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Creation and Application of the Systematically Merged Pacific Ocean Regional Temperature and Salinity (SPORTS) Climatology for Oceanic Heat Content Estimation

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UNIVERSITY OF MIAMI

A thesis submitted in partial fulfillment of
the requirements for the degree of
Masters of Science

CREATION AND APPLICATION OF THE SYSTEMATICALY MERGED PACIFIC
OCEAN REGIONAL TEMPERATURE AND SALINITY (SPORTS) CLIMATOLOGY
FOR OCEANIC HEAT CONTENT ESTIMATION

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The Systematically merged Pacific Ocean Regional Temperature and Salinity (SPORTS) climatology was created to estimate oceanic heat content (OHC) for the North Pacific (McCaskill et al., 2015). A technique similar to the creation of the Systematically Merged Atlantic Regional Temperature and Salinity climatology was used to blend temperature and salinity fields from the Generalized Digital Environment Model and World Ocean Atlas 2001 at a 0.25° resolution (Meyers et al., 2014). The weighting for the blending of these two climatologies was estimated by minimizing residual covariances across the basin and accounting for drift velocities associated with eddy variability using a series of 3-year sea surface height anomalies (SSHA) tracks to insure continuity between the periods of differing altimeters. In addition to producing daily estimates of the 20°C and 26°C isotherm depths (and their mean ratios), mixed layer depth, reduced gravities, and OHC, the SSHA product includes mapping errors given the differing repeat tracks from the altimeters and sensor uncertainties. These SPORTS products are available daily in near real-time on the Rosenstiel School of Marine and Atmospheric Science (RSMAS) Upper Ocean Dynamics research website and operationally at the National Oceanic and Atmospheric Administration (NOAA) National Environmental Satellite, Data, and Information Service (NEDSIS).
Using SPORTS with satellite-derived sea-surface temperature (SST) and SSHA fields from radar altimetry, daily OHC has been estimated from 2000 to 2011 using a 2.5-layer model approach. Argo profiling-floats, expendable probes from ships and aircraft, long-term TAO moorings, and drifters provide over 267,000 quality controlled in-situ thermal profiles to assess uncertainty in estimates from SPORTS. The in-situ profiles were used to evaluate the SPORTS OHC with a basin-wide regression analysis. TAO moorings and XBT transects were used to evaluate SPORTS OHC on a regional scale temporally and spatially.

A case study with the storms from the ONR-sponsored Impact of Typhoons on the Ocean in the Pacific (ITOP) 2010 experiments used SPORTS OHC to determine how OHC conditions before the storm contributed tropical cyclone (TC) intensification and TC induced ocean response. The SPORTS OHC before each TC showed that high OHC and horizontal ocean thermal gradients helped the ITOP storms intensify and maintain high TC intensity. Enthalpy fluxes were examined during the time while each TC intensified to its peak intensity to further investigate the TC intensification. The SPORTS OHC also helped explain the TC induced ocean SST cooling pattern. The momentum fluxes were calculated over the life cycle of the TCs to better understand the TC induced ocean response.

This thesis research was ultimately aimed at the public who must rely on the most advanced modeling systems to prepare for landfalling storms over the globe. An expected contribution of this research to society is a new daily real-time operational and 16-year archive SPORTS OHC that opens doors for avenues of research in the North Pacific Ocean basin.
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TABLE OF CONTENTS

| LIST OF FIGURES                  | …vi |
| LIST OF TABLES                  | …vii |
| LIST OF ABBREVIATIONS           | …viii |

CHAPTER

1 INTRODUCTION AND GOAL AND OBJECTIVES …1

1.1 Introduction …1
1.2 Goals and Objectives …1
  1.2.1 Creating OHC …2
  1.2.2 Assessing OHC …3
  1.2.3 Case Study …3
1.3 Summary …4

2 BACKGROUND REVIEW …6

2.1 OHC and Pre-storm Conditions …6
2.2 Enthalpy and Momentum Fluxes …8
2.3 Momentum and Current Response …10
2.4 Thermal Response …15
2.5 El Niño/Southern Oscillation …17

3 CREATION OF SPORTS OHC …20

3.1 Data Resources …20
  3.1.1 World Ocean Atlas 2001 …20
  3.1.2 GDEM v2.1 …20
  3.1.3 GDEM v2.1 over v3.0 …21
  3.1.4 Sea Surface Temperature …22
  3.1.5 Altimetry Data …23
  3.1.6 In-situ Profiles …24
3.2 Approach …26
  3.2.1 In-situ Profiles …27
  3.2.2 Objective Analysis …27
  3.2.3 Daily Climatology …29
  3.2.4 2.5-Layer Reduced Gravity Model …29
  3.2.5 Creating the SPORTS Climatology …31
  3.2.6 GDEM and WOA weighting maps …32

4 ASSESSMENT OF SPORTS OHC …35
<table>
<thead>
<tr>
<th>Section</th>
<th>Title</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>4.1</td>
<td>Basin-wide Analysis</td>
<td>35</td>
</tr>
<tr>
<td>4.2</td>
<td>TAO Mooring Comparison – Temporal Evaluation</td>
<td>38</td>
</tr>
<tr>
<td>4.3</td>
<td>XBT Transect Comparison – Spatial Evaluation</td>
<td>39</td>
</tr>
<tr>
<td>4.4</td>
<td>Linear Regression Based OHC Comparison</td>
<td>40</td>
</tr>
<tr>
<td>5</td>
<td>ITOP CASE STUDY – APPLICATION OF SPORTS OHC</td>
<td>42</td>
</tr>
<tr>
<td>5.1</td>
<td>Typhoon Fanapi</td>
<td>43</td>
</tr>
<tr>
<td>5.2</td>
<td>Typhoon Malakas</td>
<td>46</td>
</tr>
<tr>
<td>5.3</td>
<td>Typhoon Megi</td>
<td>49</td>
</tr>
<tr>
<td>5.4</td>
<td>Concluding Remarks</td>
<td>53</td>
</tr>
<tr>
<td>6</td>
<td>SUMMARY AND CONCLUDING REMARKS</td>
<td>55</td>
</tr>
</tbody>
</table>

REFERENCES 60
## LIST OF FIGURES

<table>
<thead>
<tr>
<th>CHAPTER</th>
<th>PAGE</th>
</tr>
</thead>
<tbody>
<tr>
<td>2</td>
<td>BACKGROUND REVIEW</td>
</tr>
<tr>
<td>2.1</td>
<td>TC satellite image and conceptual ocean response (Shay, 2010)</td>
</tr>
<tr>
<td>3</td>
<td>CREATION OF SPORTS OHC</td>
</tr>
<tr>
<td>3.1</td>
<td>October climatology temperature profiles compared to TAO</td>
</tr>
<tr>
<td>3.2</td>
<td>In-situ data positions (2000-2011)</td>
</tr>
<tr>
<td>3.3</td>
<td>Hovmöller of SSHA in western North Pacific</td>
</tr>
<tr>
<td>3.4</td>
<td>Weighing maps for GDEM and WOA</td>
</tr>
<tr>
<td>4</td>
<td>ASSESSMENT OF SPORTS OHC</td>
</tr>
<tr>
<td>4.1</td>
<td>SPORTS scatter plots and histograms</td>
</tr>
<tr>
<td>4.2</td>
<td>SPORTS Typhoon season normalized RMSD values</td>
</tr>
<tr>
<td>4.3</td>
<td>TAO vs SPORTS comparison</td>
</tr>
<tr>
<td>4.4</td>
<td>XBT vs SPORTS comparison</td>
</tr>
<tr>
<td>5</td>
<td>ITOP CASE STUDY – SPORTS APPLICATION</td>
</tr>
<tr>
<td>5.1</td>
<td>ITOP storm tracks</td>
</tr>
<tr>
<td>5.2</td>
<td>Fanapi pre-storm OHC and SSHA</td>
</tr>
<tr>
<td>5.3</td>
<td>Fanapi before, after, and difference SST</td>
</tr>
<tr>
<td>5.4</td>
<td>Fanapi enthalpy fluxes</td>
</tr>
<tr>
<td>5.5</td>
<td>Malakas pre-storm OHC and SSHA</td>
</tr>
<tr>
<td>5.6</td>
<td>Malakas before, after and difference SST</td>
</tr>
<tr>
<td>5.7</td>
<td>Malakas enthalpy fluxes</td>
</tr>
<tr>
<td>5.8</td>
<td>Megi pre-storm OHC and SSHA</td>
</tr>
<tr>
<td>5.9</td>
<td>Megi before, after, and difference SST</td>
</tr>
<tr>
<td>5.10</td>
<td>Megi enthalpy fluxes</td>
</tr>
</tbody>
</table>
## LIST OF TABLES

<table>
<thead>
<tr>
<th>CHAPTER</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>3  CREATION OF SPORTS OHC</td>
<td></td>
</tr>
<tr>
<td>3.1 Altimetry for SSHA used each year (2000-2011)</td>
<td>23</td>
</tr>
<tr>
<td>3.2 Objective analysis parameters</td>
<td>27</td>
</tr>
<tr>
<td>4  ASSESSMENT OF SPORTS OHC</td>
<td></td>
</tr>
<tr>
<td>4.1 Percent of SPORTS values near observed values</td>
<td>37</td>
</tr>
<tr>
<td>5  ITOP CASE STUDY – SPORTS APPLICATION</td>
<td></td>
</tr>
<tr>
<td>5.1 Estimated Momentum Response Variables</td>
<td>54</td>
</tr>
<tr>
<td>6  SUMMARY AND CONCLUDING REMARKS</td>
<td></td>
</tr>
<tr>
<td>6.1 Altimetry for SSHA used each year (2000-2015)</td>
<td>57</td>
</tr>
</tbody>
</table>
LIST OF ABBREVIATIONS

AMSR-E – Advanced Microwave Scanning Radiometer - EOS
AOML – Atlantic Oceanographic and Meteorological Laboratory
AVHRR – Advanced Very High Resolution Radiometer
CBLAST – Coupled Boundary Layer Air-Sea Transfer
D20 – Depth of the 20°C isotherm
D26 – Depth of the 26°C isotherm
ENSO – El Niño/Southern Oscillation
Envisat – ENVIronmental SATellite
EOS – Earth Observing System
EPIC – Eastern Pacific Investigation of Climate
ERS – European Remote Sensing
GDEM – Generalized Digital Environmental Model
GFO – Geosat Follow-On-Missions
GOES – Geostationary Operational Environmental Satellite
GPS – Global Positioning System
ITOP – Impacts of Typhoons on the Ocean in the Pacific
JTWC – Joint Typhoon Warning Center
MLD – Mixed Layer Depth
NASA – National Aeronautics and Space Administration
NOAA – National Oceanic and Atmospheric Administration
NESDIS – National Environmental Satellite, Data, and Information Service
NRMSD – Normalized Root Mean Squared Deviation
OA – Objective Analysis
OI – Optimally Interpolated
OHC – Ocean Heat Content
OML – Ocean Mixed Layer
ONI – Oceanic Niño Index
ONR - Office of Naval Research
PMEL – Pacific Marine Environmental Laboratory
POES – Polar-orbiting Operational Environmental Satellite
RMSD – Root Mean Squared Deviation
RSMAS – Rosenstiel School of Marine and Atmospheric Science
SARAL – Satellite with ARgos and ALtiKa
SCS – South China Sea
SHIPS – Statistical Hurricane Intensity Prediction Scheme
SMARTS – Systematically Merged Atlantic Regional Temperature and Salinity
SPORTS – Systematically merged Pacific Ocean Regional Temperature and Salinity
SSHA – Sea Surface Height Anomaly
SST – Sea Surface Temperature
ST – Super Typhoon
TAO – Tropical Atmosphere Ocean
TC - Tropical Cyclone
TMI – TRMM Microwave Imager
TRMM – Tropical Rainfall Measuring Mission
WOA – World Ocean Atlas
WCR – Warm Core Ring
XBT – Expendable Bathythermograph
Chapter 1: Introduction and Goal and Objectives

1.1 Introduction

Tropical cyclone (TC) research has been important for decades in the effort to more effectively forecast these storms to save life and property. An accurate intensity forecast provides critical important information to local, state and federal government entities needed to best protect and prepare the general population (Marks and Shay, 1998). Energy release from latent and sensible heat fluxes at the air-sea interface contribute to the intensification of the TC, making the ocean an integral part of the intensification process (Emanuel, 1986). Over two decades of research have proved that ocean heat content (OHC), relative to the 26°C isotherm, is the best indicator of the ocean’s thermal energy available for TC intensification and maintenance. In particular, studies show areas of high OHC support rapid intensification and high intensity TCs (Shay et al. 2000; Hong et al. 2000; Mainelli et al. 2008; Jaimes and Shay 2009). This phenomenon can be observed regularly in the North Pacific where intense TC activity occurs over regions of oceanic warm features with deep depths of the mixed layer and the 26°C isotherm (Lin et al. 2005). To help identify these oceanic warm features across the North Pacific, a realistic OHC product must be created.

1.2 Goal and Objectives

The overarching goal of this thesis was to create an accurate OHC product in the North Pacific Ocean following the methods used by Meyers et al., (2014). To achieve this goal, two objectives focused on:
(i) Creating an OHC product for the North Pacific (Section 1.2.1), and

(ii) Assessing the accuracy of the satellite-derived of the OHC product (Section 1.2.2).

In addressing Objective (ii), a case study examined an application of the OHC product during the Impacts of Typhoons on the Ocean in the Pacific (ITOP) 2010 experiment (Section 1.2.3).

1.2.1 Creating OHC

A Systematically merged Pacific Ocean Regional Temperature and Salinity (SPORTS) climatology was created to estimate ocean heat content (OHC) for tropical cyclone (TC) intensity forecasting and other applications (McCaskill et al., 2016). A technique similar to the creation of the Systematically Merged Atlantic Regional Temperature and Salinity (SMARTS) climatology was used to blend temperature and salinity fields from the Generalized Digital Environment Model and World Ocean Atlas 2001 at a 0.25° resolution (Meyers et al., 2014). The weights for the blending of these two climatologies were estimated by minimizing residual covariances across the basin between the climatologies and 267,000 quality controlled in-situ profiles. Drift velocities associated with eddy variability were accounted for using a series of 3-year sea surface height anomalies (SSHA) to ensure continuity between the periods of different altimeters. In addition to producing daily estimates of the 20°C and 26°C isotherm depths, mixed layer depth, and OHC, the model produces mapping errors from the optimal interpolation of the SSHA due to gaps in altimeter track coverage and sensor uncertainties. Using SPORTS with satellite-derived sea-surface temperature (SST) and SSHA fields from
radar altimetry, daily OHC was estimated from 2000 to 2011 using a 2.5-layer model approach.

1.2.2 Assessing OHC

Argo profiling-floats, expendable probes (XBTs) from ships and aircraft, long-term Tropical Atmosphere Ocean (TAO) moorings, and drifters provide more than 267,000 quality controlled in-situ thermal profiles to assess uncertainty in estimates from SPORTS. A regression analysis proved that the SPORTS OHC accurately estimates OHC. The normalized root mean squared difference analysis identified regions of uncertainty in the SPORTS OHC, which correlated with regions of high eddy activity. A TAO (XBT) analysis showed how well SPORTS OHC was estimated temporally (spatially). A comparison was made to the Pun et al. 2014 western North Pacific OHC product.

1.2.3 Case Study

The objective of this case study explores TC air-sea interactions with the upper ocean, using SPORTS OHC as a tool. The study focused on the three storms from the Impact of Typhoons on the Ocean in the Pacific (ITOP) experiment occurred in 2010; Typhoon Fanapi, which occurred over moderate OHC, Typhoon Malakas over low OHC, and Typhoon Megi over high OHC (D’Asaro et al., 2011; Mrvaljevic et al., 2013). Over 1,000 atmospheric and oceanic profiles were acquired during the experiment from aircraft deployed dropsondes, floats and drifters, vessel deployed gliders and drifters, and an ocean mooring array. Pre-storm SPORTS OHC and SSHA helped determined how the upper ocean contributed to the TC’s intensification. Enthalpy fluxes calculated from aircraft deployed instruments further demonstrated the ocean’s contribution to TC
intensification. Momentum fluxes showed how the upper ocean response to TC forced winds.

1.3 Summary

The SPORTS OHC for the North Pacific is created and assessed using over 267,000 in-situ profiles, making it unique amongst other OHC products. The 2.5-layer reduced gravity model produced accurate estimates using the SPORTS climatology, and satellite SSHA and SST. The SPORTS OHC was evaluated temporally with TAO data, spatially with XBT data, and with a basin-wide regression utilizing the over 267,000 in-situ profiles.

The ITOP storms provided an opportunity to use SPORTS OHC as tool in a case study. The study investigated how pre-storm conditions impacted TC intensity. The momentum and enthalpy fluxes gave insight to the TC air-sea interaction. The SPORTS OHC helped identify how OHC contributed to TC intensification and TC induced ocean response.

This thesis research was ultimately aimed at the public who must rely on the most advanced modeling systems to prepare for landfalling storms over the globe (Marks and Shay, 1998). An expected contribution of this research to society is a new daily real-time and 16 year archive SPORTS OHC that opens doors for avenues of research in the North Pacific Ocean basin. The SPORTS OHC provides an accurate daily OHC product at a quarter degree resolution, which is used real-time operationally at the National Oceanic and Atmospheric Administration (NOAA) National Environmental Satellite, Data, and Information Service (NESDIS). The SPORTS OHC also provides a resource to help evaluate the seasonal changes of OHC impact on TC activity. The 16-year SPORT OHC
dataset covers the different El Niño/Southern Oscillation (ENSO) phases, allowing for a study how ENSO related ocean variability affects seasonal TC activity.
Chapter 2: Background Review

2.1 OHC and Pre-storm Conditions

TC’s are incredible forces of nature that have a profound impact on humanity. The ability to accurately predict these storms has always been important for the safety of lives and property. The ocean plays a significant role in forecasting TC intensity, providing a source of energy to power the storm. Leipper and Volgenau (1972) quantify ocean heat content as the thermal energy available in the upper ocean

$$OHC = \int_{D_{26}}^{56} c_p \rho (T - 26^\circ C) dz$$  \hspace{1cm} \ldots(2.1)$$

where $c_p$ is the specific heat of seawater taken as $4200 \text{ J (kg°C)}^{-1}$, $\rho$ is density of seawater, and $T$ is the temperature at depth. The $26^\circ$ isotherm was used based on results found by Palmen (1948), finding that a minimum sea surface temperature (SST) of $26^\circ$C was important for tropical cyclone formation. High values of OHC provide a source of thermal energy available for a tropical storm through heat fluxes. Several case studies have shown the importance of high OHC in tropical cyclone intensification. Shay et al. (2000) studied Hurricane Opal as it passed over a warm core ring (WCR) in the Gulf of Mexico in October of 1995. Hurricane Opal traversed over the WCR with OHC values over $100 \text{ kJ cm}^{-2}$, allowing the storm to rapidly intensify from a category one storm to a category four storm over an approximate 14-hour period. A coupled ocean-atmosphere model showed that Hurricane Opal would not have reached such intensity without the presence of the WCR (Hong et al., 2000). Similarly, Hurricane Katrina and Rita in 2005
experienced rapid intensification over areas of high OHC associated with the Loop Current and WCR complex (Jaimes and Shay, 2009).

Mainelli (2000) proved the value of OHC in hurricane intensity forecast by using OHC as a forecast variable in the Statistical Hurricane Intensity Prediction Scheme (SHIPS; DeMaria et al., 2005). The intensity forecast improved by 5-6 % for category five storms and by 22% in the case of Hurricane Ivan (2004) (Mainelli et al., 2008).

In the Northern Pacific Ocean, a SPORTS daily OHC product was created following Meyers et al. 2014 SMARTS OHC product to aid in TC intensity forecasts. The SPORTS OHC uses a 2.5 layer reduced gravity model and a daily blended SPORTS climatology in concert with satellite daily SSHA and SST data to produce estimates of OHC. Approximately 267,000 in-situ data points were used in the creation and evaluation of SPORTS OHC. Further detail on the creation and verification of SPORTS can be found in Chapter 3 and 4. Pun et al. 2014, (further referenced as Pun14) also created an OHC product to help in the intensity forecast of TCs in the Western Pacific Ocean basin. Comparisons between the SPORTS and Pun14 OHC are made in Chapter 4.4.

The strength of the thermocline also has an impact on TC intensification (Shay and Brewster, 2010). A strong thermocline, as observed in the eastern tropical Pacific Ocean during the Eastern Pacific Investigation of Climate (EPIC), will inhibit mixing of cooler waters from the thermocline, allowing for lower OHC values necessary for intensification of TCs over that oceanic regime. Horizontal temperature changes (e.g., thermal gradients) also have an impact on TC intensification. Areas of sharp thermal gradients have the strongest air-sea fluxes during a storm passage (Jaimes and Shay,
2009; 2015) which help to increase the storm’s intensity. These strong thermal gradients suppresses the vertical mixing and weakens the ocean response, therefore reducing the negative feedback to the atmosphere (Shay and Uhlhorn, 2008).

Clearly, OHC, the strength of the thermocline across the base of the oceanic mixed layer (OML), and horizontal ocean thermal gradients are all conditions existing prior to TC passage that impact the level of the oceanic response, and hence the storm’s intensity. In this context, the upper ocean responds to a TC in three ways: (i) upwelling of cooler thermocline water through Ekman pumping induced by wind-driven currents from the storm’s cyclonically rotating wind field, (ii) turbulent entrainment of cooler thermocline waters through wind-driven current shear across the base of the OML, and (iii) latent and sensible heat loss through the air-sea interface (Shay et al., 1992; Jacob et al., 2000; Jaimes and Shay, 2009). The first two responses depend on momentum fluxes while the third one depends enthalpy fluxes.

2.2 Enthalpy and Momentum Fluxes

The maximum intensity of a TC depends largely on a balance between surface moist enthalpy fluxes, energy lost to frictional dissipation, and heat lost to the surroundings in the atmosphere (Emanuel, 1986). Therefore, momentum and enthalpy fluxes at the air-sea interface are crucial to TC intensity (Emanuel, 1986; Malkus and Riehl, 1960). The momentum ($\tau$), sensible ($Q_s$), and latent heat ($Q_l$) surface fluxes can be estimated by the bulk formulas:
\[ \tau = \rho_a c_p |U_{10}|^2, \]  
\[ Q_s = \rho_a c_p c_h |U_{10}|(SST - T_a), \]  
\[ Q_l = \rho_a L_v c_q |U_{10}|(q_s - q_a), \]

where \( \rho_a \) is the density of the atmosphere, \( c_p = 1004 \text{ J (kgK)}^{-1} \) is specific heat of air at constant pressure, \( L_v = 2.5 \times 10^6 \text{ J kg}^{-1} \) is the latent heat of vaporization, \( C_D, C_h, \) and \( C_q \) are the momentum, sensible, and latent heat exchange coefficients, \(|U_{10}|\) is the 10-m wind speed, \( T_a \) is the 10-m air temperature, \( q_a \) is the 10-m air specific humidity, and \( q_s = 98\% \) saturation (Buck, 1981) is the saturated specific at the SST (Jaimes et al., 2015). Note that \( C_h = C_q = C_K \). The momentum flux allows for estimations of surface friction velocity \( (u_*) \) to be made, which dictates the amount of work done on the ocean mixed layer through instantaneous turbulent mixing by wind erosion (Kraus and Turner, 1967). This turbulent mixing can act to decrease SSTs and will be described later. The combination of the latent and sensible heat fluxes make the enthalpy flux \( (Q_h = Q_s + Q_l) \). The enthalpy fluxes provide positive energy to the TC when SSTs are sufficiently warm (26°C or more) (Palmen, 1948). As the SST decreases to 26°C or below, the enthalpy fluxes decrease and begin to have a negative feedback affect on TC intensity (Chang and Anthes, 1978).

TC intensity is sensitive to the ratio of the enthalpy and surface drag coefficients \( (C_K/C_D) \) (Emanuel, 1995; Ooyama, 1969). This ratio appears in the calculations for maximum wind potential in the eye wall region and central pressure calculations.
(Emanuel, 1995). The ratio also provides a method for estimating \( C_K \), which is used in the enthalpy flux calculations, given a constant and \( C_D \). Theoretically, the \( C_K/C_D \) ratio typically ranges from 1.2 to 1.5 in the high wind regions of TCs (Emanuel, 1995). However, observational research from the ONR sponsored Coupled Boundary Layer Air-Sea Transfer (CBLAST) experiment has found the ratio to have mean values from 0.63 to 0.7 with values as low 0.5 (Zhang et al., 2008). The ratio variations are controlled by \( C_D \) which increases with wind speed until leveling off when winds reach 28 to 33 m s\(^{-1}\) (Powell et al., 2003; Donelan et al., 2004; Shay and Jacob, 2006; French et al., 2007; Jarosz et al., 2007; Sanford et al., 2007); beyond this threshold value, \( C_K/C_D \) is independent of wind speed (Black et al., 2007; Drennan et al., 2007; Zhang et al., 2008). The estimated value of \( C_K \) can then be used in Equations (2.3) and (2.4) to determine the heat lost from the ocean to a TC. Atmospheric and oceanic data collected during the ITOP storms can be input into these equations to better understand the energy exchange at the air-sea interface, ultimately impacting the TC’s intensity.

The ocean horizontal and vertical current responses are strongly dependent on the surface drag coefficient, \( C_D \) (Price, 1983). Accurate estimates of \( C_D \) allow for calculations of the momentum flux (\( \tau \)) to be made. The momentum flux then becomes an important variable in dictating the strength of the current response and upwelling, as will be seen in Equations 2.5-2.8.

2.3 Momentum and Current Response

As TC winds rotate counterclockwise, it creates a wind stress on the ocean surface below, imparting momentum into the upper ocean. The added momentum causes vertical
motion and elicits two different types of current responses, barotropic and baroclinic. The counterclockwise rotating wind stress on the ocean surface results in a divergent current flow away from the center of the hurricane. This divergent flow under the eye of the TC causes a surface depression to form and the thermocline upwells to fill the space left by the diverging waters (Figure 2.1; Shay, 2010). Also shown on the picture are surface heat fluxes and mixing induced by vertical current shear.

![Figure 2.1](Shay, 2010). (a) A cross section is taken through a TC. (b) A conceptual response of the ocean to TC forced winds. Shear-induced mixing, thermocline upwelling, ocean surface heat fluxes, ocean mixed layer deepening, and mixed layer divergence at the center of the storm is shown.

Theoretically, the upwelling in a stratified ocean at rest under a strong, stationary axisymmetric hurricane will be confined within 2 radius of maximum winds ($R_{\text{max}}$) from the center of the storm and will be surrounded by a broad area of weak downwelling from the displaced surface waters (O’Brien and Reid, 1967). Air-sea parameters can be used to estimate the upwelling ($w$), displacement of isopycnals ($\eta$), and the ocean mixed layer (OML) horizontal velocity response ($U_{\text{OML}}$) from TC forcing (Price, 1983)

\[
w = \frac{\tau_{\text{max}}}{\rho_o U_h}, \quad \ldots(2.5)\]

\[
\eta = \frac{\tau_{\text{max}}}{\rho_o f U_h}, \quad \ldots(2.6)\]
\[ U_{OML} = \frac{\tau_{\text{max}} R_{\text{max}}}{\rho_o h U_h}, \quad \ldots (2.7) \]

where \( \tau_{\text{max}} \) uses the maximum \( U_{10} \) winds, \( \rho_o = 1025 \text{ kg m}^{-3} \) is seawater density, \( U_h \) is the storm translation speed, \( f \) is the Coriolis parameter, and \( h \) is the thickness of the OML. However, the ocean is usually not at rest but has pre-existing geostrophic flows and eddy fields which impact the upwelling velocities (Stern 1965; Jaimes and Shay 2009, 2015; Jaimes et al., 2011) and the dynamical response (Shay et al., 1992; 1998). When the oceanic geostrophic flow interacts with the surface wind stress, upwelling regions can be found where the two flows align and downwelling regions occur where the two flows go against each other. A revised formulation for the vertical velocity accounts for influence of the background flow (Jaimes and Shay, 2015) to produce more realistic upwelling and downwelling fields under a TC.

\[ w_z = -R_{\text{g}} \delta (U_h + U_{OML}) + w_E \quad \ldots (2.8) \]

where \( R_{\text{g}} = \zeta_g / f \) is the eddy Rossby number, \( \zeta_g \) is the geostrophic vorticity, \( \delta = h/R_{\text{max}} \) is the aspect ratio, and \( w_E \) is the undisturbed Ekman pumping. Equation 2.8 remains valid under the assumptions of \( R_{\text{g}} \ll 1, R_{\text{g}} \ll 1 \), where \( R_{\text{g}} = U_{OML} / f R_{\text{max}} \) is the frictional Rossby number. The presence of the strong eddy field in the tropical Western Pacific will make the use of Eq 2.8 necessary to assess TC responses in that region. Equations 2.5-2.8 will be used to determine upper ocean current response fields during the ITOP storms.

Along with the vertical response, TC winds force a horizontal current response, which can be have a weak barotropic and strong baroclinic response. As mentioned previously, the divergent surface flow under the center of the TC causes a depression at
the sea surface. This depression ranges from 10 to 20 cm deep (Shay et al., 1990; Shay and Chang, 1997) and causes the barotropic response. The cold, surface depression acts as a low pressure and induces counterclockwise, geostrophic flow around it. The low sea surface height is elongated along the track of the hurricane and persists for several days after the storm passing.

The baroclinic response is an evolving 3-D response manifested by spreading isotherms and an internal wave wake with strong near-inertial motions. The response is primarily governed by the directional change of the surface wind stress over time and the speed of the TC movement, which displaces the interface between warm and cool layers of the upper ocean (Chang and Anthes, 1978). At a fixed oceanic location, the wind stress field changes, with clockwise rotation on the right side of the storm and counterclockwise rotation on the left side (Price, 1981). This phenomenon, known as resonance excitation, produces enhanced near-inertial ocean currents on the right side of the storm between 1 to 3 $R_{\text{max}}$. This enhanced near-inertial motion also excites larger vertical shears to induce stronger and more sustained vertical mixing events compared to those usually not observed on the left side of the storm. This is known as the right-hand bias. The interactions of the initial rotating currents with the rotation of the wind stress vectors cause the right side bias (Chang and Anthes, 1978; Price, 1983). The clockwise (counterclockwise) turning of the wind stress vectors on the right (left) side of the storm enhances (weakens) the ocean currents on the right (left) side of the track. The currents oscillate in a cycle that results in convergence and divergence of the mixed layer fluid over a time scale of the local near-inertial period. An example of this can be seen in the Hurricane Gilbert data (Shay et al., 1992; 1998) from current profilers deployed during
and after the passage of hurricane Gilbert in the Gulf. The convergence and divergence of the currents force fluid to upwell and downwell at the base of the OML, creating a horizontal pressure gradient on interface between the mixed layer and thermocline. The presence of this horizontal pressure gradient causes the currents in the OML to oscillate at a near-inertial frequency. The horizontal pressure gradient can also create a baroclinic current response in the thermocline with weaker amplitudes of oscillation (Shay and Elsberry, 1987). These baroclinic oscillations propagate away from the storm track as a response to geostrophic adjustment and can be found to persist for up to 4 weeks after the storm has passed (Shay and Elsberry, 1987).

The Froude (Fr) number, as described in Equation 2.9 where $U_h$ represents the translation speed of a TC and $c_1$ represents first baroclinic mode phase speed, determines which response will dominate, barotropic or baroclinic (Shay, 2001). The first baroclinic mode phase speed in a two-layer model where $\rho_1$ and $\rho_2$ represent the densities of the top and bottom layer density respectively and $\rho_2 > \rho_1$ (Equation 2.10). The depths of the top and bottom layer are represented by $h_1$ and $h_2$ respectively. If the Froude ratio is greater than one (the translation speed is faster then the baroclinic mode speed), then the baroclinic response will dominate. If the ratio is less than one, then the barotropic response will dominate.

$$Fr = \frac{U_h}{c_1} \quad \text{(2.9)}$$

$$c_1^2 = \frac{g(\rho_2 - \rho_1)h_1h_2}{\rho_2(h_1 + h_2)} \quad \text{(2.10)}$$
The Froude number also affects the level of the upwelling response (Geisler, 1970). Stronger upwelling is associated with Froude numbers less than one (suggestive of a stationary storm), while weaker upwelling occurs with Froude numbers greater than one when the baroclinic near-inertial response dominates in a spreading 3-D wave wake (Price, 1983).

2.4 Thermal Response

The thermal response of the ocean is primarily dependent on entrainment at the base of the mixed layer, but is also influenced by surface fluxes and advection by the currents (Jacob et al., 2000). The equation for the changes in OML temperature is found to be:

\[
\frac{\partial T}{\partial t} = -u \frac{\partial T}{\partial x} - v \frac{\partial T}{\partial y} - \frac{Q_o}{\rho_o c_p h} - \gamma \frac{w \Delta T}{h} \tag{2.11}
\]

The advective and surface flux terms account for up to 15% of the change in OML temperature, whereas the entrainment flux account for 70 to 80% of the OML temperature change (Jacob et al., 2000). Entrainment heat flux is caused by the shear at the base of the OML from the baroclinic near-inertial current response, wind driven current shear, or convective overturning from surface fluxes (Price, 1981). Of these three, the shear at the base of the OML due to baroclinic near-inertial current response is the dominant factor (Price, 1981; Sanford et al., 1987; Shay et al., 1992). As near-inertial motions propagate vertically through the base of the OML via the near-inertial pumping mechanism, current shears between the mixed layer and thermocline, and the thermocline buoyancy control the vertical mixing occurring at the OML base. The gradient
Richardson number determines how the ocean will respond with the given buoyancy and current shear field. The gradient Richardson number is defined as:

\[ Ri = -\frac{N^2}{\left(\frac{\partial U}{\partial z}\right)^2} \]  

...(2.12)

where \( N^2 \) is the buoyancy frequency and \( \frac{\partial U}{\partial z} \) is the vertical current shear. When the gradient Richardson number is less than 0.25, mixing at the OML base occurs. The mixing acts to increase the MLD, as the cooler thermocline waters mixes with the warmer upper ocean waters. The OML depth will exhibit a maximum on the right side of the track due the resonant near-inertial current excitation at 1-3 \( R_{\text{max}} \) from the storm track as noted above (Price, 1981; Shay et al., 1992). This creates enhanced mixing on the right side, leading to deeper MLDs. The converging (diverging) regions of the near-inertial currents can also act to increase (decrease) the OML depth via downwelling (upwelling) (Jacob et al., 2000). The divergent transport of the upper ocean layer causes cooler thermocline waters to be upwelled into the mixed layer (Figure 2.1). Therefore, the OML depth will be decreased directly beneath the track. Again, the Richardson number controls the mixing of this process. Entrainment mixing at the base of mixed layer always acts to reduce the SST (Price, 1981), bringing cooler temperatures to the surface.

In addition, although considerably weaker, surface heat fluxes can induce convective overturning in the upper ocean (Jacob et al., 2000). Evaporation of ocean water into the atmosphere causes the surface of the ocean to be saltier and cooler through latent and sensible heat release. This temperature response causes the surface water to
become denser than the water below and therefore sink, mixing with the warmer waters below. As stated previously, this process only accounts for a small amount of the decrease in the SST and OML temperature observed during a TC. Thus, the resultant SST decrease from entrainment mixing and surface fluxes can range from 1°C to 6°C with a rightward bias described above (Price, 1981; Jacob et al., 2000). These SST changes can remain present about a week to ten days following TC passage depending on the intensity of the TC and the background stratification.

Precipitation in a TC acts to freshen the surface waters, stabilizing the water column and can decrease the SST by as much as 0.5°C (Jacob and Koblinsky, 2007). Salinity changes of 0.2 to 0.4 psu are typically seen as a result of TC precipitation (Pudov and Petrichenko, 2000). Storm translation speed and rainfall rates impact the amount of freshening to the surface waters due to precipitation (Shay and Uhlhorn, 2008). Faster storm translation speeds result in less freshening of the salinity while heavier rainfall rates result in more freshening of the salinity.

OHC can be used to help predict the amount of SST cooling resulting from the passage of a TC. Areas of high OHC act to dampen the amount of SST cooling due to upwelling and loss of heat to the storm, resulting in lower than normal SST cooling. Ocean thermal profile measurements and satellite SST taken before, during, and after a TC can be used to assess the thermal response to a TC, as will be done for the ITOP storms.

2.5 El Niño/Southern Oscillation

The El Niño/Southern Oscillation (ENSO) consists of a linked variation between tropical trade winds and upper ocean thermal patterns in the equatorial Pacific (Bjerkens
El Niño (La Niña) conditions describe an ENSO event where the tropical eastern Pacific surface waters are anomalously warm (cold) (Bjerkens, 1969). In neutral conditions, the atmospheric Walker circulation sets up with deep convection and rising motion over the tropical western Pacific Ocean, sinking motion in the tropical eastern Pacific, and easterly trade winds at the surface between the rising and sinking air (Walker and Bliss, 1932). The winds during the neutral period act to push the warm surface waters to the west, resulting in deep warm pool in the tropical western Pacific with deep thermocline depths (Wyrtki, 1975). The movement of the surface waters westward leads to upwelling and cooler surface waters in the eastern tropical Pacific. One of the theories behind the mechanisms of the ENSO cycle is the delayed oscillator theory (Suarez and Schopf, 1989; Battisti, 1988; Battisti and Hirst, 1989; Wakata and Sarachik, 1991; Mantua and Battisti; 1994). During El Niño conditions, the westerly wind stress anomalies excite eastward propagating downwelling Kelvin waves and upwelling westward propagating Rossby waves. These waves deepen the thermocline in the central and eastern Pacific and lift the thermocline in the western Pacific. The deeper thermocline in the central and eastern Pacific allows for higher than normal OHC values across the equatorial Pacific during the El Niño. The upwelling Rossby waves travel westward and reflect off the topography in the far western Pacific, becoming eastward propagating upwelling Kelvin waves. When the upwelling Kelvin waves reach the central and eastern Pacific, they interact with the downwelling Kelvin waves, weakening then and eventually overtaking the downwelling. This process takes a little over a year and becomes the onset of La Niña. The upwelling in the eastern Pacific causes the thermocline in the eastern and central Pacific resulting in a cold SST anomaly. The cold
SST anomaly causes the trade winds to strengthen over the central Pacific, exciting westward propagating downwelling Rossby wave that create a greater pooling of warm water in the west and eastward propagating upwelling Kelvin waves which further amplify the cold SST anomaly in the east. The thermocline deepens in the western Pacific, increasing OHC, and sharply rises to east outside of the deep warm pool. While the western Pacific retains high OHC values during a La Niña, the central and eastern Pacific have lower than normal OHC values due to the shallower thermocline depths. The ENSO process restarts when the reflected Rossby waves during the La Niña become eastward propagating downwelling Kelvin waves that counteract the upwelling Kelvin waves forced from the central Pacific winds.
Chapter 3. Creation of SPORTS OHC

3.1 Data Resources

The SPORTS climatology was created as a blend of two climatologies to maximize the regional strengths of each. Approximately 267,000 in-situ data points were used to evaluate which climatology produced the most accurate depth of 20°C (D_{20}) and 26°C (D_{26}) isotherms, MLD and OHC calculations for the 2.5-layer reduced gravity model. In addition to the blended climatology, the 2.5-layer model also required daily measurements of satellite SST and SSHA.

3.1.1 World Ocean Atlas 2001

WOA 2001 is a monthly climatology product of NOAA’s National Oceanographic Data Center. Temperature and salinity profiles were derived from the World Ocean Database, comprised of over 7 million profiles from Conductivity, Temperature, and Depth data, mechanical bathythermographs, expendable bathythermographs, and moored buoy data (Conkright, 2002). An objective analysis was performed on the in-situ data and the 1° WOA product to produce a climatology at a 0.25° resolution. A median smoother was used to smooth the quarter-degree climatology, using five data points on either side of the datum point. These vertical profiles consisted of up to 24 depths, reaching depths of 5500 m where possible, with higher-resolution in the upper ocean.

3.1.2 GDEM v2.1

The Generalized Digital Environmental Model (GDEM) v2.1 is a product of the Naval Oceanographic Office. The monthly climatology was produced using over 5.5 million temperature and salinity profiles from the Master Oceanographic Observation
Dataset in a set of four-dimensional steady-state digital models (Teague et al., 1990). The model produced profiles on a 0.5° grid, providing a three dimensional representation of the ocean. Vertical profiles consisted of up to 34 depths, measuring to as deep as 9000 m, with higher vertical resolution in the upper ocean. The GDEM climatology has global coverage from 40°S to 90°N. An objective analysis (OA) scheme by Mariano and Brown (1992) was used to objectively map the resolution of GDEM v2.1 from 0.5° to 0.25° for the use of SPORTS. The OA scheme is briefly explained in section 3.3.2.

3.1.3 GDEM v2.1 over v3.0

The SPORTS climatology differs from SMARTS by using GDEM v2.1 instead of GDEM v3.0. Generally, the North Pacific Ocean is more stratified than the North Atlantic Ocean. Research has shown that the GDEM v2.1 climatology is more suited for the North Pacific than GDEM v3.0 (Shay and Brewster, 2010). That study focused only in the tropical eastern North Pacific; however, the western North Pacific is a prolific TC region with an active oceanic eddy field. Comparisons of the October monthly temperature profiles from GDEM v2.1, GDEM v3.0, and WOA 2001 to the monthly averaged October temperature profile from TAO mooring (8°N, 137°E) confirm that GDEM v2.1 fit the observed profile better than GDEM v3.0 in the western North Pacific (Figure 3.1). Daily data from the TAO mooring was averaged over 11 years (2001-2011) during the month of October to determine the average monthly profile. Further analysis was performed over the entire North Pacific Ocean basin to determine which version of GDEM was more consistent with recent data. In this context, over 267,000 in-situ profiles were used to determine that the GDEM v2.1 climatology estimates of the vertical temperature profile for the North Pacific Ocean are better than GDEM v3.0 for over 95%
of the domain. Therefore, GDEM v2.1 was used for the creation of SPORTS OHC instead of GDEM v3.0.

![Figure 3.1. October monthly climatology temperature profiles from GDEM v2.1 (squares), GDEM v3.0 (triangles), and WOA 2001 (circles) compared to the average monthly October temperature profile from TAO mooring (dashed line) located at 8°N, 137°E in the western North Pacific.]

**3.1.4 Sea Surface Temperatures**

Sea surface temperatures (SST) were obtained from two different sources. Remote Sensing Systems’ Optimally Interpolated (OI) data at a 0.25° resolution blends satellite radiometer data from the National Aeronautics and Space Administration’s (NASA) equatorial satellite Tropical Rainfall Measuring Mission (TRMM) Microwave Imager (TMI) radiometer and NASA’s polar orbiting satellite *Aqua*’s Advanced Microwave Scanning Radiometer for Earth Observing System (EOS; AMSR-E) (Gentemann et al., 2009). Regions of TMI-AMSR-E SST data inhibited by rain, sunglitter, and close to land, and data gaps were filled by interpolation (Reynolds and Smith, 1994). Once available in late 2009, SSTs were also obtained from a globally blended
product of NOAA’s Polar-orbiting Operational Environmental Satellite (POES) Advanced Very High Resolution Radiometer (AVHRR) SST and Geostationary Operational Environmental Satellite (GOES) imager SST (Maturi et al., 2008).

3.1.5 Altimetry Data

Repeat tracks from satellite radar altimetry provided SSHA measurements for the twelve-year period of the in-situ dataset were used in creating SPORTS and the real-time OHC product (Table 3.1). The TOPEX/Poseidon, Jason-1 and Jason-2 satellites have 10-day repeat tracks whereas European remote sensing (ERS) satellite, and ENVironmental SATellite (Envisat) have 35-day repeat tracks. The U.S. Navy Geosat Follow-On-Missions (GFO) has a repeat track of 17 days. Two or more satellites must be active to resolve mesoscale features from the altimetry data (Rosmorduc and Hernandez, 2003). Altimetry data were acquired from the Naval Oceanographic Office and SSHAs were calculated using a background mean SSH from the Collecte Localisation Satellites’ Combined Mean Dynamic Topography (Rosmorduc and Hernandez, 2003; Rio and Hernandez, 2004). For the daily product, SSHA data are provided by the US Navy Data Fusion Center as discussed in Lillibridge et al. (2011).

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<tr>
<th>Altimeter (period)</th>
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<th>'08</th>
<th>'09</th>
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<tr>
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<tr>
<td>Jason-2 (10d)</td>
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</table>

Table 3.1 List of satellites used for SPORTS with dates in operation and repeat track time in parentheses.
3.1.6 In-situ Profiles

More than 267,000 quality controlled in-situ profiles from a 12-year time period (2000-2011) were used in the creation of SPORTS climatology (Figure 3.2). Profiles were taken from Expendable Bathythermographs (XBTs), ARGO profiling floats, and the Tropical Atmosphere Ocean (TAO) mooring array (Sippican; Singer, 1990; Carval et al., 2010; McPhaden et al., 1998).

Figure 3.2. Position of in-situ profiles from 2000-2011. Green indicates Argo profiles, blue denotes XBTs, and red triangles are TAO mooring locations.

1) Expendable Bathythermographs

Expendable Bathythermographs (XBTs) are ship-deployed instruments that measure temperature with a thermistor as it falls through the water column. A thin copper wire transmits the measurements back to the ship. A satellite then relays the information from the ships to receiving stations on land. The thermistors on the older XBTs have an accuracy of ±0.5°C, while the newer models have an accuracy of ±0.1°C. The probe is precision weighted and spin-stabilized, allowing for ±2% depth accuracy estimations in the upper 200m from the fall rate (Sippican; Singer, 1990).

XBT data from the Atlantic Oceanographic and Meteorological Laboratory were processed, checking for unrealistic sea surface temperatures and inversions. Data collected from high density XBT transects provide spatial variability of the upper ocean thermal structure. Repeat XBT transects contributes an added perspective of temporal
variability in the upper ocean thermal structure to examine seasonal variability in the ocean structure.

(2) Tropical Atmosphere Ocean Moorings

Forty-two tropical North Pacific moorings from the NOAA Tropical Atmosphere Ocean (TAO) array were used in assessing SPORTS (McPhaden et al., 1998). Daily sea surface and subsurface temperatures are measured at 10 to 15 different depths in the upper 500m with a concentration of measurements in the upper 100m. The thermistors measure with a resolution of 0.01°C and an accuracy of ± 0.02°C (Freitag et al., 2005).

The Pacific Marine Environmental Laboratory (PMEL) performs an extensive quality control of the data. Changes in temperature greater than 5°C and vertical gradients between adjacent sensors are flagged. Any temperature readings outside 3 standard deviations from the 90-day mean are also flagged. PMEL also performs a visual inspection of the profile. Data from the TAO moorings provide an evolving Eulerian prospective of the upper ocean thermal structure.

(3) ARGO Profiling Floats

Approximately 3000 ARGO floats make up the global array of free-floating profilers. These instruments measure temperature and salinity in the upper 2000m of the ocean. A standard ARGO profiling mission occurs over ten-day intervals. The profiler deflates an external bladder, allowing the instrument to descend from the surface to a depth of 1000 m where it “parks”. For approximately ten days the instrument drifts in the ocean currents. The bladder then deflates further, allowing the profiler to sink to a depth of 2000m, before inflating and causing the instrument to rise through the water column again. On the return trip to the surface, the profiler records measurements of temperature,
pressure, and salinity at approximately 200 depths. Conductivity ratios produce salinity measurements with an accuracy of ±0.01 psu. Temperature measurements are accurate to ±0.005°C, and depth measurements have an accuracy of ±5m (Carval et al., 2010). Once at the surface, the profiler transmits the recorded data to a satellite. Real-time ARGO data undergo nineteen automated quality control checks to ensure a realistic profile (Roemmich et al., 2009). The ARGO data not only fills in the gaps between the XBT and TAO mooring data, but also provides basin-wide coverage over the North Pacific Ocean.

3.2 Approach

3.2.1 In-situ profiles

Additional quality control procedures were used on all the in-situ data. First, only profiles with data available for D_{20} and D_{26} were considered for the analysis. Profiles were removed if the depth of 20°C isotherm was less than the depth of the 26°C isotherm. Additional quality control checks were preformed, flagging profiles with temperature jumps of greater than 2°C in 2 meters and profiles with SST greater than 40°C. The flagged profiles were visually inspected and, if possible, corrected to a realistic profile or removed from the analysis.

All in-situ profiles were interpolated to 1-m vertical resolution using a piecewise cubic function to provide a more continuous realistic profile. D_{20} and D_{26} were extracted using a linear interpolation between the two surrounding depths of the isotherm of interest. MLD was found by taking the depth where the temperature varied by 0.5° from the SST (Monterey and Levitus, 1997). This threshold temperature difference was also used because it was outside the precision of the instrument. A threshold temperature difference was used instead of a temperature gradient to find the MLD due to the large
spatial variation of the structure of the thermocline (de Boyer Montegut et al., 2004). OHC was calculated using a trapezoidal integration from $D_{26}$ to the surface

$$OHC = \int_{D_{26}}^{\text{Sfc}} c_p \rho_{T,S}(T - 26^\circ C)dz$$

...(3.1)

Density was calculated using temperature and pressure taken from the instrument and an assumed salinity of 35 psu in the Equation of State (EOS-80) polynomial (Millero et al., 1980).

### 3.2.2 Objective Analysis

SSHA track data were acquired from at least two satellite missions to resolve mesoscale ocean processes (Rosmorduc and Hernandez, 2003). An objective analysis (OA) was required to fill the gaps between the track paths. SSHA from 5 days before and 5 days after the day of interest were used in the analysis for optimal coverage by 10-day altimeters. The Mariano and Brown (1992) OA scheme uses a parameter matrix algorithm to grid non-stationary fields using time-dependent correlation functions. The temporal and spatial correlation scales inputs for the parameter matrix where chosen to be the same as used by Mainelli-Huber (2000) (Table 3.2). The parameters produced a reasonable SSHA field.

<table>
<thead>
<tr>
<th>Operational mode</th>
<th>Dynamic Height</th>
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<td>Roughness parameter</td>
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<td>Tension</td>
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<tr>
<td>Spline</td>
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<td>Influential points</td>
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<td>E-folding scale</td>
<td>$1.2^\circ$</td>
</tr>
<tr>
<td>Temporal decay scale</td>
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</tr>
<tr>
<td>Confidence level</td>
<td>0.95</td>
</tr>
</tbody>
</table>

Table 3.2. Temporal and spatial correlation scale inputs for OA parameter matrix (Mainelli-Huber, 2000).

Zonal and meridional drift velocities were added to the OA parameters to improve the SSHA field. The basin was separated into $10^\circ$ by $10^\circ$ boxes to account for the spatial variability of drift velocities (Chelton and Schlax, 1996; Meyers et al., 2014). Three years
(2002-2004) of OA’ed SSHA data with no drift velocities correction were used to make Hovmöller diagrams (Figure 3.3). Slopes of three different feature trends were averaged together to determine the drift velocities in each box. East to west cross-sections were taken at 7°N in the equatorial region to avoid the complexities of the equatorial waveguide. Each of the boxes’ averaged zonal and meridional propagation speeds were consistent with the Fu’s (2009) zonal and meridional velocities based on twelve years of satellite data and model simulated drift velocities. Additionally, the zonally averaged westward phase speed of nondispersive baroclinic Rossby waves (Fu et al., 2010) was within one standard deviation of the zonally averaged drift velocities from the Hovmöller diagrams. One standard deviation of the zonally averaged drift velocities calculated from the Hovmöller also fell within Chelton et al.’s (2011) range of drift velocity computations from zonally averaged westward propagating eddies, from the 25th to the 75th percentile. Using these drift velocities, the OA scheme recalculated more accurate SSHA values over the twelve-year period (2000-2011).

Figure 3.3. Hovmöller diagram of SSHA of an east to west cross section taken at 25°N between 130°W and 140°W.
3.2.3 *Daily Climatology*

The GDEM v2.1 and WOA 2001 climatologies provided the necessary historical 4-dimensional oceanic data set for implementation of the 2.5-layer model used to derive ocean thermal profiles. A piecewise cubic Hermite scheme interpolated temperature and salinity profiles from the two climatologies to a 2-m resolution and identified the approximate depths of the 20 and 26°C isotherms. MLD was then found by determining the depth at which the profile temperature deviated from the SST by 0.5°C (Monterey and Levitus, 1997). A temperature gradient threshold was not used to define the MLD due to the strong spatial variability of the thermocline (de Boyer Montégut et al., 2004). To make daily climatology from the monthly values, a 15-day running mean was applied as described in Meyers et al. (2014) to smooth monthly transitions. Calculations were made to include an additional year day for leap years. The daily climatology provided the background ocean thermal profile needed in the 2.5-layer reduced gravity model.

3.2.4 *2.5-Layer Reduced Gravity Model*

Satellite SSHA and SST and a background climatological thermal profile were used in a reduced gravity 2.5-layer model to estimate OHC at each 0.25° grid point. The creation of SPORTS climatology as the background climatology will be described in section 3e. The reduced gravity model depends on a density difference between the upper and lower layers. Reduced gravity here is defined as $g'$:

$$g' = \frac{g(\rho_a - \rho_b)}{\rho_a}$$

...(3.2)
where $g$ is the acceleration due to gravity ($9.81 \text{ m s}^{-2}$), $\rho_2$ is the density in the lower layer and $\rho_1$ is the density in the upper layer (O’Brien and Reid, 1967; Kundu, 1990; Goni et al., 1996). Density was calculated from the climatological temperature and salinity profiles. An arithmetic mean was used to estimate the density in the upper and lower layers. The 20°C degree isotherm was used as the interface between the lower and upper densities since this isotherm is typically found within the thermocline. Changes in SSHA are reflected on the interface in this model. Positive SSHA correlate with a deeper interface depth and negative SSHA correlate with a shallower interface depth. The depth of 20°C isotherm ($D_{20}$) is given by

$$D_{20} = \overline{D}_{20} + \frac{g}{g'} \eta$$

...(3.3)

where $\overline{D}_{20}$ is the SPORTS climatological value of the depth of the 20°C isotherm and $\eta'$ is the SSHA. The SSHA is updated daily; therefore the $D_{20}$ is also updated daily through the equation above. The model assumes uniform stretching and shrinking in the upper water column. Therefore, MLD and $D_{26}$ can be estimated in a straightforward manner given these ratios relative to the depth of the 20°C isotherm

$$D_{26} = \frac{\overline{D}_{26}}{\overline{D}_{20}} D_{20}$$

...(3.4)

$$\overline{MLD} = \frac{\overline{MLD}}{\overline{D}_{20}} D_{20}$$

...(3.5)

where $\overline{D}_{26}$ and $\overline{MLD}$ are the SPORTS climatological value of $D_{26}$ and MLD respectively. An adjusted value of MLD was used over the climatological value as the adjusted value proved to be closer to reality where errors ranged between 4% to 7%
(Meyers et al., 2014). These calculations were made at each grid point at a quarter degree resolution.

Assuming a homogeneous mixed layer with respect to temperature, from satellite derived SST and a constant temperature gradient below the base of the mixed layer, OHC was calculated using the following equation, adjusted from (2.1).

\[
OHC = \frac{1}{2} \rho c_p (D_{26} + MLD)(SST - 26^\circ C)
\] …(3.6)

In-situ values of \(D_{20}, D_{26}, \) MLD, and OHC were then compared to satellite estimations of these values at the nearest grid point over a 12-year period (2000-2011). Only satellite estimations with OA error values less than 0.5 were used to ensure quality results.

### 3.2.5 Creating the SPORTS Climatology

Statistical analyses were used to determine the accuracy of the two climatologies used to create the SPORTS climatology. A regressional analysis showed correlation statistics between the satellite derived variables and the in-situ measurements. Optimally, the slope of the regression line will be unity with a small bias determined from the least squares fits. Root mean squared deviations (RMSD), which measure the differences between in-situ observations and satellite obtained values, provided an estimator for accuracy.

\[
RMSD = \sqrt{\frac{\sum_{i=1}^{n} (x'_i - x_i)^2}{n}}
\] …(3.7)

The \(x'_i\) indicated estimated values of \(n\) observations. Normalized values (NRMSD) were used to account for variability of the dataset and to avoid bias errors.

\[
NRMSD = \frac{RMSD}{x_i}
\] …(3.8)
To account for spatial variability, the North Pacific Ocean basin was divided into 5°x5° subregions to assess the accuracy of the GDEM and WOA climatologies. The subregions allowed for an optimal number of in-situ observations per box, to determine the RMSD of each climatology. The edges of each box were then weighted with a 2° linear transition zone to smooth abrupt changes across the borders. Note that the 2° buffer zone was larger than the Rossby radius of deformation to avoid possible distortion of features as they propagated between adjacent boxes. Seasonal oceanic changes were accounted for by determining RMSDs for two time periods, tropical cyclone season (May 1 – November 30) and off season (December 1 – April 30). These RMSD weights determined the final value that made up the SPORTS climatology which

$$x_{SPORTS} = \frac{x_{GDEM} \text{RMSD}^2_{WOA} + x_{WOA} \text{RMSD}^2_{GDEM}}{\text{RMSD}^2_{GDEM} + \text{RMSD}^2_{WOA}}$$ ...(3.9)

was used in concert with satellite altimetry in the 2.5-layer model to calculate the optimal estimation of OHC (Meyers et al., 2014). If a subregion contained less than 50 in-situ observations, the climatologies were weighted equally.

### 3.2.6 GDEM and WOA Weighting Maps

The weighting maps resulting from the RMSD analysis depicted the regional and temporal strengths of each climatology. Weighting maps were made for $\overline{D_{20}}$, $\overline{D_{26}}$, and $\text{MLD}$. Since $\overline{D_{20}}$ represents the depth of the interface between the two density layers in the model, the weights for $\overline{D_{20}}$ were also used for the upper and lower layer density. The regional performance of the GDEM and WOA climatologies can be seen in the weighting maps (Figure 3.4). During the tropical cyclone season, a mixture of weights for the GDEM and WOA climatologies over the North Pacific Ocean show a slight preference
for the WOA climatology. The GDEM climatology dominates in the eastern North Pacific for $\overline{D_{20}}$, $\overline{D_{26}}$, and $\overline{MLD}$ as found by Shay and Brewster (2010). However, the South China Sea (SCS) favors the GDEM climatology completely for $\overline{D_{20}}$ and mostly for $\overline{MLD}$. There is a strong favoring of the WOA climatology for the $\overline{MLD}$ in the western North Pacific.

Figure 3.4. Weighting maps derived from RMSD analysis for climatological $D_{20}$, $D_{26}$, and MLD for tropical cyclone season and off season. The bluer squares favor WOA climatology and the redder squares favor the GDEM climatology.

Transitioning from tropical cyclone season to off season, a slight shift southward of the weights can be seen as the seasonal oceanic changes move isotherms south. During the off season, the mixture of GDEM and WOA climatologies can be seen again across all variables. For $\overline{D_{20}}$, the South China Sea still shows a preference for the GDEM climatology. However, the eastern North Pacific no longer exhibits the GDEM preference. The central North Pacific and South China Sea favor the GDEM climatology for $\overline{D_{26}}$ during the off season. GDEM climatology is stronger in the eastern North Pacific for $\overline{D_{26}}$. $\overline{MLD}$ strongly favors WOA in the western North Pacific and slightly
favors GDEM in the eastern North Pacific for the off season. These weights were used in equation 21 to blend WOA 2001 and GDEM v2.1 into the SPORTS climatology for TC season and off season.
Chapter 4. Assessment of the SPORTS OHC

4.1 Basin-wide Analysis

Once the creation of SPORTS was completed, daily OHC was recomputed over the same twelve-year period (2000-2011) using the new blended climatology and the same drift-velocity corrected SSHAs and satellite SSTs. Regression analyses were performed on SPORTS and *in-situ* variables, $D_{20}$, $D_{26}$, MLD, and OHC (Figure 4.1, left panel). The OHC and $D_{20}$ regressions are the most closely correlated with slopes close to unity. The slope of the regression strays the farthest from unity with the MLD, but still has a high correlation. The high correlations are similar with the results of Meyers et al. (2014) in the North Atlantic basin. The $D_{20}$ regression slope is close to unity with some *in-situ* depths trending significantly deeper than the SPORTS estimated depths. A closer examination of the $D_{26}$ regression showed that this signal is due to the presence of El Niño in the tropical North Pacific. An El Niño occurred during the months of July through December of 2002, according to Climate Prediction Center’s Oceanic Niño Index (ONI) (Smith et al., 2008). Regression analyses of the 2002 *in-situ* $D_{26}$ against SPORTS $D_{26}$ revealed the anomalously deep *in-situ* $D_{26}$ to only be present during the El Niño months and confined to tropics, below $5^\circ$N. A similar result was found when the ONI identified an El Niño in 2004, 2006, and 2009.

Absolute differences between the satellite-derived variables using SPORTS and the *in-situ* variables were normally distributed (Figure 4.1, right panel). The $D_{20}$ differences were fairly evenly distributed about zero with 84% of the SPORTS estimations falling
Figure 4.1. Density distribution (left) of SPORTS derived variables against *in-situ* variables. The white dashed line represents the 1:1 line. Histograms (right) for the difference between SPORTS and *in-situ* derived variables.
within 20 meters of the observed values and 97% within 40 meters (Table 4.1). $D_{26}$ showed a slight negative bias with 69% of the SPORTS estimated values within 15 meters of the observed values and 91% within 30 meters. MLD had a slight positive bias with 66% of satellite calculations falling within 15 meters of the observed and 90% falling within 30 meters. OHC had a slight positive bias with 80% of the SPORTS estimated OHC falling within 20 kJ cm$^{-2}$ of the observed OHC and 90% falling within 30 kJ cm$^{-2}$.

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<td>$\pm40m$</td>
<td>$\pm15m$</td>
<td>$\pm30m$</td>
<td>$\pm20kJ/cm^2$</td>
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<td>84%</td>
<td>97%</td>
<td>69%</td>
<td>91%</td>
<td>66%</td>
<td>80%</td>
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Table 4.1. Percent of SPORTS estimated values falling within a range of the observed values.

NRMSD helped to better identify the regional areas of uncertainty using the SPORTS climatology with the 2.5-layer model. RMSD is normalized by the average of the in-situ observation in the local 5$^\circ\times$5$^\circ$ subregion (Eq. 3.8). High (low) values of NRMSD correspond with greater (less) uncertainty. All of the variables showed a significant uncertainty in the western boundary current regime where the meandering current is more variable than elsewhere (Figure 4.2). Another region of high uncertainty in all the variables was the Philippine Sea due to the presence of a strong eddy field (Qiu, 1999; Roemmich and Gilson, 2001; Hwang et al., 2004). $D_{26}$ also showed high uncertainty in the eastern tropical Pacific where a cold tongue exists (Dijkstra and Neelin, 1995) with varying strength depending on the phase of the El Niño/Southern Oscillation (Bjerknes, 1969). The MLD had the most uncertainty of all the variables and had a strong impact on the uncertainty seen in the OHC field.
Figure 4.2. NRMSD of $D_{20}$, $D_{26}$, MLD, and OHC for tropical cyclone season in the North Pacific Ocean. NRMSD represents the differences relative to the average magnitude of the variable regionally.

4.2 TAO Mooring Comparison – Temporal Evaluation

Measurements from the TAO mooring array provided an Eulerian dataset for comparison to the SPORTS OHC. Data from the central North Pacific mooring located at 9°N 140°W and SPORTS-derived $D_{20}$, $D_{26}$, MLD, SST, and OHC at the same location were averaged over a twelve-year period (2000-2011) to obtain the average daily values (Figure 4.3). Seasonally averaged SPORTS $D_{26}$ was underestimated, and therefore OHC, in the winter and spring. By contrast, in the summer and fall, during the most active part of the tropical cyclone season, SPORTS better estimated $D_{26}$ and OHC. Over the average year, the TAO mooring data fell within one standard deviation from the SPORTS derived
variables. This is especially true in the latter half of the year, during the peak of tropical cyclone season.

A comparison between SPORTS and TAO was made during the strongest El Niño of the SPORTS dataset in 2009 at TAO mooring at 5°N, 265°W. The SPORTS OHC underestimated D26, and therefore OHC, during the El Niño. However, SPORTS was only off by an average of 10 kJ cm$^{-2}$ during this time.

### 4.3 XBT Transect Comparison – Spatial Evaluation

A repeat XBT transect in the western North Pacific allowed for a spatial evaluation of SPORTS calculated OHC. OHC from a repeat XBT transect taken along a ship track from Guam to the Philippines in the Philippine Sea (Figure 4.4, inset) were averaged over the month of September for a three-year period (2008-2010). The OHC was more variable in the western portion of the transect (Figure 4.4, upper). Possible
reasoning for this could correspond to stronger eddies existing in the western Philippine Sea. The shear instability generated by the opposing flows of the Subtropical Countercurrent and North Equatorial Current provided energy for the eddies to intensify as they propagate westward in this region (Qui, 1999). The presence of cold and warm eddy features was apparent with stronger, varying eddies in the western Philippine Sea containing higher levels of OHC variability. Notice the marked agreement between XBT and SPORTS calculated OHC. The XBT OHC was confined to within 2 standard deviations of the SPORTS calculated OHC with similar trends in the OHC values.

Figure 4.4. XBT transect and SPORTS data averaged over a three-year period (2008-2010) for the month of September. Upper panel shows average OHC calculated from SPORTS (black line) with ±2σ and average OHC calculated from the XBT (red). Lower panel shows averaged temperature profiles from the XBT transect with the 26°(20°C) isotherm in black (white). The XBT transect path is shown in the inset in the lower left corner.

4.4 Linear Regression Based OHC Comparison

Pun14 created an OHC product for the western North Pacific Ocean that uses satellite-derived SSHA and SST, and the WOA 2001 climatology with monthly linear regression coefficients to estimate daily OHC. Approximately 38,000 Argo profilers were
used with the WOA 2001 monthly climatological thermal profiles to compute deviations of each isotherm depth from 29°C to 4°C over a 9-year period (2000-2008) during typhoon season (May-October). Depth deviations from climatology were regressed against SSHA to create linear regression relationships. The thermal profiles obtained using this method were used to calculate daily OHC at 0.25° resolution. The results were evaluated with a regression analysis against ARGO data during 2 typhoon seasons (2009-2010). That OHC product can be used to help in the intensity forecast of TCs in the western North Pacific Ocean basin.

The most notable difference between Pun14 OHC and SPORTS OHC is the 2.5-layer model approach by SPORTS, as opposed to the Pun14 linear-regression approach. The linear-regression approach uses monthly regression coefficients with daily SSHA to construct ocean thermal profiles for daily OHC calculations. The 2.5-layer model approach constructs thermal profiles from daily SSHA and daily SPORTS climatological isotherm depth ratios. Daily SPORTS ratios smooth any discontinuities caused by monthly transitions. SPORTS allows for a daily estimation of the MLD in the 2.5-layer model compared to monthly values which is a long time scale in this eddy rich region.

The Pun14 verification dataset excluded the tropics and the highly variable western boundary current regions. Regression analysis of Pun14 OHC against 4,300 ARGO profiles generated an OHC RMSD of 15 kJ cm\(^{-2}\) with a correlation coefficient of 0.95. SPORTS OHC regression analysis against 8,000 ARGO profiles from the same time and region as the Pun14 verification domain generated an OHC RSMD of 14 kJ cm\(^{-2}\) with a correlation coefficient of 0.94.
Chapter 5. ITOP Case Study – Application of SPORTS OHC

The ITOP experiment occurred in 2010 in the western North Pacific Ocean using a variety of ocean and atmospheric instruments to measure typhoon and ocean interactions (D’Asaro et al., 2011; Mrvaljevic et al., 2013). Three typhoons were sampled over the ITOP experiment, providing a wealth of different oceanic and atmospheric conditions to make up a dataset with over a thousand data points. A mooring array was deployed in the tropical western North Pacific Ocean in a region of maximum typhoon frequency for ocean measurements. Atmospheric measurements were taken from aircraft instruments and GPS dropsondes (Hock and Franklin, 1999) deployed during flights to investigate each storm. Ocean measurements were also acquired during and after the storm from an aircraft by deploying floats and drifters. After each storm, vessels were rapidly deployed to the ocean wakes to take measurements using gliders and drifters. The three typhoons sampled were Typhoon Fanapi and Malakas in September and Typhoon Megi in October (Figure 5.1).

Figure 5.1. The storm tracks and intensity of the three ITOP storms.
5.1 Typhoon Fanapi

According to the Joint Typhoon Warning Center (JTWC), Typhoon Fanapi became a tropical storm on 14 Sept in the Philippine Sea near 18 °N, 130°E. Fanapi continued to be a weak storm until it quickly intensified, reaching maximum wind speeds of 105 knots as a category 3 on 18 Sept. Typhoon Fanapi maintained category 3 status as it made landfall in Taiwan on 19 Sept. As the Fanapi moved eastward, the storm degraded and dissipated over China.

SPORTS OHC and SSHA pre-storm conditions can help explain the intensification of Fanapi from an oceanic perspective (Figure 5.2). To examine the accuracy of SPORTS OHC, ARGO float data was examined from 16 Sept and 5 days previous to allow for sufficient data to make an assessment. Regression analyses were done between D20 and D26 from the ARGO floats and the satellite-derived values. Both variables proved to be highly correlated with each other with a slope value of 0.96 for D20 and 0.93 for D26, proving the SPORTS OHC to be an accurate estimate.

![Figure 5.2. Pre-storm conditions of SPORTS OHC with a solid grey contour of 100 kJ cm\(^{-2}\) OHC on the left, and SSHA on the right from 16 Sept for Typhoon Fanapi. Fanapi’s track and intensity changes are shown in each image.](image)
The OHC values are in the moderate range during Fanapi’s intensification, ranging between values of 50 and 100 kJ/cm² in the Philippine Sea prior to landfall. These OHC values, much higher than Leipper and Volgenau’s (1972) suggested 17 kJ cm⁻² day⁻¹, helped provide energy for Fanapi’s intensification. The SSHA field identified regions of strong horizontal thermal gradients, where strong SSHA horizontal gradients indicate strong horizontal thermal gradients. Pre-storm SSHA conditions revealed an area of strong horizontal thermal gradient in the region where Fanapi maintained category 3 status until landfall. The strong air-sea fluxes over the thermal gradient in this region helped to sustain Fanapi as a category 3 storm.

Figure 5.3. The upper panel shows SST before Fanapi (14 Sept) on the left, and after Fanapi (19 Sept) on the right. The bottom plot shows the SST difference of SST before minus SST after Fanapi. Fanapi’s track and intensity changes are shown in each image.
A baseline understanding of the upper ocean thermal response to Fanapi can be found by examining the differences between pre-storm (14 Sept) and post-storm (19 Sept) SSTs (Figure 5.3). The post-storm SST image clearly demonstrates the cold wake left by the storm and the difference image shows a maximum SST cooling of 3° to 4°C to the right of the track. The moderate OHC values allowed for moderate cooling seen at the surface.

Initial momentum responses were calculated from equations 2.2, 2.5-2.7, and 2.9 using JTWC best track and SPORTS data. Values of the momentum flux (τ), isopycnal displacement (η), upwelling velocity (w), oceanic mixed layer velocity (U_{OML}), and Froude number (Fr) were estimated for each track point. At Fanapi’s maximum intensity (18 Sept), with wind speed of 54 m s^{-1}, the ocean experienced a τ of 7.3 N m^{-2} resulting in a η of 5.5 m, w of 0.18 cm s^{-1}, and U_{OML} of 1.34 m s^{-1} (Table 5.1). However, the strongest response to momentum forces was on 17 Sept, while Fanapi was a strong category 1 typhoon. Winds of 41 m s^{-1} created a τ of 4.2 N m^{-2} resulting in a η of 6.3 m, w of 0.23 cm s^{-1}, and U_{OML} of 2.3 m s^{-1} (Table 5.1 in parenthesis). The maximum momentum response occurred at this early date due to a slower translation speed, allowing the Froude number to drop below unity and a barotropic response to dominate.

Enthalpy fluxes were calculated during the time when Fanapi was strengthening from a category 2 to a category 3 storm (17 Sept), using equations 2.3 and 2.4. Aircraft deployed atmospheric dropsondes and oceanic XBTs were used to make these calculations. The dropsondes underwent an automated quality control by the Earth Observing Laboratory’s Atmospheric Sounding Processing Environment system (Martin, 2007). The XBTs were manually quality controlled to ensure no unrealistic temperature
gradients. The latent heat flux and sensible heat flux combined to create a maximum enthalpy flux of approximately 450 W m\(^{-2}\) during Fanapi’s intensification (Figure 5.4).

![Figure 5.4. The momentum flux (a), latent heat flux (b), moisture disequilibrium (\(dq = q_s - q_a\)) (c), sensible heat flux (d), thermal disequilibrium (\(dT = SST - T_a\)) (e), and the total enthalpy flux during Fanapi’s intensification from a category 2 to category 3 TC (17 Sept).](image)

The latent heat fluxes were strongest just outside of the maximum winds and momentum forcing, under developing bands of precipitation. The radial gradient of the moisture disequilibrium, from low in the center to high in the outer reaches of the storm, was responsible for this pattern in the latent heat flux. The sensible heat fluxes were strongest under the regions of where the momentum fluxes were closest to a radial maximum. The regions where strong momentum fluxes correlated with a strong thermal disequilibrium maximized the sensible heat flux.

### 5.2 Typhoon Malakas

The second and weakest storm studied during the ITOP experiment was Typhoon Malakas, which the JTWC reported as tropical storm on 21 Sept. near 18\(^\circ\)N, 145\(^\circ\)E. The storm traveled on mostly west-northwest track until 23 Sept, when Malakas strengthened to a category 1 typhoon with a mostly northward track. Malakas reached maximum intensity on 24 Sept. as a category 2 storm with peak winds of 90 knots. Never making landfall, Typhoon Malakas transitioned to extratropical storm on 25 Sept. near 37\(^\circ\)N, 146\(^\circ\)E.
Pre-storm SPORTS OHC and SSHA field were taken from 23 September (Figure 5.5). The accuracy of SPORTS OHC was verified using ARGO float data from 23 Sept. and 5 days previous to allow for sufficient data to make an assessment. Regression analyses were done between $D_{20}$ and $D_{26}$ from the ARGO floats and the satellite-derived values. Both variables proved to be highly correlated with each other with a slope value of 0.99 for $D_{20}$ and 0.99 for $D_{26}$, proving the SPORTS OHC to be an accurate estimate. Before being classified as a tropical cyclone, Malakas was over OHC over 100 kJ cm$^{-2}$. However, once Malakas reached TC status, the storm passed over low OHC values, at or below 50 kJ cm$^{-2}$, for the duration of it’s life cycle. The SSHA field also did not reveal a strong horizontal gradient during the time Malakas intensified to its maximum intensity and maintained a category 2 status. The absence of high OHC or a strong horizontal thermal gradient made for less than optimal pre-storm conditions for significant TC intensification. The OHC was still high enough to allow for the storm reach and maintain category 2 status.

![Figure 5.5](image). Pre-storm conditions of OHC with a solid grey contour of 100 kJ cm$^{-2}$ OHC on the left, and SSHA on the right from 23 Sept for Typhoon Malakas. Malakas’s track and intensity changes are shown in each image.
Another impact from the low OHC can be found in the cold wake left by Malakas. SST before (21 Sept.) and after (25 Sept) the storm’s passage and the difference reveal a cold wake on the right-side of track with cooling of 5° to 6°C (Figure 5.6). The shallow OML and isotherm depths (low OHC values) in this region allowed for strong SST cooling, even with a relatively weak TC compared to the other ITOP storms, Fanapi and Megi.

Figure 5.6. The left image shows SST before Malakas (21 Sept), and after Malakas (25 Sept) on the middle image. The right image shows the SST difference of SST before minus SST after Malakas. Malakas’s track and intensity changes are shown in each image.

Initial momentum responses were calculated for Malakas, similarly as was done for Fanapi. Malakas’s maximum intensity (24 Sept), with wind speed of 46 m s⁻¹, produced a $\tau$ of 5.3 N m⁻² into the ocean, resulting in a $\eta$ of 3.4 m, $w$ of 0.07 cm s⁻¹, and $U_{OML}$ of 0.85 m s⁻¹ (Table 5.1). However, the strongest response to momentum forces was on 23 Sept, just before Malakas became a typhoon with wind speeds of 33 m s⁻¹. The winds forced a $\tau$ of 2.8 N m⁻² resulting in a $\eta$ of 5.5 m, $w$ of 0.15 cm s⁻¹, and $U_{OML}$ of 1.1 m s⁻¹ (Table 5.1 in parenthesis). Same as Fanapi, the maximum momentum response
occurred at this early date due to a slower translation speed, allowing the Froude number to drop below 1 and a barotropic response to dominate.

The enthalpy fluxes were calculated during the time when Malakas was strengthening from a category 1 to a category 2 storm (23 Sept), using equations 2.3 and 2.4. The atmospheric dropsondes and oceanic XBT data used in the calculations were subjected to the same quality control procedures as the data from the Fanapi storm. The latent heat flux and sensible heat flux combined to create a maximum enthalpy flux of approximately 550 W m\(^{-2}\) during Malakas’s intensification (Figure 5.7).

The latent heat fluxes were strongly dominated by the moisture disequilibrium for Malakas, as opposed to the wind forcing. The winds had more of an impact on the sensible heat fluxes, maximizing the sensible heat flux where the strong wind forcing and thermal disequilibrium where aligned.

5.3 Typhoon Megi

The final and strongest storm studied during the ITOP experiment was Typhoon Megi. On 14 Oct Megi was classified as a tropical storm near 12°N, 139°E by the JTWC. Megi moved west-northwestward, intensifying until it reached peaks winds of 160 knots as a category 5 super typhoon (ST) on 17 Oct. Maintaining ST status, Megi made
landfall in on 18 Oct. Interactions with land weakened Megi (Tuleya, 1994) to a category 2 storm before moving in the South China Sea. Megi, moving much slower than when in the Philippine Sea, trekked a short distance westward before taking a much more northern track on 19 Oct. On 20 Oct., Megi reached ST status as category 4 storm with peaks winds of 115 knots. Megi continued moving slowly northward, maintaining category 3 winds until late on 22 Oct. when it quickly degrading to tropical storm just before making landfall in China.

The pre-storm SPORTS OHC and SSHA conditions from 15 Oct. can explain some of Megi’s intensification (Figure 5.8). To examine the accuracy of SPORTS OHC, ARGO float data was examined from 15 Oct. and 5 days previous to allow for sufficient data to make an assessment. Regression analyses done between $D_{20}$ and $D_{26}$ from the ARGO floats and the satellite-derived values showed both variables proved to be highly correlated with each other with a slope value of 0.96 for $D_{20}$ and 0.91 for $D_{26}$, proving the SPORTS OHC to be an accurate estimate. While in the Philippine Sea, Megi was over high OHC values, above 100 kJ cm$^{-2}$, for the duration of it’s life cycle there. There also existed a strong horizontal thermal gradient under the storm track while Megi was a cat 5 storm before making landfall in the Philippines. The high OHC and strong ocean thermal gradient helped Megi to intensify to and maintain super typhoon status in the Philippine Sea. Once in the South China Sea, Megi intensified to category 4 status over values of high OHC values greater than 100 kJ cm$^{-2}$. After 20 Oct., Megi was over more moderate values of OHC, from 50 to 75 kJ cm$^{-2}$, before making landfall in China. The high OHC values in the eastern South China Sea helped Megi intensify to a super typhoon while the
more moderate OHC values helped Megi maintain a category 3 status over the central South China Sea.

![OHC and SSHA maps with Typhoon Megi's track and intensity changes](image)

**Figure 5.8.** Pre-storm conditions of OHC with a solid grey contour of 100 kJ cm⁻² OHC on the left, and SSHA on the right from 15 Oct for Typhoon Megi. Megi’s track and intensity changes are shown in each image.

Differencing the Pre-Megi (14 Oct.) and post-Megi (22 Oct.) SSTs show a weak cold wake in the Philippine Sea of only 1° to 2°C compared to a strong cold wake in the South China Sea of 6° to 7°C (Figure 5.9). The presence of high OHC (or deep OML) values in the Philippine Sea kept the SST cooling low, even with a cat 5 storm similar to cat 5 storms moving over the Loop Current in the Gulf of Mexico. The strong SST cooling seen in South China Sea can be attributed to the slower translation speed of Typhoon Megi and more moderate values of OHC. The deeper OML and isotherm depths (higher values of OHC) in the far eastern part of the South China Sea slightly reduced the SST cooling in that region.

Initial momentum responses were calculated for Megi in the Philippine Sea and the South China Sea. Megi’s strongest momentum response in the Philippine Sea occurred when Megi was at maximum intensity (17 Oct), with wind speed of 80 m s⁻¹, produced a $\tau$ of 15.9 N m⁻² into the ocean, resulting in a $\eta$ of 8.3 m, w of 0.3 cm s⁻¹, and $U_{OML}$ of 0.78 m s⁻¹ (Table 5.1) associated with a dominate baroclinic response ($Fr > 1$). In
the South China Sea, Megi’s strongest momentum response was dominantly barotropic (Fr <1) when Megi reached its second peak intensity with winds of 59 m s\(^{-1}\) (20 Oct). The resultant \(\tau\) of 8.7 N m\(^{-2}\) caused a \(\eta\) of 9.5 m, \(w\) of 0.5 cm s\(^{-1}\), and \(U_{OML}\) of 2.36 m s\(^{-1}\) (Table 5.1).

![Figure 5.9](image)

**Figure 5.9.** The upper panel shows SST before Megi (14 Oct) on the left, and after Megi (22 Oct) on the right. The bottom plot shows the SST difference of SST before minus SST after Megi. Megi’s track and intensity changes are shown in each image.

The enthalpy fluxes were calculated during the time when Megi underwent rapid intensification from a category 4 to category 5 storm (16 Oct), using equations 2.3 and 2.4. The atmospheric dropsondes and oceanic XBT data used in the calculations were subjected to the same quality control procedures as the data from the Fanapi storm. The latent heat flux and sensible heat flux combined to create a maximum enthalpy flux of approximately 1100 W m\(^{-2}\) during Megi’s intensification (Figure 5.10).
Figure 5.10. The momentum flux (a), latent heat flux (b), moisture disequilibrium \( dq = q_s - q_a \) (c), sensible heat flux (d), thermal disequilibrium \( dT = SST - T_a \) (e), and the total enthalpy flux during Megi’s rapid intensification from a category 1 to category 5 TC (16 Oct).

Megi’s strong winds dominated the latent and sensible heat fluxes. The momentum fluxes caused by these winds were a factor of 3 to 4 times larger than anything seen in Fanapi and Megi. A single dropsonde measurement contributed to the flux maximum seen near the center of the storm. The winds at this point were much stronger than any of the surrounding points measured by the dropsondes, but reasonable considering the JTWC best track maximum winds of the time of the measurements.

5.4 Concluding Remarks

The three ITOP storms occurred over differing ocean environments. Fanapi occurred over moderate OHC and an active ocean eddy field with a strong thermal gradient that helped the TC to intensify. Malakas lacked a strong ocean thermal gradient and occurred over low OHC, but the OHC was enough to sustain a category 2 storm. The large amount of SST cooling associated with the cold wake of Malakas was a consequence of the low OHC and shallow OML. Megi occurred over high OHC and strong ocean thermal gradients in the Philippine Sea and South China Sea, allowing it to reach super typhoon status in both Seas. The high OHC and deep OML kept the SST cooling in the Philippine Sea to a minimum despite the super typhoon winds.
Table 5.1. Upper ocean momentum response to typhoons where $\tau$ is the momentum flux, $\eta$ is isopycnal displacement, $w$ is upwelling velocity, and $U_{OML}$ is oceanic mixed layer current. For Fanapi and Malakas, values with $Fr > 1$ are outside ( ) and values with $Fr < 1$ are inside ( ). PS is the Philippine Sea and SCS is the South China Sea.

<table>
<thead>
<tr>
<th>Typhoon</th>
<th>Max Wind (m s$^{-1}$)</th>
<th>$\tau$ (N m$^{-2}$)</th>
<th>$\eta$ (m)</th>
<th>$w$ (cm s$^{-1}$)</th>
<th>$U_{OML}$ (m s$^{-1}$)</th>
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</thead>
<tbody>
<tr>
<td>Fanapi</td>
<td>54 (41)</td>
<td>7.3 (4.2)</td>
<td>5.5 (6.3)</td>
<td>0.18 (0.23)</td>
<td>1.34 (2.3)</td>
</tr>
<tr>
<td>Malakas</td>
<td>46 (33)</td>
<td>5.3 (2.8)</td>
<td>3.4 (5.5)</td>
<td>0.07 (0.15)</td>
<td>0.85 (1.1)</td>
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<tr>
<td>Megi (PS)</td>
<td>80</td>
<td>15.9</td>
<td>8.3</td>
<td>0.3</td>
<td>0.78</td>
</tr>
<tr>
<td>Megi (SCS)</td>
<td>59</td>
<td>8.7</td>
<td>9.5</td>
<td>0.4</td>
<td>2.36</td>
</tr>
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</table>

The TC-induced ocean momentum responses were impacted by the translation speed as much as the wind speed (Table 5.1). When the Froude number was less than 1 and the barotropic response dominated, slower translation speeds resulted in a greater ocean response at lower wind speeds. Enthalpy fluxes were calculated for this study during the period of intensification to maximum intensity for each storm. For all three TCs, the latent heat flux contributed the most to the enthalpy response for each of the storms. The sensible heat flux was the most affected by the winds during intensification. Regions associated with both strong momentum fluxes and thermal disequilibrium maximized the sensible heat flux. The winds had a similar affect on the latent heat, but to a much lesser extent, as the moisture disequilibrium dominated. Megi was an exception to this trend with the high winds controlling the enthalpy fluxes. The enthalpy responses were weakest in Typhoon Fanapi and strongest in Typhoon Megi during TC intensification, in agreement with Lin et al. 2013. Malakas had a stronger moisture disequilibrium during intensification than Fanapi, resulting in higher enthalpy fluxes during Malakas’s intensification. The super typhoon force winds reached during Megi’s intensification resulted in the highest momentum fluxes, by a factor of 3 to 4 more, over Fanapi and Malakas. Megi’s winds over powered the contributions of the thermal and moisture disequilibrium, giving Megi the highest enthalpy fluxes of the three TCs during intensification.
Chapter 6. Summary and Concluding Remarks

The overarching goal of this thesis was to create an accurate OHC product in the North Pacific Ocean following the methods used by Meyers et al., (2014). To achieve this goal, two objectives focused on:

(i) Creating an OHC product for the North Pacific, and

(ii) Assessing the accuracy of the satellite-derived of the OHC product.

The ITOP 2010 storms provided an opportunity to use SPORTS OHC in a case study to examine how OHC impacted TC intensification and TC induced ocean response.

Objective (i) was achieved with the creation of the SPORTS OHC. A 2.5-layer reduced gravity model produced realistic estimations of OHC at a quarter degree resolution using the SPORTS climatology, along with satellite SSHA and SSTs. The SPORTS climatology was created using a RMSD analysis to determine the weights for blending the WOA 2001 and GDEM v2.1 climatologies. Over 267,000 in-situ profiles provided the comparison points for the analysis to determine the regional and temporal strengths of the GDEM and WOA climatologies. Resulting weights for the $D_{20}$, $D_{26}$, and $MLD$ during tropical cyclone and off season demonstrated that neither climatology dominated the spatial or temporal domain, justifying the blending used for SPORTS for two different seasons.

An assessment using over 267,000 in-situ profiles accomplished objective (ii). The $D_{20}$, $D_{26}$, MLD, and OHC calculated from SPORTS and the 2.5-layer model were carefully evaluated against in-situ data points from a 12-year period in the North Pacific Ocean. Highly correlated basin-wide scatterplot analyses for $D_{20}$, $D_{26}$, MLD and OHC showed that the SPORTS variables were realistic. The largest departure from unity
occurred with $D_{26}$ variable due to the presence of El Niño in the tropical Pacific. The El Niño and other phases of ENSO caused a measure of uncertainty in SPORTS OHC, as identified by the NRMSD analysis in the region where the cold tongue exists in the eastern tropical North Pacific. The Kuroshio Extension Region and the eddy zone in the western North Pacific where other regions of uncertainty identified by the NRMSD analysis. SPORTS OHC may have larger departures from reality in these regions due to the strong mesoscale variability. A TAO analysis in the eastern Pacific at $5^\circ$N, $265^\circ$W during a strong El Niño in 2009 showed that while SPORTS OHC was underestimated during this period, the average difference between SPORTS and *in-situ* was only 10 kJ cm$^{-2}$. Records from a TAO mooring in the central North Pacific provide an Eulerian dataset for analysis of SPORTS OHC, showing a strong correlation throughout the year with the highest correlation during typhoon season. An XBT repeat transect in the western North Pacific proved SPORTS OHC to be accurate spatially, even with the strong variability in the eddy field on the western portion of the transect.

NOAA’s NESDIS use the 2.5-layer model with the SPORTS climatology for operational OHC calculations in near-real time for the North Pacific. The real-time SPORT OHC has the flexibility to ingest improved satellite SST products and new altimetry satellites (Table 6.1). The Satellite with ARgos and ALtiKa (SARAL) was added to the real-time product in October 2013. Currently, the benefits of adding altimetry from the European Space Agency’s CryoSat-2 satellite to the SPORTS and SMARTS SSHA is being investigated. An archive of real-time and past SPORTS OHC, MLD, $D_{26}$, and $D_{20}$ spans 16-years (2000-2015), creating an oceanic dataset that can be used for weather and climatological studies. Daily SPORTS OHC and archived SPORTS
OHC, since becoming operation in 2013, can be found at the RSMAS Upper Ocean Dynamics Lab research website (http://www.rsmas.miami.edu/groups/upper-ocean-dynamics/research/ocean-heat-content/north-pacific/).

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Table 6.1. List of satellites used for SPORTS with dates in operation and repeat track time in parentheses.

Another OHC product in the North Pacific is the Pun14 daily OHC for the western Pacific. The SPORTS OHC uses a 2.5-layer model approach to estimate daily OHC, as opposed to Pun14’s monthly linear regression method. The abundance of in-situ data used to assess the blended SPORTS OHC product makes this product more robust than OHC products using a monthly climatology. The long time range of the SPORTS in-situ data creates an opportunity to improve the SPORTS climatology in future work. According to the ONI, 4 El Niño, 4 La Niña, and 4 neutral ENSO phases occurred during the 12-year period (2000-2011), making for an even distribution of the ENSO phases (Smith et al. 2008). Creating weighting maps for the 3 phases of ENSO during typhoon and off season will improve the representation of ENSO variability in the SPORTS climatology.

A case study of the ONR-sponsored 2010 ITOP storms demonstrated an application of SPORTS OHC. The pre-storm SPORTS OHC and SSHA showed how oceanic conditions contributed to the intensification and maintenance of the three TCs.
Satellite SST cooling between pre-storm and post-storm conditions showed how the OML is important to the TC’s thermal response. The deeper the OML, the greater the OHC will be and the smaller the ocean surface thermal response will be. Momentum responses were calculated at the peak intensity of the three TCs. The strongest $U_{OML}$ and isopycnal displacement were found when the barotropic response was dominant ($Fr <1$). The enthalpy fluxes were calculated for each storm during the time of intensification to the storm’s maximum intensity. The latent heat flux dominated over the sensible heat flux during intensification. Typhoon Megi had the strongest enthalpy fluxes, followed by Typhoon Malakas, while Typhoon Fanapi had the weakest enthalpy fluxes. Megi’s strong winds gave Megi the strongest enthalpy fluxes. Malakas had the strongest moisture disequilibrium of the ITOP TCs during intensification resulting in stronger enthalpy fluxes than Fanapi.

Atmospheric and oceanic data were collected during other phases of each ITOP storm, allowing for further research to explore how OHC impacts the changing enthalpy fluxes throughout each storm’s life cycle. Further investigation of the enthalpy fluxes during the life cycle of a TC will provide a clear understanding of this TC interaction with the ocean to help modelers produce accurate estimates of TC intensification.

In conclusion, the SPORTS OHC provides accurate daily estimates for the North Pacific Ocean. Real-time daily SPORTS OHC is now available for operations at NOAA NESDIS and can also be found at the RSMAS Upper Ocean Lab research website to continue research into the ocean-atmosphere interaction problem. SPORTS OHC can be used on a mesoscale level to investigate how OHC contributes to TC intensification and the TC induced ocean response. On a broader scale, the SPORTS 16-year dataset can be
used to examine how seasonal changes of the North Pacific upper ocean thermal structure impact seasonal TC activity, such as intensity and frequency. This research would serve as a guide for seasonal expectations of TC activity.
REFERENCES


