours.

6.2 (a-c) Microwave satellite estimated rain rate overlaid with the infrared image for Fanapi from late 14 September to the beginning of 18 September. (d) the Polarization Corrected Temperature (PCT) overlaid with the visible image at around 0600 UTC 18 September. Detail time of each swath is labeled on the left-upper corner of the plot.

6.3 (a) The initial SST (°C) field. (b) The initial T100 (°C) (calculated based on Price 2009) field. (c) TMI/AMSRE SST (°C) swath from 14 to 20 September, which presents the minimum SST during this period. The black line is the JTWC best-track. The diamond symbol denotes the observed storm center at 0000 UTC each day from 15 to 20 September and the gray circle indicated the storm center at 1200 UTC 14 September. (d) Similar to (c) but for the AO simulation. Because the ocean does not change in UA, the SST swath in UA is the initial SST.
6.4 Time series (a) SST, (b) enthalpy fluxes (W m$^{-2}$), (c) latent heat fluxes (W m$^{-2}$), and (d) sensible fluxes (W m$^{-2}$) averaged over the area within 500 km distance from the storm center. .......... 118

6.5 10-m wind speed in (a-b) AO and (c-d) UA at around 2000 UTC 18, when there is a dramatic oscillation in MWS. The green "x" indicates the location of the MWS whose value is given each panel. The number in the parenthesis is the number of grid points with the wind speed larger than the MWS - 10 kts in 1.3-km resolution domain, indicating the size of the features with extreme winds. .......... 119

6.6 Radar reflectivity of Fanapi at (a) 1450 UTC 18, (b) 1630 UTC 18, and (c) 1830 UTC 18. The magenta circles indicate the possible meso-vorticies. (Images are from Taiwanese Central Weather Bureau) .......... 120

6.7 (a) Observed SST from the AXBTs co-located with the dropsondes. Star symbols indicate the soundings that are located in the areas with SST less than 28 °C, roughly the location of the cold wake. The black line is the best track from JTWC and diamond symbols show the center of storm at 0000 UTC each day from 15 September. (b) Similar to (a), but for the simulated soundings sampled at the same storm-relative locations in AO. 121

6.8 (a) The trajectories of ADOS drifters from the dropping time, around 0000 UTC 17 September, to 0000 UTC 20 September. (b) Similar to (a), but for simulated drifters in AO. Time series of the observed and simulated (c-d) sea-level pressure, (e-f) 10-m wind speed, and (g-h) SST. .......... 123
6.9 (a) TMI estimated rain rate (mm hr$^{-1}$) at 0654 UTC 18 September overlaid with the EM-APEX floats dropped during Mission #1. The dropping locations are marked as black triangle symbols and black lines indicate the trajectory of floats for three days. The TMI image is shifted to center at the observed storm center at 0000 UTC 18 September. (b) Similar to (a) but for simulated rain rate and model-sampled floats in AO at 0000 UTC 18 September.

6.10 (a) The upper ocean temperature as a function of time and depth from EM-APEX floats. The location of each float is shown in Fig. 6.9. The thin black contours indicate the temperature isotherms of 29, 28, 26, 24, 22, and 20 °C. The black dots mark the mixed-layer depth of each profile. (b) Similar to (a), but for the model sampled floats in AO.

6.11 Similar to Fig. 6.10 but the color now indicates the upper ocean salinity.

6.12 Scatter plots of SST from ADOSs and simulations in (a) AO and (b) UA. The black solid line indicates one to one line while the red (blue) dashed line is the best-fit line in AO (UA).

6.13 10-meter winds from AO overlaid with the surface winds estimated from SFMR (square) and the 10-meter winds estimated from GPS dropsondes (stars) from (a) Mission #1 and from (b) Mission #2. (c) and (d) are analogous to (a) and (b), but from UA. The model time for Mission #1 is 0000 UTC 17 and is 0000 UTC 18 September for Mission #2, approximately the time of eyewall penetrations for both missions.
6.14 Radial distribution of observed (black line: SFMR, red dot: dropsondes, purple dots: ADOS) and simulated (gray) 10-m winds during (a, c) Mission #1 and (b, d) Mission #2. The simulated winds in (a) and (b) are from AO while they are from UA in (c) and (d). The model field is from 0000 UTC 17 September for Mission #1 and from 0000 UTC 18 September for Mission #2, approximately the time of eyewall penetrations for both missions. ................................. 130

6.15 Scatter plots of 10-m wind speed between observations and simulations in (a) AO and (b) UA. The black solid line is one to one line while the dashed line indicates the linear regression line. ......................... 131

6.16 (a) Sensible heat fluxes estimated from co-located dropsondes and AXBT. (b) The model-sampled sensible heat fluxes in AO. The model field is sampled based on the storm-relative location of each sounding. (c) Same as (b) but for UA. (d-f) Similar to (a-c) but for latent heat fluxes. The black arrow indicates the storm motion. Stars indicate the soundings with SST less than 28 °C, as shown in Fig. 6.7. ................................. 133

6.17 Scatter plots for the parameters that are used in bulk algorithmic flux calculation in observations versus that in simulations (red-AO and blue-UA): (a) 10-m wind speed (ms⁻¹), (b) air-sea temperature difference (°C), (c) air-sea moisture difference. (d) and (e) shows the scatter plots for latent and sensible heat fluxes (Wm⁻²). The root-mean-square error between observations and simulations for each parameter is printed on each plot. ................................. 135
6.18 Horizontal map of microwave satellite brightness temperature overlaid with the location of the co-located soundings. Colors of squares indicate sounding locations, green: upstream, blue: cold-wake, orange: downstream, and gray: others. Skew-T diagrams are from the numbered soundings indicated on the horizontal map. The observed time in HHmmSS, as well as the distance from the storm center, is labeled on the top of each plot.

6.19 Similar to Fig. 6.18 but for simulated rain rate and soundings in AO.

6.20 (a) The simultaneous air-sea temperature profiles for the upstream, cold-wake, and downstream soundings. (b) Similar to (a) but for AO simulated soundings. The numbers in the x-axis correspond to the numbers that labeled in Figs. 6.18 and 6.19 for (a) and (b) respectively.

6.21 Air-sea interface parameters estimated by co-located soundings: (a) SST (°C), (b) 10-m wind (m s⁻¹), (c) air-sea temperature difference (°C), (d) air-sea moisture difference, in which the positive value means SST is warmer than air temperature, (e) sensible heat fluxes (W m⁻²), and (f) latent heat fluxes (W m⁻²). Colors indicate the location of soundings: upstream (green), cold-wake (blue), and downstream (orange). The numbers in the x-axis correspond to the numbers that labeled in Figs. 6.18.

6.22 Similar to Fig. 6.21, but for AO sampled soundings. The numbers in the x-axis correspond to the numbers that labeled in Figs. 6.19.

6.23 The virtual potential temperature from (a) observed soundings and (b) AO sampled soundings. Colors indicate the sounding location: upstream (green), cold-wake (blue), and downstream (orange). The height of the colorful squares indicates the BL height. The black squares indicate the tick of 305 K for the corresponding sounding. The numbers in the x-axis correspond to the numbers that labeled in Figs. 6.18 and 6.19 for (a) and (b) respectively.
6.24 (a) Observed SST (shading) and wind vectors from composite ADOS drifter
data from 1800 UTC 16 to 0000 UTC 19 September. (b) Similar to (a) but
the wind vector is color-coded based on the inflow angle. The storm motion
is indicated by the thick black arrows. ........................................... 147
6.25 Similar to Fig. 6.24, but for the simulated drifter data in (a,c) AO and (b,d)
UA. ................................................................. 149
6.26 The 10-meter wind vectors from AO (red) and UA (blue) at 0000 UTC 18
September. The gray shading shows the 1 °C SST cooling and the storm
motion is roughly to west-northwest (black line). The cold wake and it
adjacent downstream area is enlarged at the right. ........................... 150
6.27 The initial location of the tracers (light-blue contours) and trajectories (dots)
in (a) AO, and (b) UA. In AO, the trajectories are released in the upstream
(green), cold wake (blue), and upstream (orange) area with the same cri-
teria use in Section 6.5. In UA, tracers and trajectories are released in the
same storm-relative location. The gray shading is the radar reflectivity (dbz)
and the red contours show constant SST (°C). .............................. 151
6.28 The equivalent potential temperature along the trajectories released over
the (a) upstream of the cold wake, (b) cold wake, and (c) downstream of
the cold wake in AO. (d), (e), and (f) are similar to (a), (b), and (c) but for
UA. ...................................................... 152
7.1 The diagnostic HBL from YSU scheme. ................................. 158
List of Tables

5.1 From right to left, the first three columns are the change of mass-weighted kinetic energy and surface enthalpy fluxes, and the efficiency of converting heat energy to kinetic energy within a control volume (cylinder) with the lateral boundary of 350 km. The forth column shows the efficiency. The fifth column shows a ratio of AO efficiency to the UA one. The right most one is the inward enthalpy fluxes in the BL at the outer edge of RMW (70 km). .......................................................... 106

5.2 The sensitivity test of the storm efficiency to the size of control volume. The top row is the radius of the cylinder, and the second and the third rows are the storm efficiency of UA and AO, respectively. The last row is a ratio of AO storm efficiency to the one in UA. .................................................. 107
Chapter 1

Introduction

1.1 Background and motivation

A Tropical cyclone (TC) is a warm-core, non-frontal synoptic-scale cyclone, originating over tropical or subtropical waters with organized deep convection and a closed surface wind circulation about a well-defined center\(^1\). When TCs are making landfall, the associated hazards, such as high winds and storm surge, cause numerous loss of life and tremendous economic damage. To mitigate the damage caused by TCs, it is essential to accurately forecast TC track and intensity. While TC track is determined mainly by the large-scale atmospheric environment, storm intensity is influenced by multi-scale physical processes and the interactions among them. These include:

1. Interaction between a storm and the large-scale atmospheric environment including the environmental moisture, the vertical wind shear, etc.

2. Internal vortex dynamic processes.

3. Air-sea interaction.

4. Physical processes in the boundary layer (BL).

\(^1\)From NOAA NWS: http://www.nws.noaa.gov/directives/sym/pd01006004curr.pdf
Figure 1.1: (a) Terra/Moderate Resolution Imaging Spectroradiometer (MODIS) visible satellite image of Typhoon Fanapi at 0220 UTC 18 September, when Fanapi was approaching Taiwan. (b) Tropical Rainfall Measuring Mission Satellite (TRMM) rainfall data at 0653 UTC on 18 September. The red circle marks an approximate outer edge of the inner core.
Each of the above presents a problem for the TC intensity forecast. Additionally, the 3-dimensional structure of winds and convection of a storm modulate how these processes affect TC intensity. In spite of the advent of advanced numerical weather prediction models and observations, the prediction of TC intensity and structure remains challenging because of insufficient knowledge of these physical processes. Among these processes, air-sea interaction is perhaps the most fundamental, because it leads to the surface enthalpy and momentum fluxes, which are the ultimate source and sink of energy. These fluxes are locally determined by the dynamics of and the thermal disequilibrium between the atmospheric BL and upper ocean. Therefore, this dissertation will focus on the air-sea interaction and its associated BL processes in TCs and their impact on TC structure.

Figure 2.5a shows a view of a mature TC as seen from above by satellite. The size of a TC is based on the distance from the circulation center to its outer most closed isobar or to the outer most area where the wind speed is larger than tropical storm force, 17 m s\(^{-1}\), (Merrill, 1984). In general, it is of the order of 1000 km. The circulation center is the eye in the middle of the central dense overcast in Fig. 1.1. Because the cloud tops are dominated by the upper-level cirrus, it is not easy to see the convective organization. The convective structure stands out when looking into the rainfall estimates derived from TRMM (Fig. 1.1b). Convective features include eyewall: a heavy rain band (reddish color) circling around the eye; primary rainband: a curved band of heavy rain east of the eyewall; outer/distant rainband: spiraling bands with moderate rainrate further outside (greenish) of the primary rain band. Houze (2010); Fudeyasu and Wang (2011) further separate a TC into inner core, outer core, and storm environment area (Fig. 1.2). The inner core is the region including the eye, eyewall and the primary rainband; we denote it as the red-dashed circle in Fig. 1.1 or the gray-dashed circle in Fig. 1.2. The outer-core is defined as the region immediately outside the inner-core region including the outer rainbands and the associated anvil clouds. Outside of the outer-core is the storm environment area.
Figure 1.2: A schematic illustration of radar reflectivity in a Northern Hemisphere TC (Adapted from Willoughby 1988 and Houze 2009). The black arrow shows the direction of cyclonic wind, and the red arrow indicates the location of the schematic diagram in Fig. 1.4.
Figure 1.3: Example of TC structure from a numerically simulated TC: (a) Horizontal map of 10-m wind speeds. (b) Horizontal map of rain rate. (c) Vertical cross section of radar reflectivity (shading), equivalent potential temperature (red contour), and radial and vertical winds (black arrows) along the thick black arrow in (a) and (b).
Figure 1.3 illustrates the typical dynamic structure (winds) of TCs. The curved vectors and wind barbs in Figs. 1.3a and b depict the strong surface winds spiraling toward the storm center, which is one of the unique features TCs have. The surface wind reaches its maximum in the eyewall, at the radius of maximum wind (RMW). From the RMW, surface winds decrease inward rapidly and outward more gradually. As the primary circulation results in the tangential (circular) component of winds, the secondary circulation results in the radial (inward) component. Such radial inflow is a consequence of frictional retardation and the resulting gradient wind imbalance, and is called frictional inflow. In addition to the frictional inflow, secondary circulation contains also the eyewall updraft and the tropopause outflow because of the diabatic heating in the eyewall (Fig. 1.3c). In between the BL and tropopause, there is a weak middle level inflow due to the convective heating in the eyewall. In the rainbands, the convective-scale updrafts and downdrafts lead to the import of middle-level air into the BL. Below the stratiform precipitation (bluish area in Fig. 1.3b), mesoscale descent also affects the BL by altering the properties of the air that will entrain into the top of the BL.

The thermodynamic structure of a mature TC is illustrated with a vertical-cross section of equivalent temperature ($\theta_e$) in Fig. 1.3c. $\theta_e$ is chosen because it combines both aspects of thermodynamics (temperature and moisture) and represents the heat energy contained in the air flow. $\theta_e$ in general has a minimum in the mid troposphere, around 5 km (this number might vary a little in various cases). Near the surface, $\theta_e$ increases gradually toward the storm center in response to the surface fluxes acquired by the frictional inflow. Near the eyewall, the entire profile of $\theta_e$ shifts to higher values, and the mid tropospheric minimum is nearly eliminated as moisture content increases. This is a consequence of the redistribution of surface fluxes by convective updrafts that lift up high $\theta_e$ air.

It is the redistribution of the surface fluxes that brings in the important role of air-sea interaction in TCs. A detailed discussion of air-sea interaction in the literature will be given in the next section; here we use a simple schematic to illustrate the current understanding...
Figure 1.4: A conceptual model of air-sea interaction in a TC. This is a vertical cross section from the storm center to the outer rainband, as indicated by the red arrow in Fig. 1.2. Black, purple, and orange arrows indicate the direction of airflow at the selected locations. The thick black arrows represent the secondary circulation, which includes the frictional inflow and the eyewall updrafts. $\theta_e$ along the frictional inflow increases with decreasing radius ($\theta_{e3} > \theta_{e2} > \theta_{e1}$) and then it remains constant along the eyewall updrafts. The thin black arrows indicate the convective updrafts and downdrafts in rainbands. The purple and orange arrows indicate the mid-level inflow and upper level outflow. The dashed green line represents the top of the HBL. Below the vertical cross section is the SST, where the colder color (bluish) represents colder SST; this is the location of storm induced SST cooling. At the air-sea interface, there are upward heat fluxes and downward momentum fluxes depicted by the red and blue arrows, respectively. The size of arrows indicates the magnitude of these fluxes.
of air-sea interaction in TCs. The strong surface winds produced the upward enthalpy fluxes (from \(\sim 100 \text{ W m}^{-2}\) in the outer core to \(\sim 1000 \text{ W m}^{-2}\) near the eyewall). Because of the upward air-sea enthalpy fluxes, the BL air gains energy from the ocean and the frictional inflow brings it into the eyewall. Eventually, this heat energy is released through condensation in the eyewall updraft to intensify and maintain the strength of TCs. Strong winds also induce downward momentum fluxes (from \(1 - 5 \text{ N m}^{-2}\) in the outer core to \(\sim 10 - 15 \text{ N m}^{-2}\) near the eyewall), which results in SST cooling, ocean waves and sea spray. The SST cooling feeds back to the storm intensity negatively, by reducing the surface heat fluxes. The ocean waves increase the surface roughness, which further increase the downward momentum fluxes from atmosphere to ocean. This process feeds back to TC as well. 

In the past two decades, a plethora of studies have focused on the feedback of these oceanic responses (SST and waves) on TC intensity. Although TC intensity is strongly affected by the structure, only limited studies examined how air-sea interaction affects the TC structure. Moreover, although it has been widely accepted that the BL plays a role in connecting the deep convection in rainbands to the air-sea interface, studies about the BL in TCs rarely acknowledge the impact of the storm induced SST cooling and surface waves on BL structures. One important reason why the impact of air-sea interaction on TCs and their BL structures has not been thoroughly studied is the lack of two crucial components: a fully physics coupled air-sea model and a comprehensive data set of coupled air-sea observations. In the past decade, an air-sea coupled model was developed (Chen et al., 2007), offering the possibility to study this issue from the perspective of numerical simulations. Meanwhile, two field programs, the Coupled Boundary Air-Sea Transfer (CBLAST) Hurricane Field program in 2003-2004 and the Impact of Typhoons on the Ocean in the Pacific

\[2\text{This paragraph is a summary from lots of previous studies, such as Malkus and Riehl (1960); Emanuel (1986); Rotunno and Emanuel (1987); Leipper (1967); Shay and Elsberry (1987); Schade and Emanuel (1999); Chen et al. (2007); Wright et al. (2001), etc. The work of each of these studies will be discussed individually in Chapter 2.}\]
(ITOP) in 2010 collected a massive set of air-sea coupled observations, which further provided an opportunity to study this complicated problem.

In short, the insufficient understanding of the air-sea interaction in TCs motivates this Ph.D. study, and the coupled atmosphere-(wave)-ocean model along with CBLAST and ITOP observations will be extensively used to advance our understanding of how air-sea interaction affects boundary layer structure and, in turn, storm structure and intensity.

1.2 Science objectives

The leading question of this study is whether and how air-sea interaction modulates the TC structure. Does air-sea interaction affect storm structure in a simple manner that could be parameterized, or instead in a complex manner that requires a full-physics coupled model framework? Since deep convection in the eyewall and rainbands are connected to the ocean surface through the HBL, this study aims to examine the physical mechanisms by which the HBL connects the upper-ocean and air-sea interface to the storm structure throughout the troposphere.

1.3 Outline

The structure of this study is as follows. In Chapter 2, we thoroughly review the current understanding of air-sea interaction in TCs, along with the relevant BL theory in TCs. In Chapter 3, we first introduce the University of Miami Coupled Model system (UMCM), including an existing fully coupled atmosphere-wave-ocean model (Chen et al., 2007), and a newly developed atmosphere-ocean model. The overview of CBLAST and ITOP field programs, and the observational dataset will be illustrated in the same chapter. Chapters 4, 5, and 6 describe the findings in this study and a short discussion about the impact of the chosen BL parameterization on the results will be given in Chapter 7. The conclusions are summarized in Chapter 8 and the future plan is addressed in Chapter 9.
Chapter 2

Literature Review

2.1 Air-sea interaction in TCs

2.1.1 Theory of TC energetics

The air-sea interaction process has been recognized important for TC energetics in the literature since the 50s. Riehl (1954) proposed that the air-sea exchanged heat fluxes are essential to TCs. Malkus and Riehl (1960) showed that there is elevated $\theta_e$ in the core of TCs via the simulated trajectory of BL inflow from an analytical model. This elevated $\theta_e$ is the transfer of sensible and latent heat from the sea surface to the atmosphere. Therefore, the drop of the central pressure is proportional to the $\theta_e$ in the BL, i.e., the intensification of TCs is determined by the surface heat fluxes.

Following Riehl’s work, Emanuel (1986) developed an air-sea interaction theory for TCs that can be understood as a modified Carnot engine (Fig. 2.1). In the modified Carnot engine, the frictional inflow acquires heat energy from the warm ocean through evaporation cooling (from point A to point B). Then the air turns and flows upward as eyewall updrafts. In this segment, the latent heat is converted into sensible heat due to condensation. In this process, there is only a conversion between these two forms of heat without gaining or losing heat energy. The air then spirals outward and the adiabatic expansion occurs.
Figure 2.1: The energy cycle of a mature hurricane. Air spirals inward close to the sea surface from point A to point B and acquires heat energy from the ocean. Air then ascends in the eyewall from B to C without gaining or losing energy. Between C to D, air loses the heat into the storm environments. Eventually, from D to A, air returns to its starting point. (Adapted from Emanuel, 1986)
Between points C and D, the air loses its energy through radiative cooling. At the mean time, air gradually sinks down toward the tropopause. The air returns back to the TC ambient and to the sea surface from point D to A without gaining or losing energy. The kinetic energy produced through the latent heat release between points B and C strengthens the surface winds, which injects more heat from the ocean. This becomes an energy cycle, which positively feeds back to storm intensity. Note that Emanuel’s theory was derived from a highly idealized axisymmetric analytical model, which does not include some of the complex physical processes occurring in reality. Nevertheless, the Carnot engine theory did give insights of the role of air-sea interaction in TC intensification.

Emanuel (1986) suggested also that surface exchange fluxes at the RMW are important for balance between heat fluxes and drag in a TC, while advection of surface heat fluxes from the area outside of twice of the RMW does not play an important role. He made such argument based on the radially near-constant $\theta_e$ beyond twice of the RMW. However, one of his assumptions is that the relative humidity is constant outside of twice the RMW, and this limits the ability of that analytical model to describe the balance between advection, fluxes, etc in the outer region (outside of twice of RMW).

Rotunno and Emanuel (1987) constructed a non-hydrostatic axisymmetric numerical model and confirmed the crucial importance of the air-sea interaction in controlling the evolution of a storm as suggested in Emanuel (1986). They also examined the radial distribution of $\theta_e$ in the BL. For a storm with a RMW of 40 km, $\theta_e$ hardly increases beyond 240 km in radius. This is somewhat matches Emanuel (1986)’s idea of and “unimportant outer region”, but with a radius much larger than twice the RMW. By examining the $\theta_e$ budget in BL, Rotunno and Emanuel (1987) exhibited that the advection of entropy is gradually gaining importance from 240 km to the RMW. A recent work of Wang and Xu (2010) also showed that the energy gained by BL inflow air due to surface enthalpy fluxes outside of and prior to the eyewall contributes significantly to the energy balance in the eyewall through the lateral inward energy fluxes.
Figure 2.2: (a) SST around the track of Hurricane Tess (1975). Tess moved toward the northwest at ~ 6 m s$^{-1}$. The SST minimum occurs 50 – 150 km to the right of the hurricane track. (b) Temperature along section AB in (a). The base of the ML is shown as a heavy dashed contour. The 200 km wide and 40 m amplitude upwelling beneath the track is the response to the positive stress curl of the hurricane. (From Price 1981)

The energetic impact of air-sea interaction on TCs was summarized by three sentences in Rotunno and Emanuel (1987): [It is a form of air-sea interaction instability through which the wind–induced latent heat fluxes from the ocean lead to locally enhanced values of $\theta_e$ in the boundary layer which, after being redistributed upward along angular momentum surface, leads to temperature perturbations aloft. These temperature perturbations enhance the storm’s circulation, which further increase the wind-induced surface fluxes, and so on. The TC will continue to intensify as long as this BL process permits steadily increasing values of $\theta_e$ near the core or until the BL air becomes saturated.]

An implication from this is that any physical process that affects the gain or loss of the inflow $\theta_e$ would ultimately impact TC intensity. One of them is the feedback of the storm induced ocean response as discussed in the next section.
2.1.2 Feedback of TC induced ocean response

In the 1960s, observational studies (e.g. Leipper 1967) found that the downward momentum fluxes in TCs cause cooling in the warm upper ocean layers and SST. Fedorov et al. (1979) and Pudov et al. (1978) showed that such cooling is spatially inhomogeneous with the rightward bias, cold wake, in Hurricane Ella and Hurricane Tess respectively (Fig. 2.2). The strength of the SST cooling depends on several air-sea environment parameters and TC conditions including the surface wind speed, the mixed layer (ML) depth, the temperature gradient below the ML, the translation speed of a storm, and storm size. In general, the strength of storm-induced cooling reaches 1 to 4 °C (Leipper and Volgenau 1972; Price 1981; Zedler et al. 2002; D’Asaro et al. 2007) while for some extreme cases, it can be as strong as 6 °C (Lin et al. 2003). Price (1981) found that the vertical mixing, caused by the wind-driven current shear, is responsible for about 85% of the cooling. The upwelling, a consequence of divergence and convergence, is responsible for enhancing the cooling. He also pointed out that the cold wake is a consequence of the alignment of directions between surface wind and inertial current on the right side of the tracks during most of the storm passages.

Numerical simulations of both idealized case and real cases from a coupled atmospheric-ocean model have been able to simulate the strength and rightward bias of the storm-induced cooling reasonably. For example, Fig. 2.3 shows the resemblance between the simulated SST cooling and the satellite-observed one of Hurricane Frances (2004) from Chen et al. (2012a). While the enthalpy fluxes from the ocean to the atmosphere are the main energy source for TCs, the reduction of SST is expected to have a negative feedback to TC intensity via reducing the upward enthalpy fluxes.

Chang and Anthes (1979) used an axisymmetric coupled hurricane-ocean model to show the weakening of a hurricane due to the cooling of the sea surface. Sutyrin and Khain (1984) coupled an axisymmetric hurricane model to a 3-dimentional ocean model.
Figure 2.3: (a) Spatial variation of SST after Hurricane France (2004) passage from Geo-stationary Satellite (GOES). Blue line shows the observed storm track. (b) Similar to (a) but from high resolution fully coupled model and magenta line shows the simulated storm track. (From Chen et al., 2012b)
and showed that smaller storm translation speeds and smaller initial ML depths lead to stronger negative feedback effect. In the 1990s, air-sea coupling and its impact on TC intensity became a popular research topic. Schade and Emanuel (1999) used an idealized axisymmetric coupled model to exhibit that the oceanic negative feedback has a first order impact on TC intensity, and is not negligible. The strength of the SST feedback depends not only on the large-scale SST pattern, but also on the synoptic-scale subsurface ocean conditions, such as ocean ML depth and the lapse rate of the ocean temperature. Bender et al. (1993) and Bender and Ginis (2000) used a high resolution (for that time), fully physics coupled hurricane-ocean model showing that introducing ocean coupling significantly improves the TC intensity forecast. Then, Lin et al. (2005) and Wu et al. (2007) addressed the insulation effect of the pre-existing warm ocean eddies which mitigate the oceanic negative feedback from observational and numerical study respectively. In this regard, the warm ocean eddies could be seen as a “booster” for some category-5 TCs.

In addition to the TC intensity, storm-induced ocean cooling could also modulate TC structure. By using an axisymmetric coupled atmosphere-ocean model, Anthes and Chang (1978) first documented the influence of air-sea interaction on BL structure from the perspective of numerical simulations. Their results conclude that colder SST results in weaker thermal structures. This change is especially significant at large radii where the wind-driven mechanical mixing and vertical motion are weak, and the BL thermodynamics is mainly controlled by the ocean temperature. About 15 years later, a observational study of Black and Holland (1992) also found the asymmetries in the BL thermal structure that is induced by the cold wake in TC Kerry (1979). Based on a coupled model simulation of Hurricane Katrina (2005), Chen et al. (2010) showed storm induced SST cooling results in the change of the thermal fields and convergence patterns in the simulated storms. The later changes the convective organization because it is related to the developing convection. The results from Black and Holland (1992); Chen et al. (2010) are related to the increase of BL stability over the cold wake; this will be further discussed in Section 2.2.2. Other
than the BL structure, ocean coupling also increases the inner core asymmetry, decreases TC circulation size by 10%, broadens the eye, and reduces the radius of hurricane-force wind (Zhu et al. 2004; Chen et al. 2010).

2.1.3 Impact of surface waves on TC intensity and structure

The downward momentum fluxes induce also surface waves. With NASA airborne Scanning Radar Altimeter (SRA) carried aboard on NOAA WP-3D hurricane hunter, Wright et al. (2001) showed that the wave field is not symmetrically distributed (Fig. 2.4) in Hurricane Bonnie (1998). Older waves with higher wave height and longer wave length were observed in the front-right quadrant, while younger waves with lower wave height and shorter wave length were observed in the rear-left quadrant. Chen et al. (2012b) found a
Figure 2.5: (a) Simulated significant wave height (color, m) and mean wave propagation direction (white vector) and (b) mean wavelength (color, m) and surface wind (black vectors) for Hurricane Frances (2004) from high resolution fully coupled model. The black “+” indicates the storm center and the arrow in the lower left corner indicates the direction of the storm motion. (From Chen et al., 2012a)
similar asymmetry in a numerical simulation of Hurricane Frances (2004) (Fig. 2.5). They explained that the asymmetries in the wave fields are partially due to the stronger winds in the front-right quadrant because of the storm motion. Another cause of wave asymmetry is the angle between wind and wave propagation direction. In the front-right (rear-left) quadrant, wave propagate in the direction that aligns (mis-aligns) with the wind direction.

Although the surface momentum flux depends on both sea state and surface wind field, it has been parameterized as a function of wind speed only in the atmospheric model. The first approach of parameterized wave effect was conducted by Charnock (1955), in which the sea surface roughness, hence the drag coefficient, increases with the wind speed. With airborne dropsonde data, Powell et al. (2003) found that the surface momentum fluxes saturate as the wind speed increases above the hurricane force. A laboratory study of Donelan et al. (2004) also suggested the surface roughness reaches its saturation point when the wind speed is large than 33 m s\(^{-1}\) (Fig. 2.6). Beyond this wind speed, the airflow sepa-
rates from surface waves and is not able to follow the crests and troughs of waves. Some hurricane models, such as Weather and Research Forecast Model (WRF), then use surface drag-coefficient parameterization derived from Donelan et al. (2004). Nevertheless, it still treats momentum fluxes as a function of wind speed only (Davis et al. 2008), regardless an asymmetries in the sea state. In this regard, Chen et al. (2012a) further demonstrated the variability of wind-wave relationship in different quadrants within a hurricane with and without wave coupling effect (Fig. 2.6). Their results suggested the asymmetric sea state in Fig. 2.5 results in an asymmetry in the surface drag coefficient, and therefore the asymmetry in surface momentum fluxes.

Unlike that of ocean coupling, the impact of wave coupling on TC intensity is unclear. The Maximum Wind Speed (MWS) usually increases with wave coupling because of the saturation of downward momentum fluxes in the high wind regime (>33 m s\(^{-1}\)) (Chen et al. 2012a; Liu et al. 2011). The Minimum Sea Level Pressure (MSLP), however, has no systematic tendency with wave coupling. Doyle (1995) and Doyle (2002) suggested that MSLP varies from 8 hPa deeper to 3 hPa shallower with wave coupling from one case to another. The results from Chen et al. (2012a) show that MSLP with wave coupling is about 2 hPa shallower than the one without.

The impact of wave coupling on TC structure is also unclear. Doyle (1995) used a fully coupled atmospheric-wave-ocean model with an idealized cyclone to show that a nonlinear interaction among the BL frictional affects the ocean wave growth and the surface exchange fluxes. He concluded that the ocean waves affect the BL convergence and therefore the rainfall in TCs. Recently, Chen et al. (2012a) showed that the observed concentric eyewall can only be simulated when both ocean and wave coupling processes are considered.
Figure 2.7: The spatial variation of the simulated drag coefficient in Hurricane Frances from (a) fully coupled model (b) uncoupled model. The black lines indicate the four quadrants relative to the storm motion as shown by the black vectors. (c) and (d) are the corresponding scatter plots of the drag coefficient and enthalpy exchange coefficients. Data from the four quadrants are shown in different color as indicated at the top right corner. (From Chen et al., 2012a)
2.2 Atmospheric boundary layer in TCs

The atmospheric BL is the layer that is directly influenced by the Earth’s surface (Stull 1988). The same concept is valid for the boundary layer in TCs. However, the BL in TCs has characteristics that are very different to those in general atmospheric BLs because the strong rotational winds and the organized convection. The BL in TCs is important as it controls the radial distribution of moisture, vertical motion and absolute angular momentum. It also connects the deep convection in the eyewall and rainbands to the ocean surface. For convenience, the BL in TCs is referred to as TC BL (TCBL) or Hurricane BL (HBL) hereafter.

2.2.1 Conventional definitions and the characteristics of the boundary layer depth

Historically, it has never been easy to define precisely what the atmospheric BL is, and the definition of atmospheric BL varies with applications. Similarly, the definition of boundary layer in TCs has been a complex issue and several conventional definitions can be found in the literature. The first commonly used definition is based on one of the atmospheric BL definition in which the height of the HBL is the level where the virtual potential temperature \( \theta_v \) is 0.5 K higher than the surface value (e.g. Anthes and Chang 1978; Powell 1990). It represents both a well-mixed layer and a transition layer aloft in which \( \theta_v \) increases quickly with height. In other words, the well-mixed layer is a base of the inversion or stable layer. The depth of the mixed layer is controlled by the surface heat fluxes, the entrainment from the top of the mixed layer, and the tropospheric subsidence. Observational studies (e.g. Zhang et al. 2011) have shown that the mixed layer depth increases from about 200 m in the inner core to about 500 m at the outer region of hurricanes (Fig. 2.8). The HBL definition is based on the mixed layer concerns with the thermodynamic property of the HBL, which will be referred to as the thermodynamic HBL (THBL) in this study.
Figure 2.8: Vertical cross section of the composite differences in the virtual potential temperature (K) between each level and the near-surface value. The thick black lines denotes the constant contour with a value of 0.5 K. (From Zhang et al. 2011)
Figure 2.9: (a) Vertical cross section of the axisymmetric inflow layer. Arrows are wind vectors scaled in the same proportions as the axes. Heavy lines are the unsmoothed and smoothed tops of the inflow layer. The solid lines are the contours with constant absolute angular momentum ($X10^5 \text{ m}^2\text{s}^{-1}$). Dashed lines are for constant equivalent potential temperature, $\theta_e$. (From Frank 1984) (b) Vertical cross section of composite radial wind velocity. The white dashed line is the height of inflow layer while the back thick line is the height where radial wind speed is 10% of the peak inflow. (From Zhang et al. 2011)
The second definition is based on an inflow layer associated with the secondary circulation in hurricanes. The boundary inflow is a result of the gradient-wind imbalance due to surface friction (Smith 1968) and the top of the inflow layer is where the inflow vanishes (e.g. Smith et al. 2009; Zhang et al. 2011). This definition makes the implicit assumption that the inflow in hurricanes is only induced by the surface friction. However, the inflow is caused by both surface friction and latent heat release from deep convection in the inner core and rainbands of a hurricane in the real world. Although the former dominates near the surface, the latter contributes throughout the lower troposphere (e.g. Pendergrass and Willoughby 2009). Therefore, it is ambiguous when using inflow layer as a definition of the HBL. Nevertheless, to distinguish the inflow layer from that of the mixed layer, here we refer to it as Dynamic HBL (DHBL).

A number of numerical and observational studies have shown the characteristics of the inflow layer. From aircraft data of Hurricane Frederic (1979), Frank (1984) concluded that the inflow layer decreases with decreasing radius from 2.2 km at 135 km radius to 0 km at the storm center while the strongest gradient occurs when the radius is smaller than 100 km (Fig. 2.9a). Using a composite of dropsondes from multiple hurricanes, Zhang et al. (2011) also showed the azimuthally-averaged DHBL increases outward from a few hundred meters in the inner core (~ RMW) to about 1.5 km in the outer region (~4-5 RMW) (Fig. 2.9b). These results are both similar to an idealized hurricane-like vortex study by Montgomery et al. (2001). Just above the DHBL, a layer of outflow due to the upward transport of angular momentum, as explained in Kepert (2001), is expected. However, the studies of Kepert(2006a, b) and Schwendike and Kepert (2008) using dropsondes analysis from several hurricanes show a significant asymmetry in winds and the depth of the DHBL. Their results indicated a large variability in the depth of inflow around each storm. Many have a much deeper inflow layer in parts of the hurricanes than the composite in Zhang et al. (2011), which raises a question whether the composite inflow can present the true structure in hurricanes.
Figure 2.10: Schematic diagram of the characteristics height scales of HBL based on the composite analysis of the dropsondes data. The $h_{\text{infl}}$ is the inflow layer depth (red dashed line), $z_i$ is the mixed layer depth (green dashed –dotted line) and $h_{\text{vmax}}$ is the height of the maximum wind speed. The solid black line represents the height where the bulk Richardson number is equal to 0.25. (From Zhang et al. 2011)
The third definition is related to the Ekman layer. The Ekman layer depth ($\sqrt{\frac{2K}{f}}$) is proportional to the square root of the turbulent diffusivity ($K$) and inversely proportional to the square root of Coriolis parameter ($f$). However, in hurricanes, the relative vorticity is high and therefore the inertial and curvature effects have to be considered in the definition. Rosenthal (1862) was the first to use this definition in hurricanes and found that the Ekman-like layer decreases in depth toward the center of a hurricane. Kepert (2001) and Foster (2009) show that the scale of the HBL height is given by an Ekman-like BL with the Coriolis parameter replaced by inertial stability ($I$) ($\sqrt{\frac{2KI}{f}}$). Here we refer to this definition as IHBL. Maybe because $K$ is usually unknown or has a high uncertainty, there is no direct observational study on the IHBL structure. However, Kepert and Wang (2001) has shown that the depth of the IHBL decreases toward the center of a hurricane due to the increase of inertial stability, which is consistent with the radial variation in the HBL height as determined by other observed quantities. Beyond these three definitions, a few additional definitions of the HBL can also be found in literature, such as the height of maximum wind in Bryan and Rotunno (2009) and the height where the Bulk Richardson number is equal to the critical Richardson number (e.g., Hanna 1969; Wetzel 1982; Mahrt 1981; Troen and Mahrt 1986). Zhang et al. (2011) concluded that except the HBL height defined based on Bulk Richardson number, HBL defined by all other definitions increase with increasing radius. The DHBL is the highest one, almost twice as high as THBL. The HBL that is derived from the maximum wind speed is a little bit shallower than the DHBL (Fig. 2.10).

It is obvious that no matter which definition is used, the HBL is always associated with the amount of inward and upward transport of net heat, moisture and momentum fluxes (Emanuel 1986). The HBL depth is therefore important because the net flux is the integral of flux divergence through the HBL, i.e., the flux difference between the top and bottom of HBL. There may not be a single value or a two-dimensional (2D) radial profile that can adequately describe the three-dimensional (3D) HBL structure. As indicated by Kepert(2006a, b) and Schwendike and Kepert (2008), the HBL is not symmetrically dis-
tributed. The azimuthally-averaged characteristics can not present the characteristics of a whole HBL. Factors controlling the symmetric and asymmetric structures of the HBL are complex and diverse. For instance, air-sea interaction and deep convection in real hurricanes can both contribute to the variability in the HBL structure. However, they have not been systematically studied in this context.

Powell (1990) examined characteristics of the THBL in rainband regions using aircraft data. He showed that the convective downdraft can modulate the THBL by bring down the dry air into the THBL, which vary spatially around the storm. Storm motion can induce asymmetries in the near surface winds and in the DHBL by an enhanced convergence in the front of a hurricane as shown in a depth-averaged slab model by Shapiro (1983). Kepert (2001) and Kepert and Wang (2001) later confirmed this result using both a linear analytical model and a non-linear multi-layers 3-dimensional numerical model, respectively. They also found an appreciable asymmetric component at the eyewall and outer regions in TCs. In particular, the asymmetric component induced by the storm motion dominates the wind field, which results in a wavenumber 1 asymmetry with the maximum storm-relative inflow in the front–right quadrant and the tangential wind in the front-left quadrant.

Figure 2.11: Mean profiles of storm-relative winds from dropsondes between 40- to 100-km radius for Hurricane George (1998). Each group contains averages over four quadrants and over the whole storm. (From Kepert 2006a)
Using GPS dropsondes from Hurricanes George (1998), Mitch (1998), Danielle (1998) and Isabel (2003), Kepert (2006a, b) and Schwendike and Kepert (2008) have further documented asymmetric inflow and DHBL structures that vary from storm to storm. For example, there is a deep inflow in the two rear quadrants near the RMW and in the rear-right quadrant in the outer region in Hurricane George (Fig. 2.11). Unlike the azimuthally averaged fields shown in Zhang et al. (2011), the low-level outflow layer above the DHBL exists only in the front-left quadrant in Hurricane George (Kepert 2006a). A similar deep inflow layer was also found in Hurricane Mitch in the two left quadrants (Kepert 2006b). One of the differences is that Hurricane Mitch was relatively slow moving in comparison with Hurricane George. Schwendike and Kepert (2008) found no outflow in the observed mean profiles from the inner core to the outer region in Hurricane Danielle, whereas there is deeper inflow capped by a layer of outflow near the eyewall in Hurricane Isabel. The depth and strength of the inflow and outflow in hurricanes may also be affected by moist convective heating induced imbalance that is not included in the boundary layer model used in Kepert and Wang (2001).

2.2.2 Boundary layer stability

Underneath the convective weather phenomena, such as TCs or tropical convection, BL is usually expected/assumed to be neutral to unstable. However, the storm induced cold wake results in the advection of the warmer air over a cooler water surface, which increases the BL stability. Some observational studies even suggest that there is a stable BL (SBL) over the cold wake in TCs. Powell (1982), who might be the first one suggesting a SBL in TCs, mentioned that “The larger shear in the rear side of the storm is probably caused by the stable condition over the cold wake of the storm.” Black and Holland (1992) then showed a SBL in TC Kerry (1997) from observational data from a research aircraft, ship, and automatic weather stations. They further found that the SBL had developed over the cold wake and extended downstream of it. The SBL also results in the low-level $\theta_e$ minimum
and low level jet. Lin et al. (2003) is another observational study that suggests an increased BL stability over the cold wake. However, their finding is based on a satellite-measured surface wind field a week after the passage of Typhoon Kai-Tak (2000), during which the environment was no longer TC influence.

While the SBL in TCs and its impact on the TC structure has rarely been discussed in the literature, there is a plethora of studies about the SBL over the mid-latitude cold SST front. Over the cold SST front, a SBL forms because the warmer air over colder water surface, similar to the mechanism responsible for the cold wake. A review study of Small et al. (2008) concluded that the responses of the airflow in the atmospheric BL to the cold SST front are the change of the surface and BL wind direction and the weakening of the wind speed. These changes are because of 1) the development of the internal SBL that is usually shallower than the surrounding environmental BL, 2) the decreases of the turbulent mixing and the downward transferred momentum fluxes from the top of the BL due to the formation of the SBL 3) the change of surface fluxes due to the SST change, and 4) the formation of the hydrostatic pressure anomalies (Wallace et al. 1998; Small et al. 2008 and the references cited therein). To understand which one is the most important factor, Spall (2007) analyzed the momentum budget through a series of idealized, 2-dimensional coupled model simulations. His results showed that under “high” wind conditions (~15 m s$^{-1}$, which is not high in TCs but is high in mid-latitude), the main cause of the wind reduction is the decrease of the vertical mixing of the momentum across the front. Then the wind turns following the background pressure gradient due to the momentum imbalance. The change of surface fluxes and the hydrostatic pressure gradient play minor roles because of the smaller adjustment period under high wind conditions.

Borrowing the idea of the SBL over the mid-latitude SST front (Small et al. 2008), Chen et al. (2010) suggested that the near surface airflow would respond to the cold wake in the same way as the BL flow respond to the cold SST front, i.e., a further inward turning of the surface winds due to the wind reduction. Nevertheless, Chen et al. (2010) proved this
idea based on an idealized storm in an uncoupled model with a constant patch of cold SST. Hence, whether a SBL can be observed in reality and/or simulated in full physics coupled model is still unknown.
Chapter 3

Methodology

3.1 University of Miami Coupled Model System (UMCM)

The University of Miami Coupled Atmosphere-Wave-Ocean Modeling system (UMCM, Fig. 3.1) includes three model components, the atmospheric, surface wave, and ocean circulation models. With various option of each component, UMCM can be configured as

1. UMCM-MWP: coupled with the Fifth-Generation Penn State University/National Center for Atmospheric Research non-hydrostatic mesoscale model (MM5) (Dudhia 1993; Grell et al. 1994), WAVESWATCH III (Tolman 1991, 1999; Tolman et al. 2002), and the three-dimensional Price-Weller-Pinkel (3DPWP) upper ocean model (Price 1981; Price et al. 1994). This coupled model is developed by Chen et al. 2007.

2. UMCM-WMH: coupled with the Weather Research and Forecasting (WRF, Skamarock et al., 2008) model, the University of Miami Wave Model (UMWM, Donelan et al. 2012), and the Hybrid Coordinate Ocean Model (HYCOM).

3. UMCM-WP, coupled with WRF and 3DPWP.

The basic coupling parameters, the data passed between models,, are noted in Fig. 3.1. In this study, UMCM-MWP and UMCM-WP are used. The detailed descriptions of these configurations are the followings.
Figure 3.1: Schematic of University of Miami Coupled atmosphere-wave-ocean Model (UMCM) system and the options for atmospheric, surface wave, and ocean circulation models. The coupling parameters between each of the components are printed next to the black arrows. (From Chen et al., 2012a)

3.1.1 UMCM-MWP

The atmospheric component of UMCM-MWP is a multi-nested MM5 (Dudhia 1993; Grell et al. 1994) with 45-, 15-, 5-, and 1.67-km grid resolutions. The three inner nests are vortex-following moving grids (Tenerelli and Chen 2001). There are 28 vertical levels with 9 of them in the lowest 1 km (approximately at 11, 50, 125, 230, 350, 490, 625, 780, and 950 m). We use Tao and Simpson (1993) microphysics scheme and a slightly modified Kain-Fritsch (Kain and Fritsch 1993) cumulus parameterization on the 45- and 15-km grids, and the same microphysics scheme only on the two inner-most domains. The Blackadar boundary layer scheme (Zhang and Anthes 1982) is used here with a modification of the thermal exchange coefficient over the ocean based on Garratt (1992). Garratt’s parameterization introduces different roughness scales for temperature \( z_t \) and moisture \( z_q \) which is different than the roughness length for momentum, \( z_o \). The surface roughness scale is an essential term to calculate the surface exchange fluxes based on bulk formula. For the uncoupled
MM5 and coupled MM5-3DPWP applications, the momentum roughness length over the open ocean is calculated from the Charnock’s relationship (Charnock 1955). In the fully coupled UMCM-MWP, stress is explicitly computed in vector form from the wave stress using the 2D wave spectra plus the skin drag (Chen et al. 2012a). The turbulent closure is first-order and approximated by the K-theory.

The 3DPWP (Price 1981; Price et al. 1994) is a multi-layers upper ocean circulation model with three-dimensional physical processes including vertical mixing, horizontal advection, vertical advection, and pressure gradient. The horizontal resolution of 3DPWP is 10 km. There are 30 layers in 3DPWP with resolutions varying from 5 m in the mixed layer to 20 m below down to 390 m depth. The model does not have bathymetry.

WW3 version 1.18 is used to simulate ocean surface waves in UMCM-MWP. It was developed by Tolman(1991, 1999) for wind waves in slowly varying, unsteady and inhomogeneous ocean depths and currents, and had been evaluated extensively and validated with observations (Tolman et al. 2002). The wind waves are described by the action density wave spectrum \( N(k, \theta, x, y, t) \). In this study, we use 25 frequency bands, logarithmically spaced from 0.0418 to 0.41 Hz at intervals of \( \frac{\Delta f}{f} = 0.1 \) and 48 directional bands (7.5° interval). The WW3 model domain is set to be about the same as the outer domain of MM5. The grid spacing is 1/6° in both the latitudinal and longitudinal directions. The water depth data used in the wave model is the 5’ gridded elevation data from the National Geophysical Data Center.

These three components communicate with each other through a coupler at each coupler time step, 10 minutes. In between the coupler time steps, the exchanged variables for each component are set constant respectively and each component is integrated forward.

### 3.1.2 UMCM-WP

In order to capitalize on the development of WRF, UMCM-WP is developed in collaboration with the scientists in the National Center for Atmospheric Research (NCAR). In
UMCM-WP, 3DPWP is added as a subroutine in the physical package of WRF. This design allows WRF and 3DPWP to share the same horizontal resolution and to communicate with each other at every time step in all domains. In other words, there is no horizontal interpolation and no time-lag for WRF to exchange information with 3DPWP. This is very different from the coupled modeling system in which the two components communicate to each other through a coupler while all the exchange variables are set to be constant between every coupler time step (eg. UMCM-MWP). The atmospheric component, WRF, has two moving nesting domains with 4- and 1.3-km horizontal grid resolutions embedded in a 12-km coarse resolution domain. There are 36 vertical levels with the lowest level at 17 m from the surface and 10 levels in the lowest 1 km. The microphysical scheme used here is WSM5 (Hong et al. 2004), and the boundary layer scheme is YSU (Hong et al. 2006) with wind-dependent surface roughness based on Donelan et al. (2004) and with the wind-dependent surface heat exchange coefficient based on Garratt (1992). The cumulus scheme here is the Kain-Fritsch scheme (Kain and Fritsch 1993) but it is only utilized on the coarse resolution domain. The core area of the simulated storm is entirely contained within the inner most (1.3 km) domain, on which convection can be explicitly simulated. The ocean component, 3DPWP, remains the same as the one in UMCM-MWP but with higher horizontal resolution (followed the grid spacing in WRF). Even though 3DPWP in UMCM-WP is capable to run down to 1.3-km resolution, the ocean field in the finest domain (1.3-km) is linearly-interpolated from its mother domain (4-km) because we do not expect sharp horizontal gradient of ocean field within such fine resolution. As a result, 3DPWP is computationally cheap and does not add any significant computational time compared to the running time of the uncoupled WRF.
3.2 Data

For the purposes of model initialization and verification, this study utilizes airborne in situ data collected during CBLAST (2004) and ITOP (2010), as well as satellite observations. Here, we describe the overall goals of CBLAST and ITOP along with various observations obtained during these two campaigns. We describe also the satellite data used in this study.

3.2.1 Airborne observations

Coupled Boundary Air-Sea Transfer (CBLAST)

The CBLAST field campaign was conducted from 2000 to 2005. Within this period, CBLAST-Hurricane was conducted in 2003 and 2004. The goal of CBLAST-Hurricane is to measure, analyze, understand, and parameterize the air-sea fluxes in the tropical cyclone environment. The objective of CBLAST during the 2004 hurricane season is to conduct
detailed measurements of HBL, air-sea interface and ocean ML using coincident airborne in situ and the remote sensing measurements, together with air-deployed drifting buoys and floats. In addition to these, the airborne turbulence measurements was conducted with the NOAA WP-3D (Fig. 3.2). The airborne in situ measurements include Stepped Frequency Microwave Radiometer (SFMR), Airborne eXpendable BathyThermograph (AXBT), and the GPS dropsondes. SFMR measures surface wind speed while GPS dropsonde measure the vertical structure of atmospheric winds, temperature, and moisture. AXBT measures vertical profiles of ocean temperature. In this study, we use the data collected for Hurricane Frances (2004) from 30 August to 1 September. For convenient, CBLAST-Hurricane is referred to as CBLAST hereafter.
Impact of Typhoons on the Ocean in the Pacific (ITOP)

The ITOP field campaign was conducted during the 2010 Typhoon season, from 20 August to 20 October, to improve the fundamental knowledge of the complex air-sea interaction in TCs. To achieve this goal, ITOP used a variety of experimental approaches to measure typhoons and the oceanic response to them (Fig.3.3). These include in situ observations from the C130 of the 53rd Air-Force Reserve Hurricane Hunter, such as SFMR, AXBTs, and the GPS dropsondes, same as CBLAST. In addition, floats, and drifters were deployed from both the C130 and from US and Taiwanese Research Vessels. Floats measure ocean temperature, conductivity and current while drifters measure not only subsurface ocean temperature, but also surface temperature, winds and pressure. Among all the data, the collocated GPS dropsondes and AXBTs provided simultaneous profiles of atmospheric and oceanic thermal structures in TCs; this was the first TC field program to obtain such air-sea coupled observations. With these collocated soundings, it is possible to estimate the surface fluxes through the bulk algorithms. Another new approach in ITOP is a cold-wake
Figure 3.5: The flight pattern for two of ITOP missions executed at 2200 UTC 16 (orange line) and 2100 UTC 17 September (green line). Symbols indicate the locations of various observations. Squares are GPS dropsondes, stars are AXBTs, triangles are the EM-APEX floats, and circles are the UCSD ADOS drifters.

module (Fig. 3.4), which was designed specifically for increasing the spatial coverage of GPS dropsondes and AXBTs across the storm-induced cold wake.

In this study, we use the data collected from two of ITOP missions for Typhoon Fanapi executed (started) at 2200 UTC 16 (Mission #1) and 2100 UTC 17 September (Mission #2). Figure 3.5 shows the flight paths of each mission and the cold-wake module was excited during Mission #2. The data set includes SFMR, GPS dropsondes, the co-located AXBTs, University of California San Diego CalearSat-Autonomous Drifting Ocean Station (ADOS) drifters, and University of Washington ElectroMagnetic Autonomous Profiling EXplore (EM-APEX) floats. Overall, we have 70 dropsondes from both missions covering the wind speed ranging from 5 to 40 m s\(^{-1}\). There are 34 of them (all from Mission #2) paired with the AXBTs. The criteria for defining the co-located soundings are 1) the distance between the near surface location of the AXBT and the dropsonde is less than 30 km (less than the radius of the mesoscale ocean eddies), and 2) the last dropsonde reporting point is at
a height lower than 15 m. Out of all co-located AXBTs, 26 of them are less than 15 km away from their corresponding dropsondes, 6 of them are 20-25 km and only one of them is 25-30 km. The range of SST of these co-located soundings is from 26 to 29 °C. The EM-APEX floats were deployed during Mission #1 and were deliberately deployed along the predicted storm track with the strategy of being in the eye and 50, 100, and 150 km left and right side of the storm center. ADOS drifters were also deployed from right to left across the predicted storm track during Mission#1.

3.2.2 Satellite observations

Tropical Rainfall Measurement Mission Microwave Imager/Advanced Microwave Scanning Radiometer-EOS (TRMM TMI/AMSRE) Optimally Interpolated (OI) satellite SST is used for the ocean model (3DPWP) initialization. TMI/AMSRE OI consists of 0.25 × 0.25 degree grids data representing a daily minimum SST defined to occur at approximately 8 AM, local time. To verify the atmospheric model, TMI estimated rain rate and the Morphed Integrated Microwave Imagery at CIMMS (MIMIC, Wimmers and Velden, 2007) are also used. The MIMIC product is a synthetic blend of tropical cyclone imagery from several low-Earth orbiting satellite instruments: the Defense Meteorological Satellite Program (DMSP)-13/14/15 SSM/I (85 GHz channel), the TRMM TMI (89 GHz channel) and the Aqua AMSR-E (85 GHz (A) channel). Because the signal from these channels is strongly attenuated from hydrometers generated by deep convection, the imagery is simply used as a proxy for the horizontal distribution of precipitation.

Note that sometimes it is convenient to analyze the data in the storm-relative coordinate and convert the observed winds into tangential and radial wind components. For this type of analyses, the storm position produced by NOAA/HRD based on the flight-level wind data (Willoughby and Chelmow, 1982) will be used. Also, in this study, satellite data are used for qualitative comparison while the in situ data are used for both qualitative and quantitative analyses.
3.3 Forward Lagrangian trajectory and tracer analysis

To understand the evolution of the air in the BL, two methodologies are used here: one is using Lagrangian trajectory particles and the other is using tracers. A Lagrangian trajectory is a single air-particle that is integrated forward with time, following the airflow, and only advection is considered:

\[ x_{i,t+1} = x_{i,t} + \delta x_{i,t} \]

\[ \delta x_{i,t} = U_{i,t} \times \delta t \]  

(3.1)

\( x_{i,t} \) represents the location in the \( i \)-direction at time \( t \), \( \delta x_{i,t} \) is the displacement of the trajectory because of \( U_{i,t} \) over \( \delta t \) period. One important characteristic of the trajectory is that it is an undiluted air parcel, i.e., molecular diffusion and turbulent mixing processes are not considered. By using a passive tracer, Romps and Kuang (2010) found that the undiluted updrafts are negligible above height of 4-5 km, indicating the importance of mixing processes. Additionally, the mixing processes in TCs are essential because the development and the maintenance of the vortex rely on the mixing of air between the vortex, BL, and the environment. Hence, to complete our analyses, a tracer, an air parcel that is subjected to not only advection but also mixing and diffusion, is also used in this study as:

\[ \frac{dC}{dt} = \nu_c \frac{\partial^2 C}{\partial x_i^2} - \frac{\partial (u_i C')}{\partial x_i} + S_c \]  

(3.2)

In Eq. (3.2), \( C \) denotes as a passive tracer and can be separated into mean field and perturbation field \( C' \). \( x_i \) again represents the location in \( i \)-direction, \( u_i \) represent the turbulent wind field in \( i \) direction and \( \nu_c \) is the molecular diffusivity. The terms on the right-hand side are (from left to right) the mean molecular diffusion, the divergence of turbulent tracer flux, and the sink term. Here, the sink term considers only when the tracer moves out from the model domain.
The tracer calculation originally belongs to one of the WRF-chemistry packages. With collaboration of NCAR scientist, we are able to modify WRF and use it as one of the scalar variables without enabling the whole chemistry package. Then the turbulent mixing processes are added in YSU scheme. The trajectory calculation is developed in WRF for in this study. It is written as a function that can be easily switched on and off prior to submit a WRF simulation. For the simulation with trajectory switch on, WRF call trajectory subroutine at the end of each big Runge-Kutta step.

The existence of the moving nests complicates both tracer and trajectory calculations. For the tracer, the entrainment between domains is taken care of through WRF numerical software. For each trajectory particle, we first need to determine how many domains contain it. Then the trajectory function is integrated forward in each of the determined domains, respectively. However, we saved only the information, i.e., \( x_{i,t} \), from the finest-resolution domain. At the next time step, the saved location will be used to decide the number of domains needed again. By doing so, we repeat trajectory calculation couple times when the trajectory is in a nested domain, but this redundant calculation does not cause a significant amount of computational time. Note that the \( x_{i,t} \) is not limited to be an integer, and therefore the location of each trajectory is not constrained to be at the model grid point. Hence, the \( U_{i,t} \) is linearly interpolated onto the trajectory point from the eight surrounding grid points. The grid stagger and map projection of tracer and trajectory is also be solved by adopting WRF functions when converting tracer/trajectory location from grid point to geographic location.

Each tracer is saved as a 3-dimensional array, and trajectories are saved as three 1-dimensional arrays that contain the longitude, latitude, and height. The size of each trajectory array depends on the amount of air-particles released. We design both calculations in the way that we can initiate them at any time during the simulations. The biggest advantage of our implementation of tracer and trajectory calculations is the improved accuracy, that avoids the potential error when using a low frequency model output to backward calculate
trajectories as pointed out by Dahl et al. (2011). Meanwhile, it does not require high frequency massive output, and therefore we can follow the air particles/parcels for a longer period with the limited computational storage space. However, one drawback associated with it is that, backward trajectory could not be calculated.
Chapter 4

Symmetric and Asymmetric Hurricane Boundary Layer Structure in TCs

An overview

The work presented in this section has been published on the Journal of Atmospheric Science (Lee and Chen, 2012). In this study, we study symmetric and asymmetric hurricane boundary layer in TCs through a fully coupled atmosphere-wave-ocean model and observations. Numerical experiments of Hurricane Frances (2004) with and without coupling to an ocean model and/or a surface wave model are used to examine the impacts of these coupling processes. Model results are compared with the airborne dropsonde and surface wind measurements on board of the NOAA WP-3D aircraft.

The atmosphere-ocean coupling reduces the mixed layer depth in the rear-right quadrant due to storm-induced ocean cooling, whereas the wind-wave coupling enhances boundary inflow outside of the radius of maximum wind. Storm motion and deep tropospheric inflow create a significant front-to-back asymmetry in the depth of the inflow layer. These results are consistent with the dropsonde observations. Azimuthally averaged inflow layer and mixed layer as documented in previous studies are not representative of asymmetric HBL. The complex, three-dimensional asymmetric structure in both thermodynamic and dynamic
properties of the HBL induced by atmosphere-wave-ocean coupling indicates difficulty of parameterizing the coupling processes.

4.1 Hurricane Frances (2004)

Hurricane Frances formed from a tropical wave that moved off the west coast on 21 August. It moved westward for several days. Thunderstorms associated with the wave began organizing on 24 August and became a tropical depression. The depression strengthened to tropical storm status on the next day, and was moving west northwestward. It became a named storm on 26 August, and was upgraded to a hurricane late that afternoon. On 27 August, Frances moved more northwesterly due to the upper level trough. It became a category-4 hurricane late on 28 August with maximum wind speed (MWS) of 59 m s$^{-1}$ and minimum sea level pressure (MSLP) of 946 hPa. It then went through an eyewall replacement cycle on 29-30 August while it slowly weakened (Beven 2004). Re-intensification began at about 1200 UTC 30 August, and Frances reached its peak intensity as a category-4 hurricane as it passed north of the Leeward and Virgin Islands at 1800 UTC on August 31, 2004. It had MWS of 63 m s$^{-1}$ and the MSLP of 938 hPa. As the vertical shear remained low, the storm maintains its category-4 hurricane status. Moderate westerly vertical shear developed later on 2 September, and Frances weakened notably during the next two days. It eventually made landfall east-southeast of West Palm Beach, Florida on 5 September as a category-2 hurricane. Its long residence over the open ocean makes it an ideal case for this study. Another advantage of choosing Frances for this study is that there are massive airborne in situ data from the NOAA and the Air Force reconnaissance missions during CBLAST (as addressed in Chapter 3).
4.2 Numerical experiments

Here, we use three experiments conducted in Chen et al. (2012b,a): 1) the uncoupled atmospheric model, MM5 (UA), 2) the coupled atmospheric-ocean model, MM5-3DPWP (AO), and 3) the fully coupled atmospheric-wave-ocean UMCM-MWP (AWO). For MM5, the model domains are $120 \times 120$, $121 \times 121$, $121 \times 121$, and $151 \times 151$ for 45, 15, 5-, and 1.67-km domain, respectively. The simulations start from 27 August to 6 September. The National Center for Environmental Prediction (NCEP) global analysis fields (6-hourly and $1^\circ \times 1^\circ$) is used to initialize the MM5 and provide continuous lateral and lower boundary conditions. The model SST is initialized from NCEP reanalysis SST blended with satellite (TMI/AMSR-E) SST using the method described in Chen et al. (2001). The NCEP reanalysis and TMI/AMSR-E SST is from one day prior to the beginning of the model simulation. Vertical structure of ocean is initialized using observed and climatological profiles. The temperature profile is blended with a pre-storm AXBT observation from the NOAA research aircraft mission and LEVITUS94 climatological data (Levitus et al., 1994; Levitus and Boyer, 1994) for depths greater than sampled by AXBT observation. The salinity profile is from LEVITUS94 climatology, since there is generally no in situ pre-storm observation available. Although the 3DPWP can be initialized using the operational Hybrid Coordinate Ocean Model (HYCOM), the HYCOM fields were not used in this case because of a systemic temperature bias compared with the AXBT data during the time period of Hurricane Frances in August 2004.

4.3 Simulated track and intensity

Figure 4.1a shows the simulated storm tracks in comparison with the NHC best track data. The tracks from UA, AO, and AWO simulations are similar and are close to the best track. The fully coupled AWO produced a track that is the closest to the best track in terms of
Figure 4.1: (a) The NHC best-track (black) and three model simulated tracks of Hurricane Frances (UA-blue, AO-green, and AWO-red) from 27 August to 6 September 2004. (b) Observed and simulated maximum wind speeds from 1200 UTC 30 August to 0000 UTC 01 September during which when airborne observations are used in this study.

timing, whereas UA and AO are 3-6 h slower than the best track after 1 September. The model simulated storm intensities are much more diverse than the tracks. All the simulated storm went through a similar eyewall cycle with a relative minimum intensity at about 0000 UTC 31 August, ~12 h later than the observations (Fig. 4.1b). The focus here is from 1800 UTC 30 August-0000 UTC 1 September when the dropsonde data are available and the storm was away from the land mass. Another advantage of choosing this period is that the model-simulated storm tracks and translation speeds are very close to that of the NHC best track estimates during this time (Fig. 4.1a), which is important for comparing the storm structure and HBL properties with observations. From 0000 UTC 31 August-1 September, the three model simulations show a similar trend of intensification as the observations, but have a significantly difference in the model simulated MWS. Although UA seemingly under-predicts storm intensity in terms of MWS, the MSLP in UA is much lower than the best track data (not shown here). The inconsistency in the pressure-wind
relationship in the uncoupled model is attributed to the lack of coupling to the ocean waves and the unrealistic roughness formulations using the Charnock relationship in UA and AO. Storm-induced SST cooling in AO led to an even weaker MWS than UA as expected. AWO improves the model simulated surface winds significantly due to the wind-wave coupling that reduces surface stress at high wind speeds (Chen et al. 2012a). The question remains as how the air-sea coupling affects the atmospheric boundary layer structure in hurricanes, which may be of importance to model-simulated overall storm structure and intensity.

4.4 Surface winds and air-sea fluxes

4.4.1 Spatial and temporal variations

Although the storm intensity is traditionally estimated by the MWS anywhere in the storm, there is a significant spatial variation of surface winds within each storm at any given time. The spatial distribution of the surface winds is a much better representation of the overall storm structure that not only affects storm evolution and intensity, but is also the most relevant measure of hurricane impact on the ocean. In this section, all the surface wind analyses of model simulations and observations are Earth-relative. Figure 4.2 shows the surface wind, enthalpy (sensible + latent heat) and momentum fluxes from the UA, AO, and AWO simulations at 1800 UTC on 31 August. The model simulated surface winds are verified against the SFMR from the NOAA WP-3D aircraft across the center of the storm from 1650-1800 UTC on 31 August (Fig. 4.3). All three simulations are able to capture the asymmetry in the inner core with a relative minimum in the southwest quadrant (Figs. 4.2a-c and 4.3), whereas the model simulated eyewall are slightly larger than the SFMR measurement. Overall, AO is relatively weaker than UA and AWO. While the modeled wind profiles are broader than the observation, AWO improves the surface wind speed, especially in the outer region.
Figure 4.2: The model simulated surface wind speeds (a-c), enthalpy fluxes (d-f), and momentum fluxes (g-i) from UA, AO, and AWO, averaged over a 2-h period centered at 1800 UTC 31 August. The black arrows indicate the direction of storm motion. The white line in (a)-(c) marks a reconnaissance flight path of the NOAA WP-3D aircraft, where the aircraft measurement will be shown in Fig. 4.3.
Figure 4.3: Surface wind speeds from the Stepped Frequency Microwave Radiometer (SFMR) measurement (black) and three model simulations (UA-blue, AO-green and AWO-red) along the flight path indicated in Fig. 4.2. The SFMR data is collected during the time period from 1650 UTC to 1800 UTC 31 August while the model fields are sampled at 1700 UTC 31 August.
The surface enthalpy fluxes are significantly different among the coupled and uncoupled simulations (Figs. 4.2d-f). The hurricane-induced SST and upper ocean cooling reduces the enthalpy flux, especially in the rear-right quadrant in AO and AWO as compared to that of UA. There are substantial differences in enthalpy fluxes between UA and AO not only in the region of strong cooling, but also near the RMW (~30 km in both simulations), where there is relatively little cooling. This is because the largest reduction of winds in AO occurs at RMW, which has a large effect on the enthalpy flux.

The spatial distributions of momentum fluxes are similar to that of surface wind speeds. The momentum flux in UA is stronger in the eyewall region near the radius of maximum wind (RMW) than AO and AWO (Figs. 4.2g-i). The difference between UA and AO is mostly due to the weaker surface winds because of ocean cooling in AO. However, the wind speed is stronger in AWO than AO, which is due to the difference in the stress formulations with and without the wind-wave coupling. Wind-wave coupling reduces stress at higher wind speed in AWO compared to that of Charnock relationship used in UA and AO (Chen et al. 2012a).

The evolution of surface properties is examined over the period from 1200 UTC 30 August-0000 UTC 1 September. Figure 4.4 shows the time series of storm-averaged SST anomaly (i.e., the difference between SST and its initial value) and air-sea fluxes averaged over an annular area between the radii of 0.5 and 5.0 times of RMW, and azimuthally averaged peak wind speed at RMW. On average, about 0.5 °C SST cooling led to more than 100 W m\(^{-2}\) reduction in latent heat flux (LH) and about 20 W m\(^{-2}\) in sensible heat flux (SH) in AO and AWO. The averaged reduction of LH and SH over the entire 36-h period in AWO (AO) are 60 (80) W m\(^{-2}\) and 10 (15) W m\(^{-2}\), respectively, which resulted a decrease of ~20% in enthalpy flux in AO and ~15% in AWO compared to that in UA.
Figure 4.4: Time series of model simulated (a) sea surface temperature anomaly (°C), (b) azimuthally averaged peak surface wind speed at the radius of maximum wind (RMW), (c) latent heat fluxes (W m\(^{-2}\)), (d) sensible heat fluxes (W m\(^{-2}\)), (e) enthalpy fluxes (W m\(^{-2}\)), and (f) momentum fluxes (N m\(^{-2}\)) averaged over an annular area between radii of 0.5 and 5.0 times of RMW from 1200 UTC 31 August to 0000 UTC 01 September (UA-blue, AO-green, and AWO-red).
Figure 4.5: Azimuthally averaged surface tangential (a) and radial (b) wind speed at 1800 UTC 31 August from the model simulations.
4.4.2 Symmetric and asymmetric structures

Azimuthally averaged fields are used here to represent the symmetric structure. The tangential and radial wind components ($V_t$ and $V_r$) are used to describe the mean vortex structure including the storm size (e.g., the RMW) and the secondary circulation (e.g., inflow and outflow), respectively. The azimuthally averaged $V_t$ and $V_r$ at the surface from all three simulations are shown in Fig. 4.5. The mean $V_t$ is consistent with the surface profiles including the SFMR observation shown in Fig. 4.3. The AO is slightly weaker than UA due to the coupling to the ocean. Comparing AO and AWO, the wind-wave coupling tends to reduce the $V_t$ outside of RMW and increase the peak $V_t$ near RMW (Fig. 4.5a), which corresponding to an increase of radial inflow outside of RMW (Fig. 4.5b). The near surface radial inflow is driven by friction. The increase in radial inflow outside of RMW can be attributed directly to the effect of coupling to the surface waves that reduces (increases) the drag coefficient in high (low) wind speeds compared to the Charnock relationship used in UA and AO (Donelan et al. 2004; Chen et al. 2012a).

The characteristics of the asymmetric component of winds are shown in Fig. 4.6. The storm is divided into four quadrants according the direction of storm motion: front-left (FL), front-right (FR), rear-left (RL), and rear-right (RR). Near the RMW, there is a noticeable decrease in tangential wind in RR quadrant in AO compared to that in UA (Fig. 4.6a), which is corresponding to the storm-induced SST cooling and the large reduction in enthalpy flux (Figs. 4.2d and e). A similar reduction is found in radial wind speed as well (Fig. 4.6b). Wind-wave coupling in AWO increases the mean tangential wind on the right side of the storm more than the left side (Fig. 4.6a). Outside of RMW, the most noticeable difference is in the enhanced inflow in AWO with the largest departure from UA and AO in FL (Fig. 4.6b). The large difference in radial inflow from AWO compared to that of UA and AO is mostly due to the effect of wind-wave coupling that reduces (increases) drag coefficient at higher wind speeds in the inner core (outer region). Figure 4.7 shows the
Figure 4.6: Similar to Fig. 4.5, except averaged fields over each of the four quadrants divided based on the storm forward motion that points to the top (FL: front-left, FR: front-right, RL: rear-left, RR: rear-right).
Figure 4.7: Same as in Fig. 4.6, except for the drag coefficients ($C_d$).

drag coefficients in the four quadrants. AWO has higher value of drag coefficient outside of inner core (> 50 km) but smaller value at RMW compared to UA and AO. There is no significant difference from quadrant to quadrant.

### 4.5 Hurricane boundary layer structure

The GPS dropsonde data from the NOAA WP-3D flights during the CBLAST-Hurricane field campaign in 2004 provides a unique opportunity to examine HBL structure in Hurricane Frances. Here we stratify the dropsonde data into sub-regions in both radial and azimuthal directions around the center of the storm. The locations of all dropsondes deployed from the three WP-3D flights in Hurricane Frances from 30 August-1 September are shown in Fig. 4.8. In order to compare the model simulations with observations, vertical profiles of model fields are sampled at the same storm-relative locations and times. The model simulations are verified accordingly. Both the mean properties of HBL and spatial
Figure 4.8: Storm-relative locations of 34 dropsondes from Hurricane Frances research flights used in this study. The dropsonde data are collected from 30 August to 1 September. The numbers indicate the dropsondes that will be shown individually in Figs. 10 and 11, whereas "*" is dropsondes included only in the mean. The circles indicate radii of 50, 100, 150, and 200 km, respectively. The inner core region within the 50-km radius is enlarged shown at the right. The storm forward motion is pointed to the top.
variability are examined. To investigate the effects of full atmosphere-wave-ocean coupling on HBL, a comprehensive analysis of both the traditional mixed layer property defined by THBL and the inflow layer defined by DHBL will be presented. All the wind analyses are Earth-relative, except Section 4.5.3 in which the influence of the storm motion on HBL is examined in both Earth-relative and storm-relative framework.

4.5.1 Vertical profiles of winds and temperature

DHBL

Azimuthally averaged vertical profiles of $V_t$ and $V_r$ from the dropsondes and the model simulations are shown in Fig. 4.9. Although Frances was not in a steady state, the intensity change from 31 August - 1 September was relatively small (within a 10 m s$^{-1}$ range). Nevertheless, it is possible that the composite could provide a somewhat skewed depiction of the storm structure depended on the data coverage in time and space. In the inner-core region, the maximum $V_t$ occurs near but below the top of the inflow layer, i.e., at a height of ~600 m at RMW (Figs. 4.9a and 4.9d) and ~1000 m at 2RMW (Figs. 4.9b and 4.9e), respectively. An outflow region is right about the inflow layer. These are similar to that described in Kepert (2006aa, b) and Schwendike and Kepert (2008). Although all three simulations have captured the general features of DHBL, AWO is most close to the observation with the best tangential winds and inflow strength, whereas UA overestimated $V_t$ throughout the lower troposphere, which means that UA produced a much stronger vortex than AWO even though the surface winds are relatively close to each other (Fig. 4.9a). However, all model simulations have a much weaker outflow than that of observed at RMW. The outflow is mainly a consequence of the upward transported super-gradient momentum carried by the eyewall updraft (Kepert and Wang 2001). This result indicates that the models may be under-predicting the upward momentum transport. One possible reason could be that the vertical velocity in model is under-predicted due to the coarse vertical resolution. The in-
Figure 4.9: The mean profiles of tangential (a-c) and radial (d-f) winds of all dropsondes shown in Fig. 4.8 at RMW (left), 2 times RMW (middle), and outer region greater than 5 times RMW (right) and azimuthally averaged profiles from UA (blue), AO (green), and AWO (red) simulations at 1800 UTC 31 August.
flow at 2RMW is slightly stronger than that at RMW as shown by both the dropsonde data and the AWO simulation (Fig. 4.9b). In the outer region, the inflow diminishes at ~1500 m at 5RMW, and there is no outflow above the inflow layer (Figs. 4.9c and 4.9f), which is in agreement with Kepert (2006a). One of the differences in model simulated $V_t$ compared to the observations is that model tends to produce a linear, rather than a logarithmic profile near the surface, which may be due to a problem in the implementation of the Blackadar PBL scheme in MM5 as discussed in Kepert (2012).

There is a large spatial variability that deviates from the mean properties of the boundary layer properties from the dropsonde data. Figures 4.10 and 4.11 show $V_t$ and $V_r$ from individual dropsondes around the center of the storm. To fit the limited space, a subset of the dropsondes shown in Fig. 4.8 is used, which covers four quadrants where the data are available. There were more dropsondes on the right side than the left. Overall the tangential wind profiles are similar to the mean, except some do not show the low-level maximum in the inner region (i.e., from the center to 2RMW) as in the mean. There is a left-right asymmetry with the strongest $V_t$ on the left, which is similar to that of model simulations shown in Figs. 4.2a-c.

The spatial variability is more apparent in the radial wind profiles. The inflow layer is the shallowest at ~200-300 m near the RMW close to the center (sondes 10, 31, and 32), while the inflow deepens radially outward within the inner core region to 800 - 1000 m near 2RMW (sonde 12, 16, 18). Some dropsondes display a clear outflow layer about the inflow layer in the inner core (Fig. 4.11). In the outer region, however, the inflow layer is similar or even shallower than that at 2 RMW in the front quadrants (sondes 14 and 15) and much deeper in the rear quadrants where there is no outflow (sonde 19, 20, 28). These features are different than the mean inflow layer that described in Zhang et al. (2011), but is consistent with that in Hurricane George (1998) shown in Kepert (2006a). This asymmetric structure will be discussed further in the next section.
Figure 4.10: Observed profiles of tangential winds from 12 dropsondes inside of the 50-km radius and 9 dropsondes in the outer region. The layout of the dropsondes here are only proxy to what shown in Fig. 4.8.
Figure 4.11: Same as in Fig. 4.10, except for the radial winds (m s\(^{-1}\)). Negative (positive) values are inflow (outflow).
Figure 4.12: Same as in Fig. 4.9, except for virtual potential temperature.

**THBL**

Because THBL is essentially defined by the mixed layer, virtual potential temperature ($\theta_v$) profiles are used to examine the properties of the THBL. The azimuthally averaged and spatial distributions of $\theta_v$ around the storm are shown in Figs. 4.12 and 4.13, respectively, using the same dropsondes as in Figs. 4.9-4.11. The depth of the mixed layer increases from the inner core to outer region, from about 200 m at RMW to 600-700 m at 5RMW on average (Fig. 4.12). The model simulations produced THBL heights similar to that of dropsonde observations. However, the values of $\theta_v$ in models are 2-3 K higher than the observations in the inner core, whereas the values are similar to observations in the outer region. The coupled AO and AWO model simulations were able to reduce the high $\theta_v$ value by 0.5 - 1.0 K, but still higher than the observations in the inner core (Fig. 4.12). It is possible that the SST in the model, initialized from the TRMM/AMSRE satellite data, is too warm. The rear-right quadrant has the shallowest mixed layer (sonde 21 and 22, locations...
Figure 4.13: Same as in Fig. 4.10, except for virtual potential temperature (K). The gray line indicated the THBL calculated from each sounding.
are shown in Fig. 4.8) than other quadrants at 2RMW where the hurricane-induced SST cooling is the most pronounced (Fig. 4.2e).

To further examine the thermodynamic properties of the THBL, the equivalent potential temperature ($\theta_e$) are used to represent moist potential energy of the boundary layer. Again we separate them into the inner and outer core regions (Fig. 4.14). Moreover, the dropsondes located in the rear-right (RR) quadrant are shown separately from other three quadrants (OQ) to examine the influence of the storm-induced SST cooling. The $\theta_e$ is about 363 K in the inner core and 353-355 K in the outer region from the dropsonde data with values in RR 2-3 K generally lower than in OQ, except near the surface in the inner core (Figs. 4.14a-b). The model simulated $\theta_e$ profiles are sampled in the same storm-relative locations as the dropsondes. The results show that the uncoupled UA overestimates the $\theta_e$ by more than 5 K in both inner core and outer regions due to the lack of storm-induced cooling in the upper ocean. The fully coupled AWO produced the $\theta_e$ profiles that are closest to the observations both near the surface and up to 2 km level (Figs. 4.14g-h).

### 4.5.2 Symmetric and asymmetric structures in HBL height

While the symmetric structure of HBL has been documented in axisymmetric models (e.g., Smith 1968, 2003; Kepert 2001) and by azimuthally averaged fields from observations (e.g., Zhang et al. 2011), the asymmetric structures are difficult to examine systematically due to the lack of a full 3D observation of winds and thermodynamic fields in time and space. Given that the fully coupled AWO simulation has been verified well with the dropsonde observations as shown in Figs. 4.9 and 4.12, here we compare the azimuthally averaged THBL and DHBL (Fig. 4.15) with that of four quadrants (Fig. 4.16) from the three model simulations to examine the symmetric and asymmetric HBL structures in Hurricane Frances. The general characteristics of the model simulated DHBL and THBL show the height of the HBL increases radially outward from the RMW to the outer region as in observations (Figs. 4.9 and 4.12), which is in agreement with the previous studies of Smith
Figure 4.14: Mean $\theta_e$ profiles from dropsondes and model simulations. The left column (a, c, e, g) is from the dropsondes inside of 50-km radius inner-core region, whereas the right column (b, d, f, h) from the outer region. The solid lines are the mean profiles in the rear-right quadrant (RR), and the dashed lines are from all other three quadrants (OQ). Numbers of dropsondes in each group are indicated in the parentheses. The model fields are sampled according to the storm-relative locations and times of the dropsondes.
Figure 4.15: Azimuthally averaged radial winds (color) as a function of radius and height at 1800 UTC 31 August from (a) UA, (b) AO, and (c) AWO simulations. The heights of THBL and DHBL are shown in solid and dashed contours, respectively. The gray lines mark the surface RMW.
Figure 4.16: Similar to Fig. 4.15, except for averaged over each of the four quadrants. The storm forward motion points to the top.
The depth of the DHBL is almost twice as high as that of the THBL (Fig. 4.15), which is consistent with the dropsonde data shown in this study as well as in Zhang et al. (2011). A noticeable feature is that the height of DHBL increases rapidly from inside of RMW to 2RMW where the inflow ascends sharply into the eyewall, and then become somewhat flat for radius > 150 km in the outer region. The height of THBL increases outward gradually. The THBL in AO and AWO is slightly shallower than that in UA.

The actual HBL structures shown in all four quadrants (Fig. 4.16) are quite different from the azimuthally averaged mean fields as shown in Fig. 4.15. There is a dramatic front-rear asymmetry in the DHBL height. The inflow layer is much shallower in the front than the rear quadrants. This strong front-rear asymmetry exists in all model simulations, which indicates it is unrelated to air-sea coupling but a possible storm-motion induced asymmetry. Unlike the azimuthally averaged DHBL, the height of the inflow layer in the front quadrants decreases outward from 2RMW. Furthermore, the deep inflow layer is likely associated with deep convective heating in the eyewall and rainbands of the hurricane, rather than the surface friction-induced inflow. It raises two questions: 1) the representativeness of the azimuthally averaged HBL properties as shown in Zhang et al. (2011) and 2) whether the use of inflow layer to define the HBL is a valid approach.

The most noticeable asymmetry in THBL is in the rear-right quadrant in AO and AWO simulations, where the THBL is the shallowest around the storm (Figs. 4.16b-c). It is due mostly to the storm-induced SST cooling and associated stabilizing effects where the warm air flows over the colder ocean surface.

4.5.3 Effects of storm motion and deep tropospheric inflow on HBL

In order to remove the influence of the storm motion on the HBL asymmetry, we compute the storm-relative inflow by subtracting the storm translation velocity. For a comparison, horizontal maps of both the earth-relative and the storm-relative inflow layer depth in Hur-
Figure 4.17: The coupled model simulated the earth-relative DHBL (a-c) and storm-relative DHBL (d-f) heights in Hurricanes Frances (2004) and Floyd (199), and Typhoon Choiwan (2009). The inflow fields are averaged over a 2-h period in all three cases and the storm forward motion points to the top
ricane Frances from AWO simulation are shown in Figs. 4.17a and d. Both inflow fields are averaged over a 2-h period to smooth out some temporal fluctuations. For the earth-relative inflow layer, there is a front-rear asymmetry in the inflow depth with a relative minimum in the front (mostly less than 800 m), whereas the inflow layer in the rear can be as deep as 10 km. This deep inflow layer is apparently unrelated to the direct boundary layer processes. The deep inflow in the rear quadrants is also shown in the dropsonde data up to the flight level in Fig. 4.11. In comparison, the storm-relative inflow depth shows a left-right asymmetry, which varies from less than 1 km on the right to as high as 6 km on the left (Fig. 4.17d), which is consistent with the observed asymmetry shown in Kepert (2006a, b). Although the exact reason for the asymmetry in the inflow is unclear, there are indications that it may be related to the asymmetry in convective structure. As convective cells grow and decay in the cyclonic flow of a hurricane, they tend to evolve from convective to stratiform rain downwind from right to left as shown in Black et al. (2002). Stratiform rain regions are usually more inductive for enhanced mid-level inflow than convective regions, which is a topic beyond the scope of this study. The asymmetry in the inflow depth in both the earth-relative (Fig. 4.17a) and storm-relative (Fig. 4.17d) framework will be masked in the azimuthally averaged DHBL as shown in Fig. 4.15c. It is clear that the mean inflow depth does not represent the actual inflow depth or DHBL in a 3D hurricane structure. In this regard, using the inflow layer to define HBL will not be valid, which has also been pointed out previously in Kepert (2010). The question is whether these characteristics of inflow are unique to Hurricane Frances? It is difficult to fully address this question without a significantly large data set. Nevertheless, with the limited space in this paper, we show two examples to demonstrate the asymmetry in the inflow depth is independent of numerical models and/or boundary layer parameterizations and exists in other tropical cyclones. Hurricane Floyd (1999) and Typhoon Choiwan (2009) are both major tropical cyclones. The Floyd simulation is done using UMCM-MWP with the exact same AWO configuration as for Frances, while the simulation of Choiwan is from the coupled model consisted
of UMCM-WP (Section 3.1). The YSU PBL scheme is used in the WRF model. In both cases, the front-rear asymmetry in the earth-relative inflow depth is similar to that of Hurricane France (Figs. 4.17a-c). The asymmetry is shifted to the left-right orientation in the storm-relative inflow in both cases with a deep inflow layer > 6 km located on the left side (Figs. 4.17d-f). Comparing inflow in the earth-relative and storm-relative framework, the storm motion seems to enhance the inflow asymmetry in the rear with the inflow depth up to 10 km or more in all cases (Fig. 4.17a-c). The inflow fields in Floyd and Choiwan are also averaged over a 2-h period same as in Frances. These coupled model simulations and the observations from Hurricane Frances shown here as well as in Kepert (2006a, b) and Schwendike and Kepert (2008) suggest that the deep inflow layer and the asymmetry in the inflow are common features in hurricanes, although they may vary in detail from storm to storm.

4.6 Conclusions

The characteristics of the hurricane boundary layer in Hurricane Frances (2004) are examined in both numerical model simulations and the GPS dropsonde observations. The effects of the air-sea coupling on the surface winds, air-sea fluxes, and HBL structure are analyzed in detail using the fully coupled atmosphere-wave-ocean model UMCM-MWP (Chen et al. 2007, 2012a). Three numerical experiments are conducted to isolate the atmosphere-ocean and wind-wave coupling effects. Overall, the fully coupled AWO simulation of Frances produced the best storm intensity and structure in terms of the wind, surface values and vertical profiles of $\theta_v$ and $\theta_e$ compared with the dropsonde and flight-level observations from the NOAA WP-3D aircraft research flights (Figs. 4.4, 4.9 and 4.13).

The coupling to the ocean with storm-induced cooling in SST resulted in an overall weaker storm with a reduced surface enthalpy flux from the ocean. It also induces a strong asymmetry in enthalpy flux with relatively lower values in the rear-right quadrant of the
storm because of the presence of a persistent cold wake (Fig. 4.11). This feature leads to a similar asymmetry in THBL with a shallower mixed layer in the rear-right quadrant in the coupled AO and AWO model simulations, which is absent in the uncoupled UA simulation (Fig. 4.16). The wind-wave coupling enhances the surface friction-induced inflow outside of RMW due to the ocean surface wave-induced changes in the drag coefficient, which tends to produce a deeper inflow layer and DHBL in the AWO simulation compared to that without the wind-wave coupling in AO (Figs. 4.15b-c).

One of the most intriguing results of this study is that the inflow layer in tropical cyclones is highly three-dimensional and can be induced by both surface friction and convective heating in hurricane eyewall and rainbands. The azimuthally averaged inflow layer tends to misrepresent the overall inflow structure in tropical cyclones, especially the asymmetric structure (cf. Figs. 4.15-4.17), as also noted in Kepert (2010). The depth of inflow layer can be several kilometers as shown in both the dropsonde observations and full-physics model simulations. It raises the question of validity of using the inflow depth to define DHBL in tropical cyclones, especially because the frictionally and convectively induced inflows are impossible to separate in real storms. It also cautions the representativeness of the azimuthally averaged HBL properties as shown in Zhang et al. (2011) which mask some dominate features in the inflow depth and asymmetry.
Chapter 5

Tropical Cyclone Structure and Stable Boundary Layer in Super Typhoon Choiwan (2009)

An overview

In this section, we examine the mechanism via which the storm induced cold wake affects the Tropical Cyclone (TC) structure and intensity using numerical simulations. Numerical experiments of Typhoon Choiwan (2009) with and without ocean coupling are conducted. The simulation with ocean coupling shows development of a persistent stable boundary layer (SBL) in TCs over the cold wake as warm air flowing over cold sea surface. The tracer and forward Lagrangian trajectory analyses indicate that the SBL suppresses the adjacent convective activities thermodynamically and weakens the near surface wind speed dynamically. The suppressed convection leads to the longer residence time of the cold and stabilized air in the boundary (BL), which enhances surface fluxes; the wind reduction leads to the further inward turning of the near surface flow because of the gradient wind imbalance. In contrast, BL is overall neutral to unstable in the uncoupled simulation, and the air tends to go into rainbands rather than stay in BL. The air that experiences SBL in the ocean-coupled simulation therefore enters the eyewall with higher heat energy than
Figure 5.1: (a) The JMA and JTWC best-track (black) and the forecast tracks from AO (red) and UA (blue) of Choiwan from 0000 UTC 13 to 0000 UTC 17 September 2009. (b) Similar to (a), but for predicted maximum wind speed (MWS). The black dashed line is the best-track from JTWC while the black solid line is that from JMA, indicating the discrepancy between two operational agencies.

its analogy in uncoupled simulation, which experiences neutral to unstable BL. The ratio of the mass-weighted kinetic energy change and air-sea enthalpy flux further indicates that the storm with ocean coupling is more efficient than the one without ocean coupling. These results suggest that the cold wake and the associated SBL partially mitigate the direct weakening effect of the cold wake and increase the efficiency of a mature storm.

5.1 Super typhoon Choiwan (2009)

Choiwan was a super-typhoon (>100 kts) in the 2009 typhoon season in the Western North Pacific. It formed on 11 September as a tropical depression and quickly intensified as a tropical storm on 12 September. Due to the strong subtropical ridge situated northeastern of the storm, Choiwan kept moving westward along the southern edge of the ridge and passed through the warm tropical ocean (Figs. 5.1a, 5.2a, and b). Choiwan became a category-4 equivalent super typhoon on late 14 September (Fig. 5.1b) and was still steered by the subtropical ridge toward northwestward. The storm reached its maximum intensity later on
Figure 5.2: (a) The initial SST (°C) field. (b) The initial T100 (°C) field (calculated based on Price, 2009). (c) TMI/AMSRE SST (°C) swath from 13 to 17 September, which presents the minimum SST during the whole period. The black line is the JMA best-track and the blacks dot denote the observed storm center at 0000 UTC each day during the simulations. (d) Similar to (c) but for the forecast from AO. Because the ocean does not change in UA, the SST swath in UA is the initial SST.
15 September as a category-5 super typhoon. With the help of excellent poleward outflow and warm ocean conditions, it remained its peak intensity, 140 kts, until late 16 September. From 13 to 17 September, with the strong surface wind and slow translation speed (varying from 2 to 4 m s\(^{-1}\)). Choiwan induced SST cooling as strong as 3 °C (Fig. 5.2c).

Different from JTWC, the official best track data, which is provided by the Japanese Meteorology Agency (JMA, solid line in Fig. 5.1b), ranked Choiwan as a category-3 super typhoon, about 40 kts and two categories weaker than JTWC did. In general, the storm intensity issued by JTWC is about 10% stronger than that issued by JMA because of the one-minute sustained wind (JTWC) versus 10-minutes sustained wind (JMA) (Lander 2008). However, for Choiwan, the intensity issued by JTWC is 40% higher than by that issued by JMA. This large discrepancy between JMA and JTWC indicates the uncertainty in the best track data. Such uncertainty has to be considered when comparing model forecast to the best track data.

5.2 Numerical experiments

4-days forecasts for Super-Typhoon Choiwan (2009) from uncoupled WRF model (UA) and from coupled UMCM-WP (AO) are conducted. WRF version 3.1.1 was configured with triply nested domains with 12, 4, and 1.3 km horizontal grid resolution, respectively. The two inner nests are storm-following moving domains. The forecast starts at 0000 UTC 13 September 2009 with a single 12-km horizontal resolution domain spinning up for the first 5 hours. Then both the 4-km and 1.3-km moving domains join the simulation to the end. The number of grid points in 12-km, 4-km, and 1.3-km domains are 400 × 300, 202 × 202, and 283 × 283 respectively. The WRF is initialized with the 1º Global Forecast System (GFS) forecast field which is initialized at 0000 UTC 13 September. The PWP is initialized by the global Hybrid Coordinate Ocean Model (HYCOM) real-time output on 12 September, blended with real-time TRMM TMI/AMSRE SST on 12 September as well.
Because there is no data at the location of the continents in HYCOM, we apply a Laplacian smoothing function from the edge of the continent to fill in the data gaps. By using the real-time HYCOM field blended with the satellite SST, the initial ocean field contains the ocean feature appropriately and is believed to be close to the reality.

5.3 Forecasts of Choiwan in the coupled model

Figure 5.1 shows the forecast track and intensity from UA and AO in comparison with the best tracks from JMA and JTWC. The simulated tracks in both AO and UA have southward bias. By comparing the 500 hPa-geopotential height field in these two forecasts to that in the European Centre for Medium-Range Weather Forecasts (ECMWF) analysis field (not shown), we found that such southward bias comes from the overestimated strength of the ridge northeastern of the storm. The track error causes the simulated storms to pass over the waters 0.5 to 1 °C warmer (T100) than those passed over by the actual storm, which might lead to intensity forecast error.

The Maximum Wind Speed (MWS, Fig. 5.1b) shows that both storms quickly intensify on 13 September, and become mature storms at the beginning of 14 September, same as the observed storm. Later, they stop strengthening for about 18 hours. Late on 14 September, storms in two simulations start to evolve separately. UA starts to rapidly intensify and reaches its peak intensity late on 15 September. AO, with smaller intensification rate, peaks near the beginning of 16 September. They are both at quasi steady state early on 16 September and start to weaken later on the same day. At the quasi steady state, UA is a strong category-4 typhoon, which is close to the JTWC best-track data but is too strong when compared to the JMA. On the other side, AO is a weak category-4 typhoon that is close to JMA but too weak when compared to JTWC. Although for various reasons, neither AO nor UA could match the two best tracks, both forecasts produce reasonable storm intensities that lie in between. Without direct measurements, the best-track uncertainty (ad-
Figure 5.3: Time series of model simulated surface (a) enthalpy fluxes, (b) latent heat fluxes, and (c) sensible heat fluxes averaged within 400 km in radius from storm center (AO-red, UA-blue).

dressed in the previous section) also makes it almost impossible to assess the accuracy of forecast. Nevertheless, the near identical tracks of UA and AO implies resemble large-scale environment (at least the steer flow) for both storms. Therefore the differences of the storm intensity could be assumed solely due to the air-sea coupling, i.e., the change of surface enthalpy fluxes.

Figure 5.3 is the area-averaged surface heat fluxes in AO and UA, which shows that AO gains sensible and latent fluxes about 75% of what UA does. Although ocean coupling has started affecting the storm energetically in the beginning of the forecast, it takes about one day for the MWS responds to the reduced enthalpy fluxes (comparing Fig. 5.3 to Fig. 5.1b).
Figure 5.4: Surface enthalpy fluxes (W m$^{-2}$) from (a) AO and (b) UA averaged over 12 hours from 0000 to 1200 UTC 16 September. The black contours in (a) and (b) are the SST isotherms. The storms move toward west-northwest as indicated by the black arrow.

As what we have seen in Chapter 4, the most dramatic enthalpy fluxes reduction in AO occurs at the eyewall and at the cold wake area because of the largest wind reduction and the strongest SST cooling in AO respectively (Figs. 5.4 and 5.5).

During the whole forecast period, we choose the first 12 hours on 16 September as our period of interest. It is chosen because both AO and UA have reached their peak intensity and are in a nearly steady state and the eyewall in both storms are rather symmetric (Fig. 5.1). Because there is no in-situ observation available for Choiwan, in the following analysis we focus on describing the air-sea coupling processes through analyzing the differences between AO and UA.

5.4 **Stable boundary layer in TCs**

Chapter 4 has showed that the cold wake affects the thermal and dynamic structures in the BL. Similar results are also found in our simulations and we will not repeat the same analy-
Figure 5.5: Similar to Fig. 5.4 but for 10-m wind speed (m s$^{-1}$).

Historically, several stability parameters are designed/used to present various physical processes. The focus of the BL stability in this study is the capability of the turbulent mixing both at surface layer and that of the buoyancy convection in the BL over the cold wake. In this regard, the surface stability ($\zeta$) is defined as Monin-Obukhov stability index:

$$\zeta = \frac{z}{L} = \frac{-kzg(w'\theta'_c)_{sfc}}{\overline{\theta'_i}u_*^3}$$

(5.1)

while the Obukhov length is given by:

$$L = \frac{-\overline{\theta'_i}u_*^3}{kg(w'\theta'_c)_{sfc}}$$

(5.2)

In Eq. (5.1), $k$ is the Von Karman constant, $w$ is the vertical velocity, $g$ is the gravity, $u_*$ is the frictional velocity. All the fields can be separated into mean (denoted as “$-$”) and perturbation components (denoted as “$'$”). The sign, not the magnitude, of $\zeta$ can be
related to static stability near the surface: negative implies unstable while positive implies statically stable. In the simulations, Monin-Obukhov stability index is directly output from the surface layer scheme from numerical model.

For BL, we use the static stability parameter that is defined as the vertical gradient of $\theta_v$:

\[
\frac{\partial \theta_v}{\partial z} > 0 \quad \text{stable BL} \\
\frac{\partial \theta_v}{\partial z} = 0 \quad \text{neutral BL} \\
\frac{\partial \theta_v}{\partial z} < 0 \quad \text{unstable BL}
\] (5.3)

The static stability is calculated on the resolving scale at each grid point, and it can be seen as the vertical gradient of mean $\theta_v$ over the individual grid cell. The SBL is then defined as the BL within which every model level is static stable and the BL here is defined as a well-mixed layer (THBL in Chapter 4), the height at which the $\theta_v$ is 0.5 K higher than its surface value.

As addressed in the Chapter 2, one particular feature of SBL is the shallower BL depth. In the simulations in this study (Fig. 5.6), there are shallow BLs (<500 m) in the eyewall and primary rainband. Outside of the inner core, another shallow BL in UA appears in a spiral shape. In AO, an additional shallow BL occurs at where the cold wake is.

The shallow BL in the eyewall is partially due to the dynamic constrain associated with the high relative vorticity in TCs, as addressed in Chapter 4. Another potential cause of the shallow BL in eyewall is the strong eyewall convection. Beneath the strong convection, the evaporative cooling results in the convective downdraft that brings down the cold air into the BL. At the rear part of convective band, there is mesoscale descent that could suppresses the depth of the BL by entraining mid-level dry air in at top of the BL (Powell, 1990). The shallow BL associated with the primary rainband as well as the one with spiral shape in UA
are also the consequence of these mechanisms. The shallow BL over the cold wake in AO, however, is more likely due to the increase of the BL stability that enhances the vertical stratification.

Based on Eqs. 5.1 and 5.3, the horizontal distribution of stable surface layer and SBL in UA and AO are presented in Fig. 5.7. The stable surface layer covers the cold wake area outside of the primary rainband; the SBL covers a smaller region inside the stable surface layer area. This indicates that the cold wake first stabilizes the surface, and then vertically stabilizes the whole BL. In other words, the stable BL is a result of the temporally and spatially integrated influence of the cold wake. Beneath the convection, convective downdraft is not able stabilized the surface layer and BL layer although it can suppress BL depth (or increase the BL stratification). Besides, winds are stronger near convection, so less residence time of air parcels to be affect by the cold SST. Without ocean coupling processes, there is no SBL in the UA.

To further demonstrates the vertical thermal structure in SBL and that in neutral/unstable BL, we examine the vertical profile of $\theta_v$ in AO and UA (Fig. 5.8). The profiles are divided
Figure 5.7: The SBL (black) and stable surface layer in (a) AO and (b) UA averaged over 12 hours from 0000 UTC to 1200 UTC 16 September. The red contour denotes SST with an interval of 0.5 °C, and the thick red line is the 29.5 °C SST isotherm. The storm moves toward west-northwest direction as indicated by the black arrow.

Figure 5.8: $\theta_v$ profiles in the lowest 1 km from (a) AO and (b) UA at 1200 UTC 16 September. Profiles are grouped based on SST, as labeled at top of each column. For each group, we present only a subset of the sample to avoid unnecessary clutter. The gray lines indicate the profiles with unstable layers near the surface while the red lines indicate the profiles with stable layers only in the BL. The black line is the mean profile of each group. All profiles are from an annulus with an inner radius of 150 km and an outer radius of 400 km. For each figure, the x-axis ranges from 304 to 306 K.
into four groups based on the corresponding SST and are sampled outside of inner core to bring into focus of the adjustment of BL structure due to the cold wake exclusively (i.e., outside of eyewall and primary rainbands, under which the BL dynamics is dominated by convection). Profiles in UA (Fig. 5.8b) indicate that the environmental SST is around 28.5-29.5 °C. Therefore, any profile locates at where the SST is below 28.5 °C in AO is affected by the ocean cooling. The main characteristics of $\theta_v$ in BL therefore changes from one group to another. In the group of SST $>29$ °C (for both UA and AO), the mean profile (black line) can be separated into three layers with increasing height: the unstable layer - $\theta_v$ decreases with height, the well-mixed layer - $\theta_v$ keeps almost constant with height, and the stable layer - $\theta_v$ increases with height. For the individual profile, the well-mixed layer sometimes is too thin to be noticed, but the unstable and the stable layers are always clear. For the group of $28.5$ °C $\leq$ SST $< 29$ °C, the mean profiles in both simulations show three distinct layers still, although the unstable layer disappears in some profiles in AO. There is also more variability among profiles within this group. In particular, some of them have stable layers only (red lines). For the groups of SST $< 28.5$ °C (only in AO), there are more profiles with only stable layer (rad lines). According to Eq. 5.3, red lines in Fig. 5.8a represent the vertical stratification of SBL.

5.5 The impact of stable boundary layer on TC structure

In attempts to find how the SBL changes the storm convective structure, especially the organization of convection and the evolution of the near surface airflow, we try to follow the low-level air over SBL with trajectory and tracer analysis (see Chapter 3).

Both trajectories and tracers are released at 0000 UTC 16 September and are traced for 6 hours. The total amount of tracer is conserved during this 6-hour period (not shown) and the air-parcels have sufficient time to reach the eyewall. In AO, the tracers are released 150 km radius (again, roughly the outer edge of the primary rainband) away from the storm
Figure 5.9: The radar reflectivity at 80 m on 0000 UTC 16 September overlaid with the initial location of the tracers (blue contours) and trajectories (blue dots) in (a) AO, and (b) UA. In AO, the tracers released in the locations marked by dark-blue contours covers area of cold wake (cold-wake tracer), 90° downstream from cold wake (top one, downstream tracer), and 90° upstream from cold wake (lower one, upstream tracer). The tracer released in the light blue contours covers an annulus with inner radius of 150 km and outer radius of 350 km. Each dark-blue contour covers an area of 1/20 of the annulus. The trajectories are released over the cold wake. All the tracers and trajectories are released at 80 m in height. In UA, tracers and trajectories are released based on the same storm-relative location. The thick gray line denotes the 28.5 °C SST isotherm, which marked the location of the cold wake.
center over 1) the stable BL over the cold wake, 2) the region 90 degree upstream of the
cold wake, 3) the region 90 degree downstream of the cold wake and 4) the annulus with
an inner radius of 150 km and an outer radius of 350 km (Fig. 5.9a). For convenience,
we refer the tracer released from these four areas as cold-wake tracer, upstream tracer,
downstream tracer, and annulus tracer hereafter. The upstream and downstream tracers
cover an area the same size and shape as the cold-wake tracer does. These three tracers
cover the sub-region (about 1/20) of the annulus tracer does. All four regions are marked
by the blue and light blue contours in Fig. 5.9a. Vertically, they are all released at 80 m,
the first model layer that is above the surface layer in WRF. For each tracer, every grid cell
within the blue contours is set to 1 while everywhere else is set to 0. Therefore, the value
of each grid point can be seen as the density or concentration of the tracer. Note that the
tracers do not interact/mix with each other. We release 86 trajectories at cold wake region
indicated by the blue dots in Fig. 5.9a. Again, all trajectories are released at 80 m height. By
releasing the trajectories extensively, the ensemble path and the thermodynamic properties
of the particles could exhibit the evolution of BL flow under the undiluted assumption.
For comparison, tracers and trajectories are also released in UA at the same storm relative
location as in AO (Fig. 5.9b).

5.5.1 The evolution of near surface airflow

Figure 5.10 shows the 0.05 iso-surface of the cold-wake tracer at $t_{\text{tracer}} = 0, 20, 60,$ and
120 min ($t_{\text{tracer}}$ indicates integrating time for tracer and trajectory calculation). The cold-
wake tracer is well mixed in the BL within 20 minutes due to the turbulent mixing. At
$t_{\text{tracer}} = 1$ hour, the difference between cold-wake tracers of AO and UA becomes apparent.
There is less vertical transport and faster inward spiraling air in AO from the cold wake area
compared to that in UA. $t_{\text{tracer}} = 1$ hour, a larger amount of the tracer in AO is transported
into the eyewall while the tracer in UA is mostly transported upward into the rainband
convection.
Figure 5.10: The 0.05 iso-surface (gray) of the cold-wake tracer in AO (top panels) and UA (bottom panels) at $t_{\text{tracer}} = 0, 20, 60, 120$ minutes. The color shading indicates the SST.
Figure 5.11: The azimuthally-integrated tracer concentration (shading) from 0300 UTC to 0600 UTC 16 September in AO (left) and UA (right). The contour shows the azimuthally averaged BL depth, and the dots indicate the RMW at each model level.
The azimuthally-integrated tracer concentration from $t_{\text{tracer}} = 3$ to 6 hours, as shown in Fig. 5.11 exhibits higher tracer concentration in the BL in AO than in UA, especially at $t_{\text{tracer}} = 3$ and 4 hour. Thus, instead of being carried vertically into the rainband convection as the one in UA, cold-wake tracer in AO is horizontally transported into the inner core area and then is carried upward with the convective updraft in the eyewall. As a result, the cold-wake tracer concentration in the eyewall in AO is much higher than that UA at $t_{\text{tracer}} = 6$ hour.

Identical analyses for the upstream and downstream tracers are illustrated in Figs. 5.12 and 5.13. Between AO and UA, there is no striking difference of these two tracers. The upstream tracer is strongly affected by the primary rainband in both cases and immediately entrains into rainband convection with little of it entering the eyewall. Without being affected by the cold wake and primary rainband, the downstream tracers start to rise up in the eyewall until $t_{\text{tracer}} = 3$ hour. The upstream tracer concentration in the eyewall in UA is a little higher than that in AO, which could be due to the stronger inflow in the stronger storm in UA.

A more quantitative way to address the evolution of air parcels originating from these tracer locations is to calculate the fraction of tracer that ends up in the BL, in the eyewall and in other locations (Fig. 5.14). The fraction of tracer that ends up in the BL is calculated by integrating the product of tracer concentration and air mass at each grid cell over the whole BL but outside of eyewall. For convenience, we call it BL-air since a tracer is analogous to an air parcel. Following the same method, we can get eyewall-air, the fraction of tracer that ends up in the eyewall, and the outer-air, which represents the remaining air.

In the beginning, all four tracers belong to BL-air. The amount of BL-air decreases with time as the turbulent mixing and rainband convection begin to transport air into the outer area right after tracers are released. We first compare the differences of these three air-groups between UA and AO, among cold-wake, upstream, and downstream tracers (Fig. 5.14a-f). The decreasing rates of BL-air in UA and AO are similar to each other.
Figure 5.12: Similar to Fig. 5.11, but for upstream tracer.
Figure 5.13: Similar to Fig. 5.11, but for downstream tracer.
Figure 5.14: The fraction of tracer originating from (a, b) cold wake, (c, d) upstream, (e, d) downstream, (g, h) annular area that stays in BL (diamond), enters the eyewall (diamond-line), and ends up in the area outside of BL and eyewall (solid line) (AO-red and UA-blue).
from upstream and downstream tracers while they are more diverse from the cold-wake tracer. In the cold-wake tracer, the BL-air decreases in a slower rate in AO than that in UA. Similar to the decreasing rate of BL-air, the increasing rates of outer-air in UA and AO are more diverse when they originate from the cold-wake tracer. Compared to the initial value of BL-air, only about 10% of them are able to enter the eyewall, becoming the eyewall-air. The evolution of eyewall-air is very different from one originating location to another. The difference among the tracer groups and that between UA and AO are consistent to what we have found in Figs 5.11, 5.12, and 5.13. The eyewall-air in AO is about twice as much as that in the UA from cold wake tracer; there is almost no eyewall-air from upstream tracer; the eyewall-air in downstream tracer in AO is about 75% as much as that in the UA, which might simply reflect the strength of storms.

Apparently, at any given location, the fraction of air that goes in to the eyewall is controlled by various factors (or the combination of them), such as the SBL, the primary rainband, and the storm intensity. Presumably, the storm intensity dominates. The question is whether the SBL can cause a noticeable change in the total amount of eyewall-air. We study this question by comparing the evolution of eyewall-air that is originated from the annulus-tracer (Fig. 5.14g, and h) to those are originated from the other three tracers. Results shows that the eyewall-air in the AO is slightly smaller than that in the UA, but the ratio between them is larger than 75% (the ratio between AO and UA from downstream tracer). While SBL covers only a small fraction of the whole storm, it might have a noticeable impact on bring in more air from BL into the eyewall with other indirect physical mechanism. The further detail discussion will be provided in the following subsection and Section 5.6.
5.5.2 Thermodynamic and dynamic forcing associated with stable boundary layer

The airflow is driven by either dynamic and/or thermodynamic forcing. Here we try to separate the cause of the different behavior of the cold-wake tracers between AO and UA into these two factors.

As the BL stability is important for supporting convection, we began our analysis by comparing the convective organization between UA and AO. Figure 5.15 shows the simulated radar reflectivity from AO and UA at 1 km height. Often, in order to avoid getting the transient features, time-composited is applied when analyzing storm structure. However, the convective features are likely to be smoothed out or diminished when averaging over time. In this regard, we use a series of snap-shot (every 3 hours from 0000D UTC to 0600 UTC 16 September) to show the time evolution and time persistence of the convective features. The resemblances of the convection between UA and AO include the symmetric eyewall, the radius of maximum wind (RMW, ~35 km), the location of the primary rainband, and the persistent outer rainband southeast of the storm. The differences between two simulations are the convection outside of the eyewall in AO is more confined, and the convection in the northeastern quadrant, denoted as the dashed-black box, in AO is persistently less active. This northeastern quadrant is downwind adjacent to the cold wake, and this could be the stabilized air with low buoyancy moves downstream and suppresses the convective activities there. Thus, the thermodynamic forcing of SBL is to suppress rainband convection downstream adjacent to the cold wake.

The thermodynamic impact of SBL leads to the long residence of BL air, which explains why there is higher cold-wake tracer concentration in BL in AO than UA in Fig. 5.11. However, this does not explain why there is the further inward eyewall penetration of cold wake tracer in AO. According to the studies of mid-latitude SST front (eg. Small et al. 2008), we suspect that SBL in TC also dynamically reduces the surface wind. In TCs,
Figure 5.15: The simulated radar reflectivity at 1 km altitude in AO (left column) and UA (right column) every 3 hours from 0000 UTC to 0600 UTC 16 September. The magenta contours show the 28.5 and 29 °C SST isotherms. The dashed-black line encloses the area downwind adjacent to the cold wake in AO, and in the same storm-relative location in UA. The storm motion is indicated by black arrow.
Figure 5.16: (a) Perturbation field of 10-m winds (shading) and SST (contour) across the cold wake at 0000 UTC 16 September. It covers an area with an inner radius of 150 km and an outer radius of 350 km. (b) Same as (a) but for $V_t$. (c) Same as (a) but for $V_r$. The black diamonds in (a-c) indicate the center of a storm. (d) Perturbation field of 10-m winds in (a) as a function of SST. The white diamond indicates the medium value of the wind perturbations wind at given SST, while the edges of thick (thin) black bar represent 25% (10%) and 75% (90%) tile respectively.
Figure 5.17: The 10-meter wind vectors for AO (red) and UA (blue) at 0000 UTC 16 September. The gray shading shows the 1 °C SST cooling and the storm motion is roughly to west-northwest (black line). The cold wake and it adjacent downstream area is enlarged at the right.

the surface wind strongly depends on the radial distance from a storm center, and thus the radial gradient of wind velocity makes it difficult to discern the azimuthal changes caused by the cold wake. For this reason, we use the perturbation field of surface winds to examine the reduction of surface winds over the cold wake. The perturbation field is defined as the difference between the total wind and the azimuthally averaged wind at each given radius:

\[ wspd'_{10}(\phi, r) = wspd_{10}(\phi, r) - \overline{wspd_{10}(r)} \]  

In Eq.(5.4), \( \phi \) is the azimuthal angle and \( r \) is the radius from the storm center. The over-bar is the azimuthal mean at a given radius. The horizontal map of the 10-m wind perturbation across the cold wake (Fig. 5.16a) suggests the reducing of wind speed over the cold wake and downstream adjacent to the cold wake. The scatter plot of the perturbation wind field and the corresponded SST (Fig. 5.16d) suggests a trend, although weak, of which the wind perturbation decreases with decreasing SST. From SST = 29.5 °C to SST = 28.5 °C,
the wind perturbation decreases by about 2 m s$^{-1}$ in average (up to 8 m s$^{-1}$). What this means is that at the same radius, the surface wind reduced by about 2 m s$^{-1}$ (in average) when moving across the cold wake. While the wind decreases over the cold wake, the imbalanced gradient wind allows further inward turning of surface winds. This causes the increase of inflow angle and potentially the stronger surface inflow. With similar calculation, the tangential component ($V'_t$, Fig. 5.17b) of 10-m wind perturbation decreases over the cold wake while the and radial component ($V'_r$, Fig. 5.17c) increases. This is expected since the tangential component represents greater magnitude of the winds while the radial component represents more deviation of wind from the balanced gradient wind.

Another way to examine the dynamic impact is to analyze the inflow angle. Fig. 5.17 shows the 10-meter wind vectors in AO (red) and UA (blue). Because the storm in UA is 5 to 10 kts stronger than that in AO at this time, the stronger friction and therefore greater inflow angle are expected in UA. The mean inflow angle over the whole 800×800 km inner domain in UA is 21.37° with a 90% confidence interval between 21.28° and 21.46°. It is 18.89° with a 90% confidence interval between 18.8° and 18.98° in AO. Over and near the cold wake, (gray area in Fig. 5.17), there is an apparent further inward turning of 10-meter wind in AO (comparing red arrows to the blue ones) and the mean inflow angle is 21.3°. There is no significant increase in mean inflow angle in UA when considering the same storm relative area.

Comparing the time-composited horizontal maps of inflow angle and inflow velocity (Fig. 5.18) in two simulations can further identify the persistent location of enhanced inflow and inflow angle. For each simulation, the strong inflow in the inner core locates in the front (west) side of the storm, which is consistent with the previous observational findings (Shapiro 1983). Outside of the core area, the most significant high inflow angle occurs at south of the storm center, which could be the confluent flow due to the storm motion or due to the primary rainband. The different field (Fig. 5.18c and f) shows that the
Figure 5.18: Time-averaged 10-m inflow angle (a-c) and radial wind (d-f) in AO (left column), UA (middle column), and the difference field (right column). The black contours in (a), (b), (d), and (e) are the 28.5 and 29 °C SST isotherm and they indicate the 1 °C SST cooling in (c) and (f). The storm moves toward west-northwest direction as indicated by the black arrow.
Figure 5.19: (a) The downward turbulent momentum fluxes averaged over the region across the cold wake as shown in Fig. 5.16. (b) The mean SST over the same area. The x-axis is the distance related to the lowest mean SST, the center of the cold wake. From left to right is from the upstream region to downstream region.

SBL enhanced inflow/inflow angle persistently occurs downstream of the cold wake in the inner core and over the cold wake in the outer region.

Figure 5.19 shows the turbulent momentum fluxes on the grid cell averaged over the area across the cold wake, as shown in Fig. 5.16. The turbulent momentum fluxes apparently decrease toward low SST, and therefore the cause of the wind reduction over the cold wake is the decrease of downward turbulent momentum fluxes. Another possible cause for weakening surface winds is the SST-gradient-induced hydrostatic pressure gradient (Small et al. (2008)). In TC, however, the hydrostatic pressure gradient is O(5) less than the turbulent momentum fluxes in the tangential wind budget, and therefore the hydrostatic pressure gradient is not essential.
5.6 Stable boundary layer and TC energetics

5.6.1 Evolution of equivalent potential temperature along the airflow

In order to understand the energetic impact of SBL, we examine the evolution of $\theta_e$ (Bolton 1980) along trajectories. A subset of the trajectories are shown in Fig. 5.20. In both simulations, trajectories start with moderate $\theta_e$ values (orangish and yellowish), but these values in are smaller that those in UA. All the trajectories can be divided into eyewall-trajectories and rainband-trajectories based on their path individually. The eyewall-trajectories stay in BL longer and gain extra energy on its way to the eyewall ($\theta_e$ is increased near surface with reddish color). In contrast, the rainband-trajectories loose the high energy while being transported upward ($\theta_e$ is decreased with height with bluish color). In the end, the eyewall-trajectories has $\theta_e$ 10 k higher than the rainband-trajectories above the BL. These seperation can also be seen in the vertical-cross section of $\theta_e$ (Fig. 5.21). We further exam-
Figure 5.21: Vertical cross section of $\theta_e$ (shading) and 0.1 m s$^{-1}$ constant vertical velocity contour in AO (left) and UA (right) at 0000 UTC 16 September. The mean values of $\theta_e$ at surface and $\theta_e$ at RMW are showed on top of each figures.

ine the $\theta_e$ along all trajectories in Fig. 5.22. There are 22 trajectories entering the eyewall in AO while there are only 15 of them in UA. There is no doubt that the extreme value of $\theta_e$ in the eyewall in UA is higher than AO (by ~2 k) and the mean value of the near surface $\theta_e$ in UA is also ~1.5 k higher than that in AO. However, the inflow over the cold wake region in AO brings in more high energetic air particles from the BL into the eyewall, and the mean value of RMW $\theta_e$ in AO is close to that in UA (shown in Fig. 5.21). Besides, the active outer convection in UA could lead to the entrainment of low $\theta_e$ airflow associated with convective downdraft. As pointed out by Powell (1990), the disturbance of the low $\theta_e$ to BL air that may incorporate into the storm core area can potentially weaken the storm. Therefore, with the help of the SBL, AO could potentially get more energy, relative to what is available from the ocean, into the eyewall. As the latent heat released in the eyewall is the main heat energy for TC to intensify, the storm-induced cold wake could energetically affect the storm structure by stabilizing the BL, modifying the properties of inflow and increase the energy that available for storm intensification. This extra energy could mitigate the oceanic negative feedback, help maintain or strengthen the storm intensity in AO.
Figure 5.22: The equivalent potential temperature along the trajectories as a function of height in (a) AO and (b) UA for the 6-hour trajectory calculation. They are separated into 2 groups: those entering the eyewall (red), and those entering the outer rainband convection (blue). The thick lines show the mean profiles respectively and the numbers indicate the data size of each group. The dots show the mean value of each group at the lowest level.
The other point we would like to make is the non-conservation $\theta_e$ in the deep convection in Fig. 5.22. For the eyewall trajectories, the variation of $\theta_e$ in the strong updraft period is not too large, about 1 degree. This small variation could be from the assumption set in Bolton (1980), i.e., it is conserved in pseudo adiabatic processes, but not conserved for reversible adiabatic processes or anything in between. Because deep convection in the full physics model will always bring some liquid water with it, $\theta_e$ will never be strictly conserved (Bryan 2008). Nevertheless, for the trajectories, either in the eyewall or in the rainband group, with which the $\theta_e$ varies more than 5 degree, dry air entrainment will be the main reason. The bottom line is, there is no “close convective system” in a full physics model, as what should be in reality. Therefore there is no conserved $\theta_e$ along each trajectory in our simulations.

5.6.2 TC efficiency

A TC can be viewed as a heat engine converting heat energy extracted from the ocean into the kinetic energy (KE) through the diabatic heating in the moist convection (Emanuel, 1986; see detail in Chapter 2). The efficiency of converting thermal energy into KE in a storm is therefore defined as a ratio between the change of mass-weighted KE ($\Delta KE$) and the mass-weighted surface enthalpy fluxes within a control volume. There are certainly a few intermediate steps within this process, and each of them has its own intermediate efficiency. However, we ignore all intermediate processes and focus on only the initial heat source and the ultimate dynamic strength of the storm. The $\Delta KE$ is calculated based on the difference of the kinetic energy per mass between the beginning and the end of the period of interest. The surface enthalpy flux is calculated by averaging the 10-minutes instantaneous surface enthalpy fluxes per mass over the period of interest. Weighting each term by mass can avoid being affected by the change of total mass associated with the storm intensity. Then, AO efficiency is normalized by the UA one. If this value is larger than one, AO is more efficient than UA and vise versa.
Table 5.1: From right to left, the first three columns are the change of mass-weighted kinetic energy and surface enthalpy fluxes, and the efficiency of converting heat energy to kinetic energy within a control volume (cylinder) with the lateral boundary of 350 km. The forth column shows the efficiency. The fifth column shows a ratio of AO efficiency to the UA one. The right most one is the inward enthalpy fluxes in the BL at the outer edge of RMW (70 km).

<table>
<thead>
<tr>
<th>Models</th>
<th>ΔKE (J s⁻¹ Kg⁻¹)</th>
<th>EFLX (J s⁻¹ Kg⁻¹)</th>
<th>Efficiency</th>
<th>AO/UA</th>
<th>Inward MSE (1e15 J) at ~RMW</th>
</tr>
</thead>
<tbody>
<tr>
<td>UA</td>
<td>0.00139</td>
<td>0.0474</td>
<td>0.0295</td>
<td></td>
<td>1.33</td>
</tr>
<tr>
<td>AO</td>
<td>0.00135</td>
<td>0.0393</td>
<td>0.0344</td>
<td>1.166</td>
<td>1.55</td>
</tr>
</tbody>
</table>

The assumption of the efficiency calculation is that within a control volume, a cylinder, the only energy source we care is the surface enthalpy fluxes. In other words, the energy brought in from the lateral boundary of the control volume is negligible, and therefore the control volume needs to cover an area large enough. Hence the first step here is to define the size of control volume. Through analyzing the radial flux of moist static energy \( V_r \frac{\partial (\text{Moist-static-energy})}{\partial r} \), we find that the radial advection of moist static energy is getting close to zero beyond around 200 km in radius and it becomes almost zero beyond 400 km. We first use 350 km as the radius of the cylinder.

We calculated the storm efficiency over the 6 hours from 0000 to 0600 UTC 16 September. Results (Table 5.1) shows that UA has efficiency of 2.95% while AO has it of 3.44%. A weak storm in AO is 16.6% more efficient than its stronger analogy in UA. Figure 5.14 shows that SBL has noticeable effect on the fraction of air that enters into the eyewall. To further examine if the SBL actually increase the total inward moist static energy fluxes at the eyewall, we calculate the inward moist-static energy fluxes at outer edge of the eyewall within the BL (Table 5.1: right most column). There is more inward moist static energy at the outer edge of the inner-core in AO than that in UA during this period, which again indicates a higher efficient storm in AO. Identical analyses with different size of control volume are also examined (Table 5.2). Within from 300 to 500 km in radius, various radius re-
<table>
<thead>
<tr>
<th>Radius(km)</th>
<th>Models</th>
<th>300</th>
<th>400</th>
<th>500</th>
<th>700</th>
<th>1000</th>
</tr>
</thead>
<tbody>
<tr>
<td>AO</td>
<td>0.0342</td>
<td>0.0345</td>
<td>0.0325</td>
<td>0.0220</td>
<td>0.0142</td>
<td></td>
</tr>
<tr>
<td>UA</td>
<td>0.0244</td>
<td>0.0322</td>
<td>0.0324</td>
<td>0.0223</td>
<td>0.0139</td>
<td></td>
</tr>
<tr>
<td>AO/UA</td>
<td>1.40</td>
<td>1.07</td>
<td>1.003</td>
<td>0.986</td>
<td>1.02</td>
<td></td>
</tr>
</tbody>
</table>

Table 5.2: The sensitivity test of the storm efficiency to the size of control volume. The top row is the radius of the cylinder, and the second and the third rows are the storm efficiency of UA and AO, respectively. The last row is a ratio of AO storm efficiency to the one in UA.

Our results suggest that at the quasi-steady state, the energy gained by BL inflow due to the surface enthalpy fluxes outside of the eyewall plays an important role in storm efficiency. Similar idea has also been argued in Wang and Xu (2010) who showed that the lateral inward energy flux contributes significantly to the energy balance in the eyewall. With the suppressed rainband convection and enhanced inflow due to the presence of SBL, more high-energy air parcels are carried by the HBL inflow, inward transporting into the eyewall region in AO. Therefore, the storm in AO is more efficient than UA. Nevertheless, the storm in UA was stronger, because there is more overall energy available from the warmer SST.

### 5.7 Conclusions and discussions

In this study, we discussed an important issue about how the storm-induced cold wake changes the BL characteristics of a mature storm and then the convective structure energetically, which eventually increases the storm efficiency. The whole study was built up by using UMCM-WP forecast of Super Typhoon Choiwan (2009) and the comprehensive tracer and trajectory analysis. For comparison, two model forecasts with (AO) and without (UA) ocean coupling were conducted.
Forecast results revealed the similar track and intensity evolution of storms from UA and AO. Because of neglecting the negative feedback from the ocean model, the maximum intensity in UA was ~10 kts stronger than the one in AO. The summarized total surface enthalpy fluxes showed that AO got less energy from the beneath ocean than UA did by 25%. During the period of interest, the SST cooling was about 1.5 to 2 °C cooling over the cold wake. The cold wake led to the formation of stable surface layer and stable boundary layer (SBL). The stability analysis suggested that the SBL forms over the cold wake outside of the primary rain band, and it covered narrower area than the stable surface layer did. This result indicated cold wake stabilized surface layer first and then vertically stabilized the BL. Tracer and trajectories analysis suggested the thermodynamic influence of the SBL is to suppress the convection from the cold wake area to its adjacent downwind area. Dynamically, the SBL enhances the inward turning air-flow because of the imbalance momentum equation due to the suddenly reduced wind speed. By examining the momen-
tum equation, we found that the wind reduction is mainly caused by the less downward transported turbulent momentum fluxes.

Via analyzing the evolution of equivalent potential temperature along the trajectory and storm efficiency, a new mechanism in AO was found here. We present this physical mechanism as a schematic figure in Fig. 5.23: The stabilized BL over the cold wake alters the inflow properties, suppresses the convection and enhances the surface inflow. It eventually gains some extra energy, carries it into the eyewall, and releases the energy through the latent heat. In contrast, the air-particle originated from the “cold-wake” region in UA wastes their high energetic property by entering the outer rainband convection. In the end, this mechanism increased the efficiency of AO. Given that there is inflow in a storm regardless the cold wake, it is the thermodynamic impact that helps increase the high energetic air in the eyewall. While the direct negative feedback caused by SST cooling dominates the storm intensity, the stabilizing effect of the cold wake can oppose and partially mitigate this negative feedback.
Chapter 6

Coupled Observations and Modeling over the Cold Wake in Typhoon Fanapi (2010)

An overview

In this chapter, we study the spatial variability of the atmospheric and upper oceanic boundary layers (BLs) across the TC-induced cold wake, with both observations from the Impact of Typhoons on the Ocean in the Pacific (ITOP) project and numerical simulations with UMCM-WP. The case chosen here is Typhoon Fanapi (2010); during its lifetime, a rich set of air-sea observations from various instruments was collected. We also use ITOP observations to examine the physical mechanisms associated with the impact of the stable boundary layer (SBL) on TC structure.

ITOP data show that as warm air flows over the cold sea surface (over the cold wake), the air-sea thermal disequilibrium reverses direction, leading to negative (i.e. downward) sensible heat fluxes. The largest negative sensible heat fluxes occur where warm air advected from upstream has not been modified by convective downdrafts. The air-sea moisture disequilibrium decreases over the cold wake as well. However, the greatest reduction in latent heat fluxes occurs where the BL air is nearly saturated due to rain evaporating into
the BL in convection. Afterward, the airflow moves back over the warm ocean, thermal and moisture air-sea disequilibrium is enhanced, and therefore the sensible and latent heat fluxes. We showed in Chapter 5 that the SBL forms over the cold wake in numerical experiments. Here, we show that ITOP data not only confirm those results, but also show that the static stability of the BL varies from close to neutral upstream of the cold wake, to strongly stable in the cold wake, and then to unstable downstream of the cold wake. The dynamic impact of the SBL, i.e., the increased inward turning of the surface winds, is also verified through analyzing the inflow angle. A numerical simulation of Fanapi with ocean coupling has an almost perfect intensity prediction. Such a performance is a consequence of the fairly well-simulated spatial variability of the observed coupled air-sea BLs across the cold wake in UMCM-WP. This includes the horizontal distribution of SST, surface wind, air-sea fluxes and thermodynamics and dynamics in the coupled BLs, and the atmospheric BL stability. Under the assumption that the coupled model simulation is better representative of reality, a tracer and trajectory analysis in the coupled model simulations are again used to further quantitatively show the energetic impact of the SBL. An identical analysis has also been conducted in the uncoupled simulation. These analyses show that the increment of heat energy that the airflow gains while on its way from the rear-right quadrant (where the cold wake is found if ocean coupling is considered) to the eyewall is higher in the simulation with ocean coupling than the one without. This again shows that the coupled air-sea BL variability can mitigate the oceanic negative feedback to certain degree.

6.1 Typhoon Fanapi (2010)

Typhoon Fanapi was a category-3 equivalent typhoon in 2010 in the Western Pacific. It developed from a tropical easterly wave and became a tropical depression on 1800 UTC 14 September near 19.6 N and 129.1 E (Fig. 6.1). It moved northwestward at 9 kts along the southwestern periphery of a low- to mid-level steering ridge and the majority of deep
Figure 6.1: JTWC best-track (black) and simulated (red-AO, blue-UA) (a) tracks, (b) maximum wind speeds (MWS), and (c) minimum sea level pressures (MSLP). In (a), the circle indicates the center of storms at 1200 UTC 14 September, the beginning of the simulations; the diamonds are the center location at 0000 UTC each day. The initial location for the storms in UA and AO are the same. In (b) and (c), diamonds denote the storm intensity every 6 hours.
convection was in the eastern and northern periphery (Fig. 6.2a). The system became a tropical storm on 1200 UTC 15 September with a curved convective band wrapping in from all quadrants and a deep convection near and over the system center, especially to the east side of the storm. Around this time, Fanapi slowed down considerably and then started to move northward because of the mid-latitude trough moving over Kyushu (Japan) weakened the steering flow associated with the subtropical ridge. The mid-latitude trough also may have helped the storm continuously intensify by enhancing its poleward outflow. In addition to the enhanced poleward outflow, Fanapi was also in a favorable environment with low shear, high SST (Fig. 6.3a, and b). At the same time, a tropical upper tropospheric trough south of the system may have enhanced its equatorward outflow.

On 16 September, it became a named Typhoon, Fanapi, with a consolidating low-level circulation with an eye feature and convective banding over the eastern semi-circle (Fig 6.2b). After the shortwave trough propagated eastward into central Japan, the subtropical ridge started to steer Fanapi back toward west-northwest with translation speed of 4 kts. Throughout 16 September, Typhoon Fanapi intensified by 20 kts (10 hPa) to 75 kts (960 hPa). The system continued intensifying to 95 kts (950 mb) on 0000 UTC 18 September. Meanwhile, Fanapi had an nearly complete eyewall with a small gap in the north quadrant and a multiple convective bands over the eastern semi-circle (6.2c). Fanapi reached its maximum intensity of 105 kts and 944 hPa on 0600 UTC 18 September and was then at its steady-state stage for about 21 hours. On 19 September, Fanapi made landfall in Taiwan with the intensity of 115 kts (945 mb).

Fanapi developed and intensified within the observational array of ITOP. Because of its slow movement and strong winds, Fanapi induced a significant amount of SST cooling. It is the best-observed typhoon over the Western North Pacific with unprecedented airborne observations of the atmosphere and the ocean from the Air Force C-130 aircraft.
Figure 6.2: (a-c) Microwave satellite estimated rain rate overlaid with the infrared image for Fanapi from late 14 September to the beginning of 18 September. (d) the Polarization Corrected Temperature (PCT) overlaid with the visible image at around 0600 UTC 18 September. Detail time of each swath is labeled on the left-upper corner of the plot.
Figure 6.3: (a) The initial SST (°C) field. (b) The initial T100 (°C) (calculated based on Price 2009) field. (c) TMI/AMSRE SST (°C) swath from 14 to 20 September, which presents the minimum SST during this period. The black line is the JTWC best-track. The diamond symbol denotes the observed storm center at 0000 UTC each day from 15 to 20 September and the gray circle indicated the storm center at 1200 UTC 14 September. (d) Similar to (c) but for the AO simulation. Because the ocean does not change in UA, the SST swath in UA is the initial SST.
6.2 Numerical experiments

Although ITOP had provided a rich set of air-sea observations that are believed to be able to convey the spatial variability of the coupled BL, they are still limited in time and space. In this regard, we conduct the coupled model simulations for the purposes of 1) examining the time evolution of the BL characteristics and the physical processes associated with it, and 2) examining the ability of the couple model to reproduce the observed coupled air-sea BL variability.

Here, 5.5-day simulations of Typhoon Fanapi from the uncoupled WRF model (UA) and the coupled UMCM-WP (AO) are conducted. WRF version 3.2.1 is configured with triply nested domains with 12, 4, and 1.3 km horizontal grid spacing, respectively. The two inner nests are storm-following moving domains. The simulations start at 1200 UTC 14 September with a single 12-km horizontal resolution domain spinning up for the first 12 hours. Thereafter both the 4-km and 1.3-km moving domains are utilized until the end of the simulations. The number of grid points in the 12-, 4-, and 1.3-km domains are 600 X 445, 502 X 502, and 301 X 301 respectively.

WRF is initialized with National Centers for Environmental Prediction (NCEP) final (FNL) fields initialized at 1200 UTC 14 September. The PWP is initialized by the global Hybrid Coordinate Ocean Model (HYCOM) real-time output on 13 September blended with real-time satellite (TRMM TMI/AMSRE) SST on 13 September as well. We apply the Laplacian smoothing function from the edge of the continent to fill in the data gaps. By using the real-time HYCOM field blended with the satellite SST, the initial ocean fields portray the ocean features appropriately and are believed to be close to the reality. Throughout the whole simulation, the 6-hourly nudging of zonal and meridional winds is performed on the 12-km domain using the NCEP analysis to ensure the large-scale steering flow for the simulated Fanapi is close to reality, and therefore the track of the simulated storm is close to the best-track.
6.3 Simulated track and intensity

The simulated storm tracks, with the benefit of nudging, are very close to the best track data with the correct timing, except for the first 12 hours, when there is only an ill-defined center for both simulated storms and for the observed one (Fig. 6.1a). With the well-simulated storm tracks, storms in both simulations have similar evolution trend as the observed one (Fig. 6.1b and c). The difference of the MSLP and MWS between UA and AO becomes evident since 17 September when the simulated SST cooling starts to have significant impact on surface heat fluxes (Fig. 6.4). UA reaches steady state on 0600 UTC 18 September as a Category-4 Typhoon with MSLP of 920 hPa and MWS of 120 kts, approximately 25 hPa and 20 kts deeper and stronger than the best track. AO, with the ability to simulate the ocean response adequately, corrects the over-predicted storm intensity and creates a Category-3 storm with both MWS and MSLP almost right on top of the best track. Within an area of 500 km radius of the storm center, AO induces 1 °C SST cooling and causes approximately 150 W m\(^{-2}\) reduction of surface enthalpy fluxes when compared to UA at steady state.

A noticeable feature in Fig. 6.1b is the spike of MWS in UA at 2000 UTC 18 September, which indicates that UA rapidly intensifies with MWS increasing 40 kts within an hour. At this particular time, however, the MSLP does not decrease, implying that the storm is not really intensifying. By analyzing the evolution of MWS every 5 minutes from 1800 to 2000 UTC (not shown here), we found a dramatic increase of MWS from 138.3 kts to 159.6 kts in 20 minutes from 1940 to 2000 UTC. A horizontal surface wind map focusing on the eyewall region (Fig. 6.5c, and d) shows this sharp jump in MWS as a result of the passage of eyewall mesoscale vortices. The meso-vorticies are finger-like features with a length scale of approximately 5-10 km and has extreme local wind. In Fig. 6.5d, the meso-vorticies in UA cover only 4 girds, about 3 km in radius. Via examining the MWS every
Figure 6.4: Time series (a) SST, (b) enthalpy fluxes (W m$^{-2}$), (c) latent heat fluxes (W m$^{-2}$), and (d) sensible fluxes (W m$^{-2}$) averaged over the area withing 500 km distance from the storm center.
Figure 6.5: 10-m wind speed in (a-b) AO and (c-d) UA at around 2000 UTC 18, when there is a dramatic oscillation in MWS. The green “x” indicates the location of the MWS whose value is given each panel. The number in the parenthesis is the number of grid points with the wind speed larger than the MWS - 10 kts in 1.3-km resolution domain, indicating the size of the features with extreme winds.
Figure 6.6: Radar reflectivity of Fanapi at (a) 1450 UTC 18, (b) 1630 UTC 18, and (c) 1830 UTC 18. The magenta circles indicate the possible meso-vorticies. (Images are from Taiwanese Central Weather Bureau)

5 minutes in AO, we found a similar feature occurring on the east side of the eyewall in AO at 2010 UTC 18 September (Fig. 6.5a, and b).

Historically, meso-vorticies have been observed and documented with both radar and flight-level data by Aberson et al. (2006) in Hurricane Isabel (2003) and Marks et al. (2008) in Hurricane Hugo (1989) respectively. The numerical study by Nolan et al. (2009) also showed these eyewall vortices maximums in an Isabel (2003) simulation. Montgomery et al. (2006) argued that these eyewall meso-vorticies act as an additional heat sources to the eyewall, and may locally increase the convective instability.

Although this study does not focus on these meso-vorticies, we are curious to know if similar features exist in reality. The Taiwanese due-Doppler radar images of Fanapi at 1450 UTC, 1630 UTC, 1830 UTC 18 September (Fig. 6.6) exhibit the small-scale local radar reflectivity maximum (magenta circles) at the inner edge of the eyewall, which is an indication of the meso-vorticies as suggested in Aberson et al. (2006). Even though there were meso-vorticies in the reality, it is not easy to catch the local extreme wind maximum in observations and therefore the best track. However, if so, some dramatic oscillation of MWS in best track, as shown in high-resolution numerical simulations, would be expected (although it is possible that it would not be included in the best track because it could not be representative).
Figure 6.7: (a) Observed SST from the AXBTs co-located with the dropsondes. Star symbols indicate the soundings that are located in the areas with SST less than 28 °C, roughly the location of the cold wake. The black line is the best track from JTWC and diamond symbols show the center of storm at 0000 UTC each day from 15 September. (b) Similar to (a), but for the simulated soundings sampled at the same storm-relative locations in AO.

From the whole simulation, we focus on the period of 0000 UTC to 1200 UTC 18 September, when both observed and simulated storm are approaching and reaching their respective steady states and maximum intensities. Additionally, this is also the period when there are co-located air-sea soundings from ITOP observations.

6.4 Oceanic response

The improvement of simulated storm intensity in AO illustrates the crucial impact of SST cooling on storm intensity. While the atmospheric forcing is critical to the size and strength of the SST cooling, the pre-storm oceanic condition and the evolution of subsurface oceanic stratification plays an important role as well. In this section, we examine the observed and simulated ocean response that includes the SST cooling and subsurface ocean response. Because there is no oceanic response in UA, here we present only the results from AO.
From AXBT observation, SST at the very front side of the storm are above 28.5 °C (reddish) and it is below 27.5 °C (bluish) in the rear-right quadrant, revealing the location of the cold wake (Fig. 6.7). In addition to AXBTs, ADOS also measures SST. Furthermore, it provides the evolution of SST from pre-storm to post-storm. Other than SST, ADOS measures 10-m winds and sea-level pressure (SLP). In total, there were 8 drifters and Fanapi was expected to pass the drifters on 0000 UTC 18 September (Fig. 6.8). At the time of storm passage, the storm-relative locations of ADOSs on the right side of the storm were 143 km (blue), 102 km (light-blue), 51 km (green), and 21 km (light green). On the left side, they were 163 km (brown), 115 km (pink), 70 km (red), and 31 km (yellow). The pre-storm SST on the right side of Fanapi was 29.5 °C, about 0.3 °C warmer than the one on the left side. Right after storm passage, SST on the right side immediately dropped. For two ADOSs (light green and green) that are at the storm center/in the eyewall, SST dropped to approximately 26.5 °C. For the one further outside (light blue), SST dropped to 28 °C. On the left side, SST in the eyewall dropped to 28 °C (yellow). SST in the other three ADOSs shows only about 0.5 °C reduction. During the whole four days, there was no clear recovery trend of SST.

The evolution of subsurface oceanic structure is studied via using EM-APEX floats. Again, Fanapi was expected to pass the floats on 0000 UTC 18 September. Figure 6.9 shows the trajectories of all 7 EM-APEX floats overlaid the TMI observed rainrate. Since there was no good satellite coverage on 0000 UTC 18 September, we use TMI image at 0654 UTC to demonstrate the floats location relative to the storm center. From right to left relative to the storm motion, the float numbers are #4391d, #4912a, #4907a, #4910a, #4906a, #4909a, and #4914a. #4910a was in the eye of Fanapi, about 20 km from storm center, while #4907 and #4906 were underneath the eyewall, 44 and 53 km from the storm center respectively. The time series of ocean temperature and mixed layer depth from all floats except #4914a (that reported only one-day data) are in Fig. 6.10.
Figure 6.8: (a) The trajectories of ADOS drifters from the dropping time, around 0000 UTC 17 September, to 0000 UTC 20 September. (b) Similar to (a), but for simulated drifters in AO. Time series of the observed and simulated (c-d) sea-level pressure, (e-f) 10-m wind speed, and (g-h) SST.
In the pre-storm (before 0000 UTC 18 September) period, the mixed layer depths of all floats are around 50 m. The ocean temperature is warmer by more than 1 °C on the right side than the left side of the storm (compared to #4391d to #4909a). The salinity profiles also exhibit left to right asymmetry in vertical stratification in the pre-storm conditions in which there is a stronger vertical stratification on the left side than on the right side (Fig. 6.11). After the storm passage, the storm-induced cooling decreases the mixed-layer temperature by more than 2 °C in the eye and at right side of the eyewall (#4910a and #4907a). Outside of the eyewall, the float to the right of the storm track (#4319d and #4912a) have around 1 °C cooling while that to the left-side (#4909a) has cooling less than 1 °C. Although it is well known that the strong cooling on the right-side is because of the alignment between the changes of wind and current direction (Price 1981), the high stratification in the pre-storm ocean condition on the left might also help to increase the right-left asymmetry of the ocean after storm passage.
Figure 6.10: (a) The upper ocean temperature as a function of time and depth from EM-APEX floats. The location of each float is shown in Fig. 6.9. The thin black contours indicate the temperature isotherms of 29, 28, 26, 24, 22, and 20 °C. The black dots mark the mixed-layer depth of each profile. (b) Similar to (a), but for the model sampled floats in AO.
Figure 6.11: Similar to Fig. 6.10 but the color now indicates the upper ocean salinity.
To analyze the oceanic structure in the coupled model, we sample both simulated AXBTs, ADOSs and EM-APEX data in AO, as shown in Figs. 6.7, 6.8, 6.9 and 6.11. Geographically, they are all sampled based on storm-relative locations. The simulated-AXBTs are sampled from the model output at 0000 UTC 18 September. The simulated floats and drifters are sampled continuously from 17 September to the end of the simulations based on the recording times from ADOS and EM-APEX respectively. The comparison of SST between AO and observations indicates that the simulated cooling extends further south (rear-left side of the storm track). Over the rear-left quadrant, the simulated SST in AO is about is about 0.5 °C colder than the observed SST from AXBT and is about 0.5 to 1 °C colder than that from ADOSs. In the contrast, the SST on the right side of the storm is well simulated. Figure 6.12a is a scatter plots of SST from ADOS versus that from AO The linear-regression between observed and AO SST nearly matches the one to one line (the slope is 1.05). On the other hand, Fig. 6.12b shows the unrealistic SST UA due to lacking of the ocean coupling.
Below the sea surface, there is no left-right asymmetry in neither temperature nor salinity in pre-storm ocean condition in AO. After storm passage, the reduction (increase) of ocean temperature (salinity) from these sampled floats is very close to that was observed, as well as the change of the mixed layer depth. However, AO seems to induce stronger cooling on the left side of the storm than the real storm did, similar to what we found in SST fields. The possible explanation for this bias could be the lack of the left-right asymmetry in the pre-storm oceanic stratification, which limited the cooling on the left side in reality.

6.5 Spatial distribution of surface winds and air-sea fluxes

6.5.1 Observed and simulated winds

Figure 6.8 has shown that the magnitude of the 10-m wind of the simulated drifter is close to the observed one. Fig. 6.13 further depicts the horizontal distribution of the simulated 10-m wind speed from AO and UA overlaid with the observed GPS dropsondes and SFMR. Again, all the wind analyses presented here are Earth-relative but the locations are relative to the storm center.

For Mission #1, the observed wind is relatively stronger in the eastern eyewall (the reddish squares and stars) and is stronger at northeast quadrant at the outer core (~100 km from the storm center, yellowish squares). This surface wind asymmetry is possibly associated with the convective organization. As shown in Fig. 6.2b, there is more contiguous convection south-east of the center in the eyewall and south of the center in the outer core. These convective asymmetries could be a consequence of north-northeasterly shear at this particular time (from Statistic Typhoon Intensity Prediction System, STIPS) as argued in Chen et al. (2006). Another possible cause of the wind asymmetries is the interaction between storm motion and the earth surface (Shapiro 1983; Kepert 2001). However, the wind asymmetries here do not agree with the storm motion theory in which the wind is stronger
Figure 6.13: 10-meter winds from AO overlaid with the surface winds estimated from SFMR (square) and the 10-meter winds estimated from GPS dropsondes (stars) from (a) Mission #1 and from (b) Mission #2. (c) and (d) are analogous to (a) and (b), but from UA. The model time for Mission #1 is 0000 UTC 17 and is 0000 UTC 18 September for Mission #2, approximately the time of eyewall penetrations for both missions.
Figure 6.14: Radial distribution of observed (black line: SFMR, red dot: dropsondes, purple dots: ADOS) and simulated (gray) 10-m winds during (a, c) Mission #1 and (b, d) Mission #2. The simulated winds in (a) and (b) are from AO while they are from UA in (c) and (d). The model field is from 0000 UTC 17 September for Mission #1 and from 0000 UTC 18 September for Mission #2, approximately the time of eyewall penetrations for both missions.

in the front side of the storm. The relative importance of the asymmetries caused by storm-motion and by the convection (or vertical shear) requires detail analysis, and is beyond the scope of this study. Nevertheless, the distribution of winds in Fanapi implies that the impact of the asymmetric convective organization dominates.

The asymmetric structures of surface winds in two simulations are somewhat similar to each other and to observation. Both of them show the strongest wind in an elongated region from the east to southeast side in the eyewall. Outside of the eyewall, the strongest wind is in the northeast quadrant. Because of the similar vertical shear direction in simulations, the simulated wind asymmetry is close to the observations.
Figure 6.15: Scatter plots of 10-m wind speed between observations and simulations in (a) AO and (b) UA. The black solid line is one to one line while the dashed line indicates the linear regression line.

In terms of the magnitude of winds, during the duration of Mission #1, AO fits SFMR data well but is slightly weaker in the eyewall. UA is in general too strong, but it captures the local wind maximum in the eyewall at southeastern quadrant that was observed by dropsondes. When comparing 10-m winds from simulations and the observed winds (from SFMR, dropsondes, and ADOS) as a function of radius (Fig. 6.14), we notice that the winds in the outer core in UA are too strong. Strong wind in the outer core indicates larger angular momentum in the outer area. With the friction-induced BL inflow, this extra momentum in UA will be carried into the storm core area, and help spin up the storm. This might contributes to the intensification rate which is larger in UA after 0000 UTC 17 September compared to that those in AO and in reality(Fig. 6.1). What this means is that the essential corrections of the ocean coupling includes not only the correction of MWS and MSLP, but also the increasing accuracy of the winds outside of eyewall and the radial gradient of the surface winds.
In Mission #2, Fanapi was contracting and the RMW shrank from 50 km to about 20 km (Fig. 6.14). The maximum wind speed occurs again east of the storm center and the winds is stronger in the northeast quadrant outside of the eyewall. Again, the wind asymmetries here is likely related to the convection organization. Comparing AO and UA to the observations, neither simulation captures the eyewall contraction; they both have RMWs of 50 km, more than twice as large as observed (Fig. 6.14b, d). In terms of the asymmetric distribution, both storms seem to have stronger winds in the eastern portion of the eyewall and in the northeast quadrant in the outer area. Because of the larger eye, they over-predict the winds from 20 to 50 km in radius. Outside of the simulated RMW (50 km), the winds in both simulations are also larger than the observed winds. This might be another reason why the cold wake in AO is broader than that in reality. In general, both simulations capture the magnitude of the 10-m winds well but UA has systematically positive bias (Fig. 6.15). However, because of the large eye in simulations, both of them underestimate the strong 10-m winds at the observed eyewall location.

6.5.2 Observed and simulated air-sea fluxes

Figure 6.16 exhibits the spatial distribution of observed and simulated surface sensible and latent heat fluxes during Mission #2. The observed fluxes are calculated from the co-located dropsondes and AXBTs based on the bulk COARE 3.0 (Fairall et al. 2003) algorithm for low wind regime (\(\leq 15 \text{ m s}^{-1}\)) and on the bulk formula with the exchange coefficient from CBLAST observations (Drennan et al. 2007) for the high wind regime (\(>15 \text{ m s}^{-1}\)). The sounding from the cold-wake region is marked with star symbols. There is a clear reduction of both sensible and latent heat fluxes over the cold wake region. As Fig. 6.13 shows, there is only a weak spatial asymmetry in the surface winds during this particular period, and therefore the asymmetric distribution of surface fluxes is a consequence of the cold wake. Another interesting point shown here is a region of negative sensible heat fluxes, which indicates that the SST is colder than air temperature. Although there is only 1 sounding
Figure 6.16: (a) Sensible heat fluxes estimated from co-located dropsondes and AXBT. (b) The model-sampled sensible heat fluxes in AO. The model field is sampled based on the storm-relative location of each sounding. (c) Same as (b) but for UA. (d-f) Similar to (a-c) but for latent heat fluxes. The black arrow indicates the storm motion. Stars indicate the soundings with SST less than 28 °C, as shown in Fig. 6.7.
with negative sensible heat fluxes in Fig. 6.16a, there are actually 3 more soundings with negative sensible heat fluxes locate more than 200 km from storm center, and therefore they are not shown here. The effect of these negative heat fluxes will be discussed in detail in the next subsection.

Similar to the observed data, the surface fluxes in AO are smaller over the cold wake region, and there is also a negative value of the sensible heat flux (again, only one of them within 200 km radius of the storm center). Nevertheless, the magnitude of the surface fluxes in AO is clearly higher than the observed flux, and this is mainly due to the over-estimated wind speed and larger eye shown in Fig. 6.14. Other source of the errors in simulated fluxes are the errors in the air-sea temperature and moisture differences (Fig. 6.17). Without the storm induced cold wake, the asymmetric surface sensible fluxes in UA simply reflects the wind asymmetry. The root mean square errors (RMSE) of the parameters that used for flux calculation are shown in Fig. 6.16. As expected, the RMSE of 10-m wind, air-sea temperature and moisture difference in UA is larger than that in AO. Among the three parameters, the air-sea moisture error is improved the most by adding the ocean coupling, from 2.3 g kg\(^{-1}\) in UA to 0.8 g kg\(^{-1}\) in AO. With the improvements of all these parameters, AO reduces the error in the surface fluxes by more than 50%.

### 6.6 Spatial variability in TC and oceanic structures across the cold wake

Results from Chapters 4 and 5 provide an essential idea, which is that the storm-induced asymmetric ocean response, i.e. the cold wake, feeds back on to the BL structure. To examine the spatial variability of the BL across the cold wake, we first divide all of the co-located soundings from Mission #2 into four groups: 1) cold-wake soundings, for those released over the rear-right quadrant and that have SST lower than 27.5 °C; 2) upstream soundings, for those released upstream of the cold-wake and in the area of the strong SST
Figure 6.17: Scatter plots for the parameters that are used in bulk algorithmic flux calculation in observations versus that in simulations (red-AO and blue-UA): (a) 10-m wind speed (ms$^{-1}$), (b) air-sea temperature difference ($^\circ$C), (c) air-sea moisture difference. (d) and (e) shows the scatter plots for latent and sensible heat fluxes (Wm$^{-2}$). The root-mean-square error between observations and simulations for each parameter is printed on each plot.
Figure 6.18: Horizontal map of microwave satellite brightness temperature overlaid with the location of the co-located soundings. Colors of squares indicate sounding locations, green: upstream, blue: cold-wake, orange: downstream, and gray: others. Skew-T diagrams are from the numbered soundings indicated on the horizontal map. The observed time in HHmmSS, as well as the distance from the storm center, is labeled on the top of each plot.

gradient; 3) downstream soundings, for those released downstream of the cold-wake and in the area of strong SST gradient; and 4) others, for those that are not included into above three groups, which are usually in the front side half of the storm (Fig. 6.18). All of the upstream, cold-wake, and downstream soundings are at least 50 km away from the storm center to ensure that they are not in the eyewall. Although the dropsondes are not released at the same time, we assume that time evolution of the BL is not as significant as the spatial variation within the observation period. Furthermore, most of the soundings in the three relevant groups are released between 0000 UTC and 0300 UTC 18 September, except sounding 5. Therefore, we treat all soundings as if they occurred at the same time and focus only on the spatial variations. Because the analyses in the previous three sections
have shown that the storm intensity, structure, and the characteristics of the air-sea interface in AO are more accurate than that in UA, the analyses below will mainly focus on AO. The simulated soundings are grouped relative to the cold wake in AO with the same criteria that is used for observed soundings (Fig. 6.19). All the soundings are sampled from the model output at 0000 UTC 18 September. Because the horizontal shape of the cold-wake in AO is different to that in reality, especially in the southeast quadrant (rear-left quadrant of the storm track), some simulated soundings might be classified differently to the analogous observed soundings. (e.g. the soundings 3 and 4).

6.6.1 Thermodynamic structures and its relation to the distribution of convection

While air-sea interaction has a direct and substantial effect on the BL structures in TCs, the effect of the large scale atmospheric forcing and of convection associated with rainbands
can not be ignored. For example, some soundings are at the leading edge of a rainband while some of them are in the rear side of convection, which might result in varying atmospheric conditions (Zipser 1977; Barnes et al. 1983; Houze 2004) and complicate our analysis. Given the fact that there is no easy way to separate the impact of the atmospheric convection from the effect of the air-sea interaction, the criteria used to group soundings contain no information about the atmospheric conditions. However, understanding the atmospheric condition can help us to interpret the results and examine the robustness of our findings.

To relate each sounding to the distribution of convection, the thermodynamic structure of soundings is examined using composite satellite observations (MIMIC) and skew-T diagram (Fig. 6.18). The composite satellite observation in Fig. 6.18 is consist of various satellite data near 0000 UTC 18 September. Therefore some of the features might not occur during our observation window. However, when comparing the satellite image in Fig. 6.18 to Fig. 6.2c, the resemblance of these two suggests that the eyewall and primary rainband in Fig. 6.18 is representative. The outer rainband, which was eroding from 0000 to 0600 UTC, might be exaggerated in the composite image. On the other hand, the Skew-T diagram shows the detail thermodynamics of each sounding. All the numbered soundings are outside of the primary rainband. Except for sounding 10, which is 87 km from the storm center, all of the soundings are at least 120 km from the storm center. Identical analysis for AO is shown in Fig. 6.19.

For the ITOP observation, the three upstream sounding (1 - 3) are at the leading edge of the primary rainband. There are “onion” type of profiles in soundings 1 and 2; it is uniformly dry throughout the whole column in sounding 3. Above 850 hPa, the temperature lapse rates for these three soundings are close to the moist-adiabatic lapse rate, indicating the occurrence of convection in the past. The variability among the cold-wake soundings is greater than the upstream soundings. The nearly moist-adiabatic temperature lapse rate and the little dew-point depression of the sounding 4 imply the existence of convection
more recent. Meanwhile, the dry BL might be directly associated with the cold wake. The dry pattern in Sounding 5 indicates that may not have been affected by convection for a while. The clear mid-level inversion layers in soundings 6 and 7 are indicative of mesoscale downdraft, which brings down the dry air from mid-troposphere that may enhance the low-level warming. Another possible source of the warmth of BL of the sounding 6 is the upstream warm air. The near-saturated and near-moist adiabatic profiles for sounding 8 – 11 indicate that the soundings have recently been in the rain convection. The two downstream soundings (12 and 13) are in a dry environment.

The simulated storm in AO captures the overall convective asymmetries, especially the rainband convection on the east side. However, there is no clear separation between the primary rainband and the outer rainband (Fig. 6.19). The three upstream soundings are again in the convective-scale downdraft at the edge of the primary rainband. The temperature lapse rate of all of them implies convection in their history as what was found in the observed soundings. The first cold wake sounding, sounding 4, is in the stratiform cloud area associated with the primary rainband, and there is a mesoscale downdraft below the stratiform cloud base. Soundings 5-7 are in the subsidence area similar to the observed soundings. The remarkable similarity between sounding 5 in AO and sounding 6 in ITOP data is because both of them are in the environment that is dominated by a mesoscale downdraft and upstream warm air. Soundings 8-11 are in the convective zone associated with an outer rainband, and there is low-level subsidence occurring at sounding 9. Sounding 12 is in the subsidence zone while sounding 13 in AO is in low clouds. Overall, these simulated soundings exhibit atmospheric conditions similar to what is observed.

In both ITOP observations and AO, the upstream and downstream soundings are in the area with an evidence of a history of previous convection. The cold-wake soundings are in conditions indicative of a combination of a convective and mesoscale descent environment. Some of the BL features in the skew-T diagrams are related to their locations relative to the cold wake, and this will be addressed next.
6.6.2 Simultaneous temperature profiles across the boundary layers

The observed and simulated temperature profiles across the air-sea interface from ocean to atmosphere are shown in Fig. 6.20. The estimated air-sea parameters at the interface (SST, 10-m wind, air-sea thermal disequilibrium, and air-sea fluxes) are shown in Figs. 6.21 and 6.22. The observed SST decreases from 28.25 °C in the upstream region to 25.8 °C at sounding 5 in the cold wake. Then it gradually increases to 28.5 °C in the downstream soundings. The oceanic mixed layer temperature shows a similar evolution to that of SST. In the typical marine BL (not in TCs), the air temperature is lower than SST as what we found from the upstream and downstream soundings. However, over the cold wake area, the airflow maintains its upstream properties while the ocean is colder relative to upstream. Therefore the air can actually be warmer than the SST (soundings 5, 6, 7 in Fig. 6.20a), indicating downward (negative) sensible heat fluxes (Fig. 6.21c, e). A similar feature is also seen in the Fig. 6.16a. Although the SST is the lowest in sounding 5, the warmer near surface temperature in sounding 6 (Fig. 6.20a) results in the largest negative sensible heat fluxes. The near-surface warmth of sounding 6 might be due simply to horizontal advection of warm air from upstream, with this air remaining unmodified by convection. Alternatively, the warmth could instead be due to vertical advection from a mesoscale downdraft. The moisture disequilibrium decreases as well from upstream to cold wake region due to the low SST. The lowest moisture disequilibrium and the latent heat fluxes occur at sounding 8 because of the near saturated BL air due to the convection (Fig. 6.18). While the air continuous to travel over the cold wake, it becomes successively colder and drier because the low surface fluxes (sometimes negative for sensible heat fluxes). Then, when the airflow moves back over the warm ocean, i.e., at downstream of the cold wake, thermal and moisture air-sea disequilibrium are enhanced. The resulting sensible and latent heat fluxes are both greater here than those upstream of the cold wake.
Figure 6.20: (a) The simultaneous air-sea temperature profiles for the upstream, cold-wake, and downstream soundings. (b) Similar to (a) but for AO simulated soundings. The numbers in the x-axis correspond to the numbers that labeled in Figs. 6.18 and 6.19 for (a) and (b) respectively.
Figure 6.21: Air-sea interface parameters estimated by co-located soundings: (a) SST (°C), (b) 10-m wind (m s\(^{-1}\)), (c) air-sea temperature difference (°C), (d) air-sea moisture difference, in which the positive value means SST is warmer than air temperature, (e) sensible heat fluxes (W m\(^{-2}\)), and (f) latent heat fluxes (W m\(^{-2}\)). Colors indicate the location of soundings: upstream (green), cold-wake (blue), and downstream (orange). The numbers in the x-axis correspond to the numbers that labeled in Figs. 6.18.
Figure 6.22: Similar to Fig. 6.21, but for AO sampled soundings. The numbers in the x-axis correspond to the numbers that labeled in Figs. 6.19.
The simulated soundings are remarkably similar to the observed one (Figs. 6.20b and 6.22). In AO, two soundings have negative sensible heat fluxes. The lowest SST occurs in the sounding 7 while the smallest (most negative) sensible heat flux is in sounding 5 due to the warmer air temperature (Fig. 6.20b, and Fig. 6.19), similar to the observed sounding 6. As in the observations, the smallest latent heat fluxes in the simulation occur at sounding 8. However, it is a consequence of not only the low SST and the nearly saturated atmospheric conditions, but also the weak wind speed. Again, both sensible and latent heat fluxes are enhanced downstream of the cold wake, although the enhancement is larger compared to the observations.

6.6.3 Boundary layer height and stability

The temperature profiles and the skew-T diagrams in Figs. 6.18 and 6.19 have shown an indication of the BL height and stability of each sounding. For the purpose of describing the results quantitatively, we define the BL as mixed layer (THBL in Chapter 4). The top of BL is defined as where the $\theta_v$ is 0.5 K larger than its surface value and we define the BL stability as the static stability (Eq. 5.3). Figure 6.23 shows the vertical profiles of $\theta_v$ from the ITOP and AO soundings. Except sounding 6, the BL height decreases dramatically from about 500 m in the upstream region to ~150 m in the cold wake. The BL height then increases back to around 500 m in the downstream region. In addition to BL height, the vertical gradient of $\theta_v$, the BL stability, shows dramatic changes.

Among the three upstream soundings, only one of them (sounding 3) shows the near-surface unstable layer, while the others have a well-mixed layer down to the surface. The mean vertical gradient of $\theta_v$ in the BL is 0.43 K km$^{-1}$ in this group, indicating a neutral to stable BL. In the cold-wake group, all the soundings, except sounding 11, have only a stable layer, and the mean $\theta_v$ gradient increases to 2.43 K km$^{-1}$. For downstream soundings, there is a distinct unstable layer because of the cold air flowing over the warm water and the associated enhanced surface heat fluxes discussed above. The mean vertical $\theta_v$ gradient
Figure 6.23: The virtual potential temperature from (a) observed soundings and (b) AO sampled soundings. Colors indicate the sounding location: upstream (green), cold-wake (blue), and downstream (orange). The height of the colorful squares indicates the BL height. The black squares indicate the tick of 305 K for the corresponding sounding. The numbers in the x-axis correspond to the numbers that labeled in Figs. 6.18 and 6.19 for (a) and (b) respectively.
in this group is \(-1.54\) K km\(^{-1}\). The BL stability analysis for the ITOP soundings support the findings from Chapter 5 for the SBL formation over the cold wake.

The simulated soundings show features similar to the observed ones but the vertical stratification is not as strong, likely due to the low vertical resolution. The \(\theta_v\) gradients for upstream, cold-wake, and downstream are \(-0.06, 0.85,\) and \(-1.06\) K km\(^{-1}\), which also indicate the BL stability changes from neutral to unstable in the upstream region, and to stable in the cold wake region to unstable in the downstream region.

### 6.6.4 The dynamic characteristic of the surface airflow

Chapter 5 concludes that the dynamic impact of SBL over the cold wake is the surface wind reduction. From the ITOP observations, Figure 6.21 shows signs of wind reduction from upstream to the cold wake. However, those soundings are located at various radii and have different convective conditions. It is therefore hard to distinguish causes of the wind reduction (due to cold wake or due to other reasons). To avoid being influenced by the radial and convective dependence of the local wind speed, we examine the feature of the wind reduction by using the inflow angle (same as what we did in Chapter 5).

Figure 6.24a shows the composite SST and wind vector from 8 ADOS data (Fig. 6.8) during the period of 1800 UTC 16 to 0000 UTC 19 September. While the mean inflow angle over the whole area (covered by the data but outside of 50 km in radius) is 21.0\(^{\circ}\) with 90% confidence interval from 19.8\(^{\circ}\) to 22.1\(^{\circ}\), the inflow angle over the cold wake (blueish area in SST map) is as large as 43.5\(^{\circ}\) with 90% confidence interval from 42.1\(^{\circ}\) to 44.9\(^{\circ}\). Figure 6.24b shows that the high value of inflow angle occurs not only over the cold wake, but also upstream of the cold wake. In Powell (1982) and Shapiro (1983), the strongest inflow angle in the part of the rear side of Hurricane Frederic (1979) is addressed due to the storm motion, which induced the a confluent flow over the rear side of the storm. Furthermore, inflow angle also depends on other complex factors, such as storm intensity and the radial distance from storm center, etc. (Zhang and Uhlhorn 2012). Hence, it is not
Figure 6.24: (a) Observed SST (shading) and wind vectors from composite ADOS drifter data from 1800 UTC 16 to 0000 UTC 19 September. (b) Similar to (a) but the wind vector is color-coded based on the inflow angle. The storm motion is indicated by the thick black arrows.
easy to conclude that the higher inflow angle over the cold wake area is purely due to the SBL from observations. In this regard, the comparison of inflow angle between UA and AO is useful here. To conduct the same analysis on the model field, the simulated drifter data (Section 6.4) are used to calculate inflow angle (Fig. 6.25). Qualitatively, the wind vectors in UA over the rear-right quadrant are more parallel to the tangential direction than those in AO. The mean inflow angle over the cold wake in AO is about 25.2°, which is twice as large as the mean value over the whole domain (14°). In UA, however, the mean inflow angle decreases slightly from 17.0° over the whole domain to 16.8 over the rear-right quadrant. This is because the higher inflow angle in UA is at rear-left quadrant rather than in the rear-right quadrant, like what we see in AO and the observed fields. Because Figs. 6.24, and 6.25 are sampled over about two days, the observed wind directions (as well as simulated one) in the front two quadrants were measured between 1800 UTC 16 to 0000 UTC 18 September while those in the rear two quadrant were measured after 0000 UTC 18 September. During these two days, Fanapi was intensifying. Thus, we suspect that our results could be contaminated by the storm intensities. Therefore, we calculate the mean inflow angle from the model output at 0000 UTC 18 September over the whole simulated storm domain (400 km in radius) and over the rear-right quadrant in AO and UA (Fig.6.26). While the magnitude of the inflow angle varies a little bit, the increase of inflow angle over the rear-right quadrants again occurs in AO only.

Although, it is clear that the enhanced inflow angle over the rear-right quadrant occurs only when ocean coupling is considered, the simulated inflow angle in both AO and UA are much smaller than the observed value. To address this discrepancy, we should consider the uncertainty associated with both the observed and simulated wind directions. In ADOS, the wind direction are measured at a precision of 5°. From deployment tests of ITOP, the wind direction difference between two drifters within 100-300 m of each other is around 8°, and this is the typical instrument noise. Comparing the mean inflow angle at different quadrants obtained from ADOS to that from dropsondes, the ADOS shows about 5 to 10° positive
Figure 6.25: Similar to Fig. 6.24, but for the simulated drifter data in (a,c) AO and (b,d) UA.
bias, which is close to $8^\circ$ and indicates ADOS is relatively reliable. The comparisons between ADOS wind direction and other co-located scatterometer data in other TCs shows the average bias of ADOS is less than $15^\circ$, but the value can be as small as $2^\circ$ and as large as $30^\circ$. The large bias values ($15\sim30^\circ$) are mostly due to the fact that the ADOS take measurements at a point, while the scatterometer reports area averages (at least 50x50km). For our purposes, the quality-controlled ADOS data (which is used here) has the instrument noise removed and is reliable to use (Dr. Jan Morzel, personal communication). The reason why the model simulated inflow angle is systematically low requires further research and will not be addressed here.
Figure 6.27: The initial location of the tracers (light-blue contours) and trajectories (dots) in (a) AO, and (b) UA. In AO, the trajectories are released in the upstream (green), cold wake (blue), and upstream (orange) area with the same criteria use in Section 6.5. In UA, tracers and trajectories are released in the same storm-relative location. The gray shading is the radar reflectivity (dBZ) and the red contours show constant SST (°C).

6.7 The modulation of TC energetics due to the cold wake

In this section, we try to follow the energetic variation of the air parcels as they moving across the cold wake. However, it is impossible to do the Lagrangian analysis using the “in situ” observations. Because of the limited temporal sampling of the observations, we study this issue with a Lagrangian forward trajectory and tracer analysis in AO. The underlying assumption of using AO instead of observational data is that the storm in AO is representative of Fanapi. This assumption is justified based on the comparison between ITOP data and AO shown in previous sections. For the purpose of comparison, the same analysis is also conducted in UA.

In total, we release 500 trajectories. They are consisted of 145 upstream trajectories, 350 cold-wake trajectories and 105 downstream trajectories (Fig. 6.27). All the trajectories are released from 180 to 300 km from the storm center and at 80 m above the surface (the
Figure 6.28: The equivalent potential temperature along the trajectories released over the (a) upstream of the cold wake, (b) cold wake, and (c) downstream of the cold wake in AO. (d), (e), and (f) are similar to (a), (b), and (c) but for UA.

The initial value of upstream trajectories in UA and AO are both in the range of 360 to 362 K. Both simulations have some upstream trajectories that are ingested into the rainband convection and their $\theta_e$ decrease to less than 350 K (bluish color). For the trajectories in AO that are not ingested into rainband, there is a clear decreasing trend in $\theta_e$ as the air parcels passing over the cold wake. The lowest $\theta_e$ is approximately 355 K at 150 km in radius. After leaving the cold wake, the $\theta_e$ of the air parcels gradually increases because of the enhanced surface heat exchange fluxes. Just before the air parcels enter the eyewall, $\theta_e$ is in the range of 360 to 368 K. From the cold wake to the eyewall, $\theta_e$ along the trajectories in
AO increases by as much as 13 K. In UA, $\theta_e$ along the trajectories those enter the eyewall increases about 10 K maximum on their way to the eyewall. Although the absolute value of $\theta_e$ at the eyewall in UA is ultimately higher than that in AO. The increase of $\theta_e$ from the rear-right quadrant to the eyewall in AO is 3 K higher than that in UA.

Unlike the upstream trajectories, the cold-wake trajectories behave very different in AO and UA. Twelve out of thirteen trajectories in AO stay in the BL, and move cyclonically inward into the eyewall. In UA, however, only four of them enter the eyewall while the remainings are all ingested into the rainband. This is because of the lack of the stabilizing effect of the cold wake. The $\theta_e$ of trajectories in AO increases gradually on their way to eyewall with final values varying from 360 to 368 K, and the largest increase of $\theta_e$ is 13 K. In UA, the trajectories going into the eyewall again has the increase of $\theta_e$ about 10 K. The downstream trajectories in both AO and UA behaves more similarly to each other than do the trajectories initiated in the other two regions. The initial values of $\theta_e$ are in the range of 358 to 360 K in AO and are 360 to 362 K in UA.

For all 500 trajectories, trajectories in UA from the three groups behave similar to each other: some of them are ingested into the rainband while the rest of them stay in the BL, moving cyclonically inward to the eyewall with an increase of $\theta_e$ of up to 8-10 K. In AO, the air flowing from these three locations evolves differently. Some of the upstream and downstream trajectories are affected by the rainband convection, similar to UA. The cold-wake trajectories are stabilized by the SBL, and the majority of them enter the eyewall with the $\theta_e$ increased by 10 to 13 K due to the enhanced fluxes downstream of the cold wake. Upstream of the cold wake, the trajectories that stay in BL and eventually move across the cold wake behaviors similar to the cold-wake trajectories. Downstream of the cold wake, the evolution of the $\theta_e$ is similar to that in UA. Among these 500 trajectories, there are 269 of them that enter the eyewall in AO but only 196 of them in UA.

Similar to Chapter 5, we release tracer over the cold wake area. Vertically, the tracer is released at about 80 meter above the sea surface, and horizontally, it is released over the
area with SST less than 27 °C and between 180 to 300 km radius (Fig. 6.27). The tracer concentration supports what we found from trajectory analysis and from Chapter 5. Hence, we are not repeat the same analysis here.

Combining the results from the trajectory and tracer analysis, we can conclude that with the storm-induced cold wake and its associated SBL, the BL air parcels first lose their high energy due to passing across the cold wake. Then, they regain part of the lost energy on their way from the cold wake to the eyewall through enhanced upward enthalpy fluxes. This process can mitigate the negative ocean feedback by bringing more high energetic air-parcels into the eyewall.

### 6.8 Conclusions

The variation of the air-sea coupled BL across the cold wake and their impact on Fanapi (2010) were examined with ITOP observations and with the air-sea coupled model simulations. The ITOP data included the SFMR, paired GPS dropsondes and AXBTs, ADOS drifters, and EM-APES floats. The horizontal distribution of SST, surface winds, air-sea surface fluxes, and the vertical profiles of thermodynamic fields in BL were studied in detail.

The storm-induced cold wake was observed in AXBT and ADOS SST. The asymmetry in SST led to the spatial variability in surface heat fluxes, which have low sensible and latent heat fluxes because of the low SST. Furthermore, the warm air over the cold wake, causes Fanapi to lose some heat energy to the ocean through the downward sensible heat fluxes. The cold SST resulted in the decreased air-sea moisture disequilibrium as well and therefore the reduction of surface latent heat fluxes. The lowest latent heat fluxes occurred at where the atmosphere was dominated by convection, in which the surface air is nearly saturated. While the air continued to travel over the cold wake, it became successively colder and drier because the low surface fluxes (sometimes negative for sensible heat
fluxes). Then, when the airflow moved back over the warm ocean, i.e., at downstream of the cold wake, thermal and moisture air-sea disequilibrium were enhanced. The resulting sensible and latent heat fluxes were both greater here than those upstream of the cold wake (Figs. 6.21). The change of air-sea disequilibrium across the cold wake modulated the BL stability as well (Fig. 6.23). The BL varied from neutral in the upstream region, to strongly stable over the cold wake, to unstable downstream of the cold wake. The increased stability has the dynamic impact on BL air, which reduces surface wind speed and enhance the inflow (Fig. 6.24).

Overall, the simulated track and intensity in AO was almost on top of the JTWC best-track data (Fig. 6.1). The well-captured storm track was due to the nudging of winds. Compared to UA, the improvement of the intensity prediction in AO was because of the accurate prediction of air-sea coupled BL. This included not only the magnitude of cold wake, but also the size of the cold wake, the spatial distribution (especially radial distribution) and magnitude of surface wind, air-sea exchange fluxes, and the BL thermal and dynamic structure across the cold wake.

While the temporal and spatial limitations of observations prevented us from studying the evolution of the air property in the real air-sea coupled BL, the resemblance of the simulated storm in AO meant that we have an alternative way to examine this issue. The comprehensive tracer and trajectory analysis in numerical simulations showed that although the cold SST does decrease the heat energy carried by the airflow, the enhanced fluxes downstream of the cold wake compensate to a certain extent for the enthalpy reduction. Follow the $\theta_e$ of the BL air in the cold wake as the trajectories moved to the eyewall in AO, the increase of the $\theta_e$ can be as large as 13 K. In UA, the $\theta_e$ of the airflow starting from the rear-right quadrant (i.e., at the same storm relative location as the cold wake in AO) increased only by a maximum of 10 K, 3 K smaller than that in AO. The conclusion of the stabilizing effect of the cold wake found here is consistent to what we found in chapter 5 although these two storms had various structures in both storm and air-sea interface.
In summary, ITOP data and numerical simulations showed that there is a positive impact of the cold wake through a SBL formation. This is the first time the SBL in TCs is shown and analyzed in detail from observational data. By reducing the convection over the cold wake and the enhanced the convection in the eyewall through the enhanced surface inflow with more energetic air, SBL increases the amount of heat energy that is transferred into the eyewall and used to maintain/intensify storm intensity. This positive impact can mitigate the negative oceanic feedback to certain extend.
A recent study by Kepert (2012) pointed out that the results from numerical simulations might be biased because of the chosen planetary boundary layer (PBL) scheme. They suggested that the Yonsei University (YSU) scheme, the PBL scheme used in Chapters 5 and 6, should be used with caution regarding the diagnostic PBL height and resulting turbulent mixing. Thus, we examine whether the proposed mechanism by which the SBL affects TC structure (Chapters 5 and 6) is sensitive to the chosen PBL scheme. The YSU PBL scheme is an eddy diffusivity (K) profile parameterization (KPP) scheme that fits the K profile to the diagnostic PBL height and the surface conditions (Troen and Mahrt, 1986):

$$K = ku_\star z \left(1 - \frac{z}{h}\right)^p,$$

where $k$ is the von Karman constant, $u_\star$ is the friction velocity, and $p$ is the shape parameter. As Hong et al. (2006) shows, $h$ is defined as the altitude at which the bulk Richardson number is equal to the critical bulk Richardson number ($= 0$) over the neutral to unstable BL. The bulk Richardson number is defined as:
\[ R_i = \frac{gz}{\theta_{va}} \frac{\theta_v(h) - \theta_s}{|U(h)|^2}, \]

(7.2)

where \( U(h) \) is the horizontal wind speed at \( h \), \( \theta_{va} \) is the virtual potential temperature at the lowest model level, \( \theta_v(h) \) is the virtual potential temperature at \( h \) and \( \theta_s \) is the appropriate temperature at the surface (Hong et al. 2006). The PBL height is initially set to equal to \( h \), i.e. the altitude where \( \theta_v = \theta_s \). This represents the well-mixed layer. For a stable boundary layer, the critical Richardson number is increased to a value that is larger than zero. This value varies according to whether the grid is over water or land. In both cases, \( h \) is enhanced under stable conditions. Since KPP represents only the turbulent mixing below \( h \), the turbulent mixing above \( h \) is calculated using the Louis local K approach (Louis 1979). According to Kepert (2012), the KPP scheme creates stronger turbulent mixing compared to the local K approach. Thus, \( h \) determines not only the strength of the mixing but also the vertical extent of the stronger mixing from KPP.
Figure 7.1 shows the mean diagnostic BL height in YSU in both AO and UA. Compared to Fig. 5.6, the YSU BL height is overall about 200 to 300 m shallower than THBL. This is probably because THBL contains also a small transition layer right above the well-mixed layer. Despite the overall shallower BL height, YSU BL exhibits a similar pattern as THBL: BL decreases with radius and there is a shallower BL over the cold wake in AO, in particular where the BL is stable.

Nevertheless, if we use a different BL definition instead of the YSU BL, the depth of the KPP mixing layer will change. For example, with THBL, the deeper layer with KPP turbulent mixing would create more downward momentum flux and enhance the tangential component of the winds. The potential consequences are overall stronger surface winds and a decrease of the inflow velocity. However, this change does not affect the formation of the SBL and the suppressed convection because they are mostly related to the air-sea thermal disequilibrium. Since the thermodynamic impact of the SBL is the main reason for the cold wake to positively affect TC energetics, one can still expect to see this impact. Of course the change of the BL height in YSU might have other accumulated impacts on the symmetric and asymmetric structure of a TC, and this is beyond the scope of this study. Similarly, if we apply other PBL schemes, a reasonable increase or decrease of the turbulent mixing is not expected to obscure this SBL-formation mechanism, although a change of TC structure due to the accumulated impact of PBL parameterization could do so.
Chapter 8

Overall Conclusions

This study aimed to investigate how and to what extent air-sea interaction affects TC structures. Because the deep convection in the eyewall and rainbands are connected to the ocean surface through the HBL, it is also important to understand the physical mechanisms by which the HBL connects the upper-ocean and air-sea interface to the storm structure throughout the troposphere. Coupled atmosphere-(wave)-ocean models and the air-sea coupled observations from CBLAST (2003-2004) and ITOP (2010) are used. High-resolution numerical experiments for Hurricane Frances (2004), Typhoon Choiwan (2009), and Typhoon Fanapi (2010) are conducted. The convective structure, the BL structure, and the surface winds and surface fluxes for these three TCs were examined in detail in model field and observations.

The main findings are:

1. Ocean coupling changes both the surface winds, air-sea fluxes and vertical structures (in HBL) of winds, $\theta_v$, and $\theta_e$. This is due to the decrease of both sensible and latent heat fluxes caused by the storm-induced cooling. (Chapters 4, 5, and 6)

2. Wave coupling changes mainly the dynamic structure of the HBL. It increases (decreases) the surface wind that is in the high (low) wind regime. This is the effect of coupling the HBL to the surface waves, which reduces (increases) the drag coeffi-
cient in high (low) wind regime compared to the Charnock relationship used in the simulation without wave coupling. As a result, radial inflow wind increases outside of the RMW. (Chapter 4)

3. Both ocean and wave couplings induce asymmetries in the HBL height in terms of well-mixed layer and inflow layer. The well-mixed layer is shallower over the cold wake when ocean coupling is considered. On the other hand, the inflow layer is overall deeper with wave coupling. The highly asymmetric structure in both the well-mixed layer and inflow layer implies that the azimuthally-averaged profiles, as widely shown in literature, may not accurately represent the whole HBL structure. (Chapter 4)

4. The storm-induced cold wake changes the BL stability. This is mainly because the warmer air flows over the cooler sea surface over the cold wake. BL stability varies from neutral in the upstream region to strongly stable over the cold wake, and then to unstable downstream of the cold wake. In other words, there is a SBL formation over the cold wake. (Chapters 5, and 6)

5. As both sensible and latent heat fluxes are reduced over the cold wake because of low SSTs, this study found negative (downward) sensible heat fluxes over the cold wake because air is warmer than the ocean. Downstream of the cold wake, the air-sea temperature and moisture disequilibrium is enhanced because the colder air flows over warm SSTs. This is also the reason why the BL is unstable in the downstream region of the cold wake. (Chapter 6)

6. The SBL suppresses the convective activity over the cold wake and downstream of the cold wake and causes surface wind reduction. The reduced wind speed then results in the wind turning further inward (larger inflow angle) because of the gradient wind imbalance. Therefore, the air flowing over the cold wake tends to stay in the BL longer with stronger inward forcing.
7. The long residence time in the BL allows the near surface air to re-gain some energy via the enhanced enthalpy fluxes downstream of the cold wake on their way to the eyewall. This can mitigate the negative feedback and increase the storm efficiency of converting heat energy into kinetic energy. (Chapters 5, and 6)

8. Ocean coupling affects the TC structure through the reduced convection in rainbands and the enhanced convection in the eyewall due to the near surface inflow that results from the SBL over the cold wake. (Chapters 5 and 6)

While the oceanic negative feedback has been well established in the literature, this is the first time the formation of a SBL and its positive impact have been documented. This mechanism was examined by studying two typhoons (Choiwan and Fanapi). In the periods of interest, Choiwan was a moderate category-4 typhoon with a MWS of approximately 120 kts and Fanapi was a category-3 typhoon with a MWS of approximately 100 kts. Although both of them are in quasi-steady state, the structures for the storms and the underlying ocean in these two cases were quite different. Choiwan had a symmetric eyewall with a RMW of 25 km while Fanapi had an asymmetric eyewall with a RMW of 50 km. The cold wake in Choiwan (61,000 km$^2$) was much smaller than that in Fanapi (250,000 km$^2$). Here, the range of the cold wake is defined as the area with SST 0.5 °C less than the ambient SST. Interestingly, this physical mechanism is valid in both cases, which indicates that this may be a robust feature.

In summary, this study found that atmosphere-wave-ocean coupling affects the boundary layer structure and the physical properties of the near-surface airflow in TCs, which in turn changes the convective organization in TCs. These modulations of TC structures eventually directly affect the TC structure, energetics and intensity. This indicates that atmosphere-wave-ocean coupling affects the TC structure via complex physical processes, which include those that cause asymmetries in the boundary layer and the modulation of boundary layer flows. Hence it is difficult to parameterize the atmosphere-wave-ocean coupling processes in TCs without a fully coupled model.
Chapter 9

Future Work

Some research topics related to our current results could be explored in the future. First of all, the generality of the formation of a SBL in TCs needs to be examined. In addition, the physical mechanism by which the SBL positively affects TC energetics via modulating convective organization deserves further study. These topics require more coupled air-sea observations and coupled model simulations. With a massive data set (from both observations and numerical simulations), one can also quantitatively address the importance of the SBL on TC energetics systematically.

Another possible research topic for future work is the impact of surface waves on the TC structure. The majority of this study has focused on the influence of atmosphere-ocean coupling on TC structure. However, as shown in Chapter 4, wave coupling results in a change of both symmetric and asymmetric HBL structures. While the storm vortex structure is affected by its own HBL structure, the wave coupling is expected to change the TC structure as well. As the state of the art fully coupled UMCM-WMH has been developed/tested in the past two years, one can address this topic from a numerical perspective.
Bibliography


