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Surface Gravity Waves in the Gulf of Mexico and Their Role in Ocean-Atmosphere Coupling

Ehsan Abolfazli
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SURFACE GRAVITY WAVES IN THE GULF OF MEXICO AND THEIR ROLE IN OCEAN-ATMOSPHERE COUPLING

A Thesis

Submitted to the Graduate Faculty of the Louisiana State University and Agricultural and Mechanical College in partial fulfillment of the requirements for the degree of Master of Science in The Department of Oceanography and Coastal Sciences

by

Ehsan Abolfazli B.Sc., Sharif University of Technology, 2009 M.Sc., Amirkabir University of Technology, 2011 May 2019
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Finally, I must thank my dear parents and sister for their everlasting support and encouragement.
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<th>Definition</th>
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<tr>
<td>$f$</td>
<td>Absolute frequency in Hz</td>
</tr>
<tr>
<td>$\omega$</td>
<td>Absolute radian frequency</td>
</tr>
<tr>
<td>$U_x$</td>
<td>Current velocity in the x direction</td>
</tr>
<tr>
<td>$U_y$</td>
<td>Current velocity in the y direction</td>
</tr>
<tr>
<td>$C_{g,x}$</td>
<td>Group velocity in the x direction</td>
</tr>
<tr>
<td>$C_{g,y}$</td>
<td>Group velocity in the y direction</td>
</tr>
<tr>
<td>$La$</td>
<td>Langmuir number</td>
</tr>
<tr>
<td>$C_x$</td>
<td>Propagation velocity in the x space</td>
</tr>
<tr>
<td>$C_y$</td>
<td>Propagation velocity in the y space</td>
</tr>
<tr>
<td>$C_\theta$</td>
<td>Propagation velocity in the $\theta$ space</td>
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<tr>
<td>$C_\sigma$</td>
<td>Propagation velocity in the $\sigma$ space</td>
</tr>
<tr>
<td>$\sigma$</td>
<td>Relative frequency</td>
</tr>
<tr>
<td>$H_s$</td>
<td>Significant wave height</td>
</tr>
<tr>
<td>$u_{st}$</td>
<td>Stokes drift</td>
</tr>
<tr>
<td>$N$</td>
<td>Wave action density</td>
</tr>
<tr>
<td>$W_{age}$</td>
<td>Wave age</td>
</tr>
<tr>
<td>$E$</td>
<td>Wave energy density</td>
</tr>
<tr>
<td>$T$</td>
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</tr>
<tr>
<td>$C_p$</td>
<td>Wave phase speed</td>
</tr>
<tr>
<td>$K$</td>
<td>Wavenumber</td>
</tr>
<tr>
<td>$u^*$</td>
<td>Wind friction velocity</td>
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<tr>
<td>$U_{10}$</td>
<td>Wind speed at 10-m height</td>
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# LIST OF ABBREVIATIONS

<table>
<thead>
<tr>
<th>Abbreviation</th>
<th>Description</th>
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<tbody>
<tr>
<td>CFSR</td>
<td>Climate Forecast System Reanalysis</td>
</tr>
<tr>
<td>COAWST</td>
<td>Coupled Ocean Atmosphere Wave Sediment Transport</td>
</tr>
<tr>
<td>DJF</td>
<td>December-January-February</td>
</tr>
<tr>
<td>JJA</td>
<td>June-July-August</td>
</tr>
<tr>
<td>IFREMER</td>
<td>L'Institut Français de Recherche pour l'Exploitation de la Mer</td>
</tr>
<tr>
<td>MAM</td>
<td>March-April-May</td>
</tr>
<tr>
<td>MDT</td>
<td>Mean Dynamic Topography</td>
</tr>
<tr>
<td>NDBC</td>
<td>National Data Buoy Center</td>
</tr>
<tr>
<td>ROMS</td>
<td>Regional Ocean Modeling System</td>
</tr>
<tr>
<td>SLA</td>
<td>Sea Level Anomaly</td>
</tr>
<tr>
<td>SODA</td>
<td>Simple Ocean Data Assimilation</td>
</tr>
<tr>
<td>SON</td>
<td>September-October-November</td>
</tr>
<tr>
<td>SWAN</td>
<td>Simulating WAVes Nearshore</td>
</tr>
<tr>
<td>WaveSEP</td>
<td>Wave Spectral Energy Partitioning</td>
</tr>
</tbody>
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ABSTRACT

This study provides an overview of the surface gravity wave dynamics in the Gulf of Mexico (GoM) using numerical simulations. The focus is on the effects of ocean currents on waves, and the geographic distribution of a set of wave statistics and parameters related to the role of waves on both sides of the ocean-atmosphere interface. Simulations are performed using the Simulating WAVes Nearshore (SWAN) model with and without coupling with the Regional Ocean Modeling System (ROMS) model within the Coupled Ocean Atmosphere Wave Sediment Transport (COAWST) framework. In the GoM, currents alter the climatological significant wave heights ($H_s$) by up to ±15%. This alteration reduces wave heights in the southwestern GoM and generally increases wave heights in other regions. In two instantaneous snapshots representing the Loop Current variability in terms of its northward extension into the GoM, significant wave heights were modulated by as much as ±35% by the currents. A ray-tracing experiment showed that the wave rays that travel through the northern and the southern margins of the anticyclonic eddies in the GoM are refracted to the left of their direction of motion (southward) because of the negative meridional shear of zonal currents from the radius of maximum velocity towards the eddy boundary. The rays travelling through the core of the eddy are refracted to the right of their direction of motion (northward) because of the positive meridional shear of zonal currents from the center of the eddy toward the radius of maximum velocity. In winter, spring, and fall, the swell fraction increases from east to west in the GoM and reaches as high as 0.8 in the southwestern GoM, off the coast of Mexico. The dominance of swell in this region combined with weak winds results in a higher prevalence of wave-driven wind regime consistently throughout the year. The wind-driven wave regime is prevalent in fall, whereas the wave-driven wind regime is prevalent in summer, when the wind is the weakest. The spatial and temporal variability of the Langmuir
number suggests that the relative contributions of wave-driven turbulence and wind-driven turbulence are variable over the GoM.
CHAPTER 1. INTRODUCTION

1.1. Gulf of Mexico

The Gulf of Mexico (GoM) is a semi-enclosed marginal sea surrounded by the North American continent and Cuba. It provides abundant physical and biological resources and supports a variety of industries such as fisheries, energy, and tourism in the United States, Mexico, and Cuba. The warm surface water in the Gulf of Mexico is favorable for developing and supporting hurricanes. This causes the Gulf of Mexico to be on the path of many major North Atlantic hurricanes that erode and re-suspend sediments (e.g., Xu et al., 2016), cool the ocean surface (e.g., Shay et al., 2000, Walker et al., 2005), and fuel near-surface phytoplankton blooms (Walker et al., 2005). Warm Core Rings (WCRs) can also have a strengthening effect on the hurricanes in the Gulf of Mexico. Study of the passage of the Hurricane Opal (1995) over a WCR in the Gulf of Mexico showed that Hurricane Opal intensified by 17 hPa while passing over the WCR, compared to 7 hPa when no WCR was present (Hong et al., 2000).

The major current in the GoM is the Loop Current, which can extend from the surface to a depth of approximately 1000 meters. It is an integral part of the Gulf Stream Western Boundary Current System and links the Yucatan Current entering the GoM from the Caribbean Sea to the Florida Current within the Florida Straits. The Loop Current and the associated eddies that detach from it are the most energetic currents within the GoM. The Loop Current transports heat, salt, and nutrients northward from the Caribbean Sea to the North Atlantic Ocean via approximately 23–27 Sv (1 Sverdrup (Sv)=10^6 m^3/s) of water at a speed that can reach up to 1.7 m/s (Johns et al., 2002; Sheinbaum et al., 2002; Forristall et al., 2012). It also modulates the outbreak of harmful algal bloom in the West Florida Shelf (WFS) by interacting with the shelf break (Liu, et al., 2016a).
Circulation in the GoM, the extent to which the Loop Current penetrates into the GoM, the process of eddy shedding, and its seasonal and inter-annual variability have been extensively studied using observations and computer models (e.g., Sturges & Leben, 2000; He & Weisberg, 2003; Vukovich, 2007; Alvera-Azcárate et al., 2009; Vukovich, 2012; Oey L.-Y. et al., 2013; Liu et al., 2016b). In a recent study, Weisberg and Liu (2018) used the Self-Organizing Map (SOM) method to characterize the features of the Loop Current with a set of 40 extracted patterns. They found that northward intrusion of the Loop Current into the Gulf of Mexico causes the eastern side of the Loop Current to be displaced westward from the west Florida continental Shelf. Eddy separation can occur up to three times per year. About 45% of the major warm core eddies that separated from the Loop Current during the years 1976–2003 occurred in late winter and spring with an interval period of between 0.5 and 18.5 months (Vukovich 2007). In a satellite observation study, Hamilton et al. (2002) found both cyclonic and anticyclonic eddies with diameters of about 40–50 km over the central slope of the Gulf of Mexico. By examining the vertical structure of the eddies, they found uplifted isotherms (50–100 m) in the center of the cyclones. The interactions between the Loop Current and the eddies and the topography of the Gulf of Mexico affects the distribution of current energy in the water column.

Land loss caused by the reduced sediment load of the Mississippi and Atchafalaya rivers, wave action on the coastline, hypoxia, oil spills, eustatic and local sea level rise, and large human population residing near coastlines have added to the importance of GoM research. The northern GoM, specifically, has been subjected to significant changes during the past several centuries. Over 25% of the deltaic wetlands of the Mississippi Delta have been lost to the ocean (Day et al., 2007). High eustatic and local sea level rise has caused Louisiana coastlines to experience one of the greatest rates of sea level rise in the world (Louisiana Coastal Protection and Restoration
Authority). In this regard, gravity waves are also a critical factor in changing the geomorphology of the coastlines. Waves are the major contributor to coastal erosion because they gradually destroy the vegetation in wetlands and salt marches. A recent study by Rabalais et al. (2018) based on a 27-year observation database indicated that factors such as higher river discharge, an easterly (westward) wind, and reduced wind speed exacerbate hypoxia in the Gulf of Mexico and increase the area of the hypoxic region.

1.2. Surface Gravity Waves

Ocean surface gravity waves in the GoM have received less attention compared to other parameters that describe the state of the oceans such as temperature, salinity, chlorophyll a, and currents. Surface gravity waves are comprised of locally generated wind seas and remotely generated swells. As indicated by its name, the wind supplies the energy for the growth of a wind wave, while whitecapping, depth-induced wave breaking, and bottom friction drain energy from waves and cause them to decay. When wind waves leave their generation zone and no longer receive energy from local winds, they are considered swells. Swells can be very persistent and travel thousands of kilometers before they dissipate. Currently, there is no consensus on the cause for swell dissipation (The WISE Group, 2007). Possible reasons include interaction with ocean turbulence (Babanin, 2011) or interaction with airflow to create wave-driven winds (Harris, 1966). The importance of research on surface gravity waves is not restricted to the fields of navigation, offshore structure design, and coastal erosion because such research can also provide insights on the role of waves in the coupling between the ocean and the atmosphere.

The evolution and fate of surface gravity waves is affected by the ocean currents over which waves travel. Currents can induce shoaling and refraction and cause Doppler shifts in the frequency of the waves (Wolf and Prandle, 1999). Currents can also indirectly affect the waves by changing
the velocity of the surface wind relative to the surface water, which alters the wind energy input to waves. These currents effects on waves change wave characteristics, including their height and direction. Currents effects on waves have been studied in a few ocean regions. In the northern GoM close to the Alabama coast, wave–current interaction modulates wave heights by 30% (Romero et al., 2017). In the mouth of Delaware Bay, strong tidal currents alter significant wave heights by as much as 40% under stormy conditions (Kukulka et al., 2017). In northern Mobile Bay, the current field alters wave heights by as much as 50% (Chen et al., 2005). Off the coast of Western Australia, the Leeuwin Current modulates significant wave height and wave direction of the surface gravity waves by ±25% and ±20°, respectively (Wandres et al., 2017).

Ocean surface gravity waves are driven by the wind and transfer momentum and energy to the ocean when they break (Melville, 1996). Observations have shown that surface waves can also drive the wind (Grachev and Fairall, 2001) when remotely generated swells propagate substantially faster than the local wind. In contrast to wind seas, swells are not directly coupled to the local wind. Inclusion of swell effects on the wind stress and atmospheric mixing has been shown to improve simulated wind speeds (Wu et al., 2016). The degree of coupling between waves and the overlying atmosphere can be quantified by wave age (Hanley et al., 2010)

\[ W_{age} = \frac{c_p}{U_{10}\cos\Delta\theta} \]  

(1.1)

where \( W_{age} \) is wave age, which is a dimensionless parameter, \( c_p \) is peak wave phase speed, \( U_{10} \) is wind speed at the 10-m height, and \( \Delta\theta \) is the angle between waves and the wind. A wind-driven wave regime is identified by an inverse wave age larger than 0.83, while the wave-driven wind regime is characterized by an inverse wave age smaller than 0.15 (Hanley et al., 2010). Hanley et al. (2010) identified prevalent wind-driven wave regime in regions such as the Southern Ocean and the Northern Hemisphere storm tracks and a prevalent wave-driven wind regime in
regions such as the tropical eastern ocean basin off the coast of Southern California. Wave age can also be used to characterize other processes, such as the distribution of breaking waves (e.g., Sutherland and Melville, 2013, 2015) and the role of gas bubbles on air-sea gas transfer (e.g., Brumer et al., 2017; Liang et al., 2017; Deike and Melville, 2018). These processes make it important to understand the spatial and temporal distribution of wave age over the GoM.

Surface gravity waves induce Stokes drift that changes the near-surface currents through the Stokes-Coriolis effect (Polton et al., 2005). They also play an important role in vertical mixing in the ocean surface boundary layer (OSBL) by driving Langmuir circulations through interactions with currents. Vertical mixing not only mediates the exchange of momentum, heat, and trace materials between the ocean interior and the atmosphere, but also modulates horizontal transport and dispersion by controlling the vertical profiles of horizontal currents, material concentrations, and temperature in the OSBL (e.g., D’Asaro et al., 2014; Liang et al., 2018). Inclusion of wave-induced turbulence reduces the bias of the simulated mixed layer depth (MLD) (e.g., Belcher et al., 2012; Li et al. 2016) and sea surface temperature (SST) estimates (e.g., Belcher et al, 2012). These findings suggest that wind speed alone is not sufficient to quantify mixing in the upper ocean. The importance of Langmuir mixing can be quantified by the turbulent Langmuir number \((La)\) (McWilliams et al., 1997)

\[
La = \frac{u^*}{u_{st}}
\]

where \(u^*\) is the waterside friction velocity, and \(u_{st}\) is the magnitude of the surface Stokes drift. Theoretically, the Langmuir number is the ratio between turbulent kinetic energy (TKE) production by current shear and that by the Stokes drift shear in the OSBL. A large \(La\) implies a greater contribution of wind-driven mixing, whereas a small \(La\) indicates a greater contribution of wave-driven mixing in the OSBL. Because most ocean models do not incorporate a
parameterization for Langmuir circulation, knowledge of the mixing effect of waves in the GoM is beneficial to understanding potential biases in ocean model solutions for the GoM.

Existing studies on surface gravity waves in the GoM primarily focus on wave characteristics during individual events such as hurricanes (e.g., Sheng et al., 2010; Hu and Chen, 2011; Huang et al., 2013). To the best of our knowledge, there have been only two studies devoted to the long-term investigation of surface gravity waves in the GoM region (Appendini et al., 2013, Appendini et al., 2018). The former examines the long-term trend of significant wave height and extreme significant wave height (99th percentile) in the GoM using numerical simulations. The latter identifies the Norte events in the GoM and estimates wave power during those events. Nortes refer to the anticyclonic cold fronts that enter the GoM from North America and drive strong northerly winds. Significant wave heights increase during these events even during the prefrontal phase. Huh et al. (1984) showed that significant wave heights increased from 1 m to over 2 m during the prefrontal to frontal passage of the outbreaks on the Northwest Florida Continental Shelf during fall of 1978. The distributions of swell fraction of wave energy, wave age, and Langmuir number, and the effect of currents on waves have not been studied in the GoM.

The goals of the present study were (1) to examine the mean state and variability of surface wave characteristics, such as significant wave height and swell fraction, as well as a number of wave-related parameters important in air-sea coupling and ocean mixing such as wave age and Langmuir number over the Gulf of Mexico, and (2) to investigate how currents affect the surface gravity waves. The objectives were achieved by analyzing a 10-year (2001 to 2010) numerical simulation using a wave model coupled with a circulation model. The remainder of this thesis is organized as follows. Chapter 2 describes the models, the way they are configured, and the data used for model skill assessment. Chapter 3 validates the models using in situ buoy-measured wave
parameters, as well as satellite altimeter-derived significant wave height and sea surface height datasets. Chapter 4 discusses the wave dynamics and current effects on waves in the GoM. Chapter 5 focuses on parameters important in ocean-atmosphere coupling and wave-driven mixing. Chapter 6 provides a discussion and a summary.
CHAPTER 2. DATA AND METHODS

2.1. Model Description and Configuration

Simulations were carried out using the ocean and the wave modules in the Coupled Ocean Atmosphere Wave Sediment Transport (COAWST) modeling framework (Warner et al., 2010). The ocean module is based on the Regional Ocean Modeling System (ROMS) model (Shchepetkin and McWilliams, 2005), and the wave module is based on the Simulating WAves Nearshore (SWAN) model (Booij et al., 1999).

2.1.1. SWAN Model

SWAN Spectral Wave Model Description

The SWAN model is a spectral wave model based on the wave action balance equation (The SWAN Team, 2011),

$$\frac{\partial N}{\partial t} + \frac{\partial C_x N}{\partial x} + \frac{\partial C_y N}{\partial y} + \frac{\partial C_{\sigma} N}{\partial \sigma} + \frac{\partial C_{\theta} N}{\partial \theta} = \frac{S_{tot}}{\sigma}$$  \hspace{1cm} (2.1)

where \( N = E/\sigma \) is wave action density, \( E \) is wave energy density, \( \sigma \) is the relative or intrinsic radian frequency, \( t \) is time, \( \theta \) is the wave direction, and \( C_x, C_y, C_\sigma, \) and \( C_\theta \) are the propagation velocities in the \( x, y, \sigma \) and \( \theta \) spaces, respectively. The wave energy propagation velocity equals the sum of wave group velocity and current velocity (The SWAN Team, 2011)

$$C_x = C_{g,x} + U_x$$ \hspace{1cm} (2.2a)

$$C_y = C_{g,y} + U_y$$ \hspace{1cm} (2.2b)

where \( C_{g,x} \) and \( C_{g,y} \) are waves group velocity components, and \( U_x \) and \( U_y \) are the surface ocean current velocity components, along the \( x \) and \( y \) directions, respectively. On the right-hand side of Equation 2.1, the source and sink terms \( (S_{tot}) \) include the effects of generation by the wind,
dissipations by whitecapping, bottom friction, depth-induced wave breaking, and nonlinear wave-wave interactions.

At wind-wave equilibrium, the source terms are in balance. Waves grow linearly and exponentially with the wind. Linear growth follows from the work of Cavaleri & Rizzoli (1981), and exponential growth follows from the work of Komen et al. (1984). KOMEN whitecapping was applied according to Komen et al. (1984). SWAN performs Quadruplet wave-wave interaction computations using the Discrete Interaction Approximation (DIA) (Hasselmann & Hasselmann, 1985). Triad wave-wave interaction was activated using the Lumped Triad Approximation (LTA) of Eldeberky (1997). Surface wave-breaking was also considered. Bottom friction was activated based on a semi-empirical expression derived from the JONSWAP results for bottom friction dissipation (Hasselmann et al., 1973). Diffraction was also considered. All the default coefficients were used, except for a constant friction coefficient, which was set to 0.019 m²s⁻³ because the GoM has a smoother seafloor compared to the average ocean. The SWAN model has been used in various geographic locations and oceanic conditions (e.g., Ris et al., 1999; Rogers et al., 2007; Iglesias et al., 2009; Huang Y. et al., 2013; Collins III et al., 2015; Akpınar & Bingölbali, 2016; Kukulka et al., 2017).

For the purpose of this study, two quantities, namely swell energy and Stokes drift, were added as additional model outputs.

Stokes drift refers to the wave-phase-averaged Lagrangian velocity in the direction of ocean surface gravity waves (Stokes, 1847). The surface Stokes drift was computed as follows (e.g., Webb & Fox-Kemper, 2011; Breivik et al., 2016):

\[
\bar{u}_{st} = \frac{2}{g} \int_0^{2\pi} \int_0^{\infty} (\cos \theta, \sin \theta, 0) \sigma^3 E(\sigma, \theta) d\sigma d\theta
\] (2.3)

where \(\bar{u}_{st}\) denotes the surface Stokes drift vector.
The original algorithm in the SWAN code for the separation of swell energy from wind sea energy is based on an arbitrary choice of a cut-off frequency such that the spectral bins with frequencies smaller than the cut-off frequency are considered to constitute swells. This approach does not take into account the relative magnitudes of the wave phase speed and the overlying wind speed nor the wind-wave misalignment. We therefore modified the original SWAN code based on the Wave Spectral Energy Partitioning (WaveSEP) method by Tracy et al. (2007) to separate the directional wave spectrum into wind seas and swells. In this method, wave components traveling at phase speeds slower than $U_P$ are considered to be wind seas

$$U_P = C_{\text{mult}} U_{10} \cos(\theta - \theta_{\text{wind}})$$

(2.4)

where $U_P$ is the wind component in the wave direction and includes a wave age factor $C_{\text{mult}}$ that has a default value of 1.7, and $\theta_{\text{wind}}$ denotes the direction of the 10-m wind.

**SWAN Model Configuration**

The computational domain encompasses the Gulf of Mexico (Figure 2.1a, smaller box) with a horizontal resolution of 0.05°×0.05°. A simple geography of the GoM is shown in Figure 2.1b. The bathymetry of the model is obtained from the ETOPO database (Smith and Sandwell, 1997). The wave model is forced by the wind from the three-hourly NCEP Climate Forecast System Reanalysis (CFSR) product (Saha et al., 2010) with the highest available spatial resolution of 0.312°×0.312°. CFSR is a reanalysis product from a global, high resolution, coupled atmosphere-ocean-land surface-sea ice system. The wind field over the sea is extrapolated wherever necessary to obtain the wind data over regions close to the coastline and avoid the use of data from land because data over regions close to the coastline have been shown to better represent the wind near the boundary between the sea and the land (Kara et al., 2007). The simulations were conducted over a 12-year period from 1999 to 2010, during which time the CFSR
The wind is predominantly directed westward over the GoM because it is part of the Trade Wind system (Figures 2.2a-2d). However, eastward winds are also frequent in the northern Gulf of Mexico and have been shown to significant affect the sediment transport at the Atchafalaya-Vermilion Bay (Walker & Hammack, 2000). There is a clear seasonal variability in the mean wind speed, with the strongest winds during winter over most of the Gulf (Figure 2.2a) and the weakest winds during summer (Figure 2.2c). The strongest climatological winter winds (>8.5 m/s) occur over the northwestern GoM. In summer, the lowest wind speeds (Figure 2.2c) occur close to the western coast of Florida (<3.5 m/s). Over a major part of the GoM away from the coasts, the mean wind speed ranges from 3.5 to 8.5 m/s between summer and winter. The mean wind speed maps shown here smooth out the effects of short-term extreme weather phenomena such as hurricane
events in late summer and fall, and the Norte events during September-April (Appendini et al., 2018), which occur frequently over the Gulf.

Figure 2.2. Seasonal climatology of 10-m wind speed (m/s) (color) and direction (arrows) over the years 2001-2010 based on CFSR wind data for a) December-February (DJF), (b) March-May (MAM), (c) June-August (JJA), and (d) September-November (SON)

In the SWAN model, the eastern and southern boundaries are closed. The fact that a model with a similar setup but with an extended domain (Figure 2.1a, larger box) yielded almost identical results at the studied buoys over the GoM (Figure 2.3) showed that the influence of the swells travelling into the GoM from the Atlantic Ocean and the Caribbean Sea was limited. Forty-five-minute time steps were used in the SWAN model. Sensitivity tests showed that further reduction of the time step had little effect on the results. The directional resolution was 10°, and there were
a total of 36 directions. The lowest and highest frequencies in the wave spectrum were set to 0.0335 and 1 Hz, respectively, spaced logarithmically in 25 frequency bins.

Figure 2.3. Time series of $H_s$ (m) in the two models with different domains at 4 buoy locations in year 2010

2.1.2. ROMS Model

ROMS Model Description

Circulation in the GoM was simulated using ROMS. The ROMS model solves the hydrostatic primitive equations in vertical terrain-following coordinates and horizontal curvilinear grids with innovative algorithms for advection, mixing, pressure gradient, vertical-mode coupling, time stepping, and parallel efficiency. The K-profile parameterization is used for the vertical mixing effect caused by boundary-layer turbulence (Large et al., 1994).

ROMS Model Configuration
The ROMS model uses the same horizontal grid as the SWAN model. Vertically, the grid has 40 terrain-following layers. Monthly climatologies of salinity, temperature, currents, and sea surface height from the Simple Ocean Data Assimilation ocean/sea ice reanalysis (SODA) (Carton and Giese, 2008) were used at the open boundaries. The wind forcing for the ROMS model was derived from the CFSR product, the same wind forcing used for the SWAN model.

2.1.3. Model Coupling

In COAWST, the SWAN and ROMS models were coupled using the Model-Coupling Toolkit (MCT) (Larson et al., 2005). The free surface elevation and current fields computed by ROMS were passed to SWAN at user-specified intervals (45-minute intervals were used here) (Warner et al., 2010). In COAWST, computation of the current field ($\vec{U}_c = (U_x, U_y)$) that is passed to SWAN from ROMS is based on the approach by Kirby and Chen (1989), which integrates the near-surface current velocity over a depth that is a function of wave number with a weighting factor that exponentially decays with depth.

In a coupled ROMS-SWAN simulation, currents affect the wave action balance by both changing the group velocities (Equations 2.2a and 2.2b) and modifying the energy flux from the wind to waves through alteration of the relative speed between the wind and the surface water (Warner et al., 2010). Accordingly, energy flux from the atmosphere to the waves was calculated based on a relative wind velocity ($\Delta \vec{U}$), which is calculated as follows (Warner et al., 2010):

$$\Delta \vec{U} = \vec{U}_{10} - \vec{U}_c$$  \hspace{1cm} (2.5)

where $\Delta \vec{U}$ is the 10-m wind velocity vector ($\vec{U}_{10}$) relative to ocean currents.

To assess the effects of currents on waves, a second simulation was conducted that did not include the effects of currents on waves and had only the SWAN module activated within the COAWST framework (hereafter the SWAN-only simulation). Other than the ROMS module being
turned on or off, both simulations had a similar setup and configuration. A third simulation separated the direct effects of currents on waves—including current-induced shoaling (reverse shoaling) when the wind was travelling over opposing (following) currents and wave refraction over horizontally sheared currents—from the effects of currents on the wind field that forced the wave module in the coupled simulation. In that simulation, the SWAN code was modified in a way that the wind field was not affected by the current field (hereafter the semi-coupled simulation).

Table 2.1 provides a summary of the simulations.

### Table 2.1. Summary of simulations

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Direct effect of currents on waves (i.e., ROMS module turned on)</th>
<th>Wind Energy Input from the Wind to Waves</th>
</tr>
</thead>
<tbody>
<tr>
<td>Coupled</td>
<td>Yes</td>
<td>Function of $\Delta \vec{U}$ (Equation 2.5)</td>
</tr>
<tr>
<td>Semi-coupled</td>
<td>Yes</td>
<td>Function of $\vec{U}_{10}$</td>
</tr>
<tr>
<td>SWAN-only</td>
<td>No</td>
<td>Function of $\vec{U}_{10}$</td>
</tr>
</tbody>
</table>

### 2.2. Observational Datasets

Satellite altimetry and buoy datasets were used to assess the skill of the model. Buoy data were provided by the National Data Buoy Center (NDBC). Along-track, quality-controlled satellite altimeter-derived significant wave height ($H_s$) data were provided by IFREMER (L’Institut Français de Recherche pour l’Exploitation de la Mer). The satellite altimeter data includes observations from inter-calibrated missions, including Jason-1, Jason-2, Geodetic Satellite Follow-up (GEOSAT FO), Ocean Topography Experiment (TOPEX), European Remote Sensing Satellite (ERS)-2, and Environmental Satellite (Envisat) that were operational during the time period of 2001 to 2010. Mean dynamic topography data available at the $0.25^\circ \times 0.25^\circ$ resolution were provided by AVISO. Delayed time level-4 sea level anomaly data from multi-mission altimetry.
observations available at the 0.25°×0.25° resolution were provided by Copernicus Marine Environment Monitoring Service.

2.2.1. Analysis of Altimeter-Derived Significant Wave Height Data

The altimeter-derived significant wave height data were provided by IFREMER and were produced by CERSAT (Laboratoire d'Oceanographie Spatiale). The data contain the cross-corrected significant wave height measurements from a number of satellite missions.

To obtain the climatology of significant wave height based on satellite observations, the data pertaining to the period 2001–2010 were downloaded and imported into MATLAB. The data points that fell within the Gulf of Mexico were saved in a matrix. The generated matrix contained longitude, latitude, time of the measurement, and significant wave height. Figure 2.4 shows the track of the satellites over the region.

![Figure 2.4. Track of the satellites over the GoM over the simulation period (2001-2010)](image)

After compiling a matrix of the locations, times, and significant wave heights, each data point was assigned to a location bin in a 0.5°×0.5° grid to enable construction of a complete map of significant wave height climatology (Figure 2.5).
Figure 2.5. Altimeter-derived significant wave height climatology based on the 0.5°×0.5° grid
CHAPTER 3. MODEL VALIDATION

The solutions from the coupled simulation were validated against *in situ* measurements made by 11 buoys operated by the NDBC and the satellite altimetry data.

3.1. SWAN Model Skill Assessment

The $H_s$ time series at buoy stations 42001, 42002, 42003, 42007, 42019, 42020, 42035, 42036, 42039, 42040, and 42055 (see Figure 2.1a for their locations) were used in the assessment of the SWAN solution in the coupled simulation. These buoys cover both the open ocean and the coastal ocean and have a long history of observations over the GoM. The simulated significant wave height values compared well with *in situ* measurements (Figure 3.1). The Pearson correlation coefficients between the simulated and observed significant wave height values over the 10-year analysis period exceeded 0.90, and the absolute biases between the simulations and observations were less than 7 cm (Figure 3.1), where the mean significant wave heights in this period ranged between 0.66 m and 1.31 m during this period at different buoys. This good agreement indicated that the model could successfully reproduce the variations in $H_s$ and was thus able to fill the gaps in measurements.

In addition to significant wave heights, the simulated direction-integrated wave energy spectrum, $\phi(f) = \int_{0}^{2\pi} E(f, \theta) d\theta$, which is the mean squared sea surface displacement height, agreed well with observations (Figure 3.2). Here, $f$ is the wave frequency in Hz and can be written as $f = \omega / 2\pi$, where $\omega$ is the absolute radian frequency. When the wind forcing is sustained, the spectral peak may stay somewhat constant at lower wave frequencies, which is indicative of waves reaching fully developed wind-wave equilibrium. The model (Figure 3.2b) was able to capture the observed (Figure 3.2a) spectral peaks and their changes both quantitatively and qualitatively. For instance, frequency downshifts can be an indication of a sustained wind condition and a developing
sea, whereas frequency upshifts can be due to a recent change in the wind direction. The fact that the energy density in the spectral peaks was smaller in both the observations (Figures 3.2a) and the simulation (Figures 3.2b) during April–June was due to the smaller wind energy input to the waves when the wind was weaker (Figure 3.2c). Although the comparison shown here was for buoy 42001 during a randomly chosen period of January to June 2008, other sample time periods and buoy selections yielded similar results.

Figure 3.1. Scatter plots of simulated $H_s$ (m) against buoy observations at 11 NDBC-operated buoys located in the GoM. Colors indicate the number of observations. The values $b$ and $R$ represent the bias and Pearson correlation coefficient between the simulated and observed $H_s$. The black line represents the 1:1 line. Comparison is based on the entire analysis period for all the data that were available for each buoy in that period.

To assess the model performance using a different dataset, model solutions were also compared to satellite altimeter-derived $H_s$. There was a reasonably good comparison between the buoy-measured and altimeter-derived $H_s$ during the analysis period (Figure 3.3).
Figure 3.2. (a) Observed and (b) simulated direction-integrated energy spectrum $\Phi(f)$ ($m^2 s$), and (c) wind speed and direction during January-June 2008 for buoy 42001

The along-track measurements were mapped to a $0.5^\circ \times 0.5^\circ$ grid to allow for the construction of $H_s$ bias and correlation maps. This relatively low spatial resolution grid was selected to ensure obtaining complete gridded maps. Some cells adjacent to the coastline were excluded because the significantly smaller number of observations in those cells was not sufficient to represent the strong spatial variability in $H_s$ associated with change in water depth in the cells. Bias was smaller in the central and western GoM, where the biases were less than 10 cm, than in the northeastern region close to the WFS (Figure 3.4a). The correlation coefficients exceeded 0.85 over almost the entire GoM (Figure 3.4b). Although the simulated and observed values did not compare well over the Caribbean Sea because of the closed boundary conditions that prevented
swells from traveling into this region from the Atlantic Ocean and the eastern Caribbean Sea, the simulation and observation agreed well within the GoM.

Figure 3.3. Scatter plots of $H_s$ (m) measured by satellite altimeters inside a 0.1°×0.1° box from the location of the buoys against buoy observations

Figure 3.4. (a) Bias (m) and (b) Pearson correlation coefficient between simulated and altimeter-derived $H_s$
3.2. Validation of the ROMS Model

Sea surface height and sea level anomaly were used to characterize the mean geostrophic currents (e.g., the Loop Current) and mesoscale variabilities (e.g., the Loop Current eddies). The fact that the mean sea surface height and the standard deviation of the high-frequency sea level anomaly (SLA) from the ROMS model (Figures 3.5a and 3.5c) compared reasonably well with those from satellite altimetry data (Figures 3.5b and 3.5d) implied that the model was able to successfully capture the mean and mesoscale variability of geostrophic currents in the GoM. The evident water mass with higher mean sea surface height (SSH) represented the region enclosed by the Loop Current (Figures 3.5a and 3.5b); its locations were coincident in the simulated and observed maps. The persistent anticyclonic feature in Figures 3.5a and 3.5b over the western Gulf was one (or more) warm-core ring(s) separated from the Loop Current (Sturges and Kenyon, 2008). The long-term mean, seasonal cycle, and 90- and 180-day running means were removed from the SLA fields to obtain the high-frequency SLA fields. The strong variability in the high-frequency SLA in the northern parts of the Loop Current was likely associated with the fluctuation of the Loop Current, which occasionally extended close to the Louisiana-Texas and Mississippi-Alabama continental shelves (Figures 3.5c and 3.5d). High variability of the high-frequency SLA west of the Loop Current was a result of the Loop Current warm-core rings or eddies, which dominate the dynamics in the western GoM (e.g., Walker, 2005, Cardona & Bracco, 2016). Mesoscale variability west of the Loop Current was stronger in the model compared to altimeter observations, consistent with previous modelling studies (e.g., Xue et al., 2013). This was likely due to the presence of eddies at a scale comparable to or smaller than the resolution of the gridded SLA maps. In this region, the first baroclinic Rossby deformation radius is around 20 kilometers (Chelton et al., 1998), smaller than the resolution of the gridded altimeter-derived SLA maps,
whereas in the region of the Loop Current, the scale of the Loop Current is well resolved by the satellite altimeters.

Figure 3.5. (a) Model-derived mean surface height (m), (b) altimeter-derived mean dynamic topography (MDT) (m), and (c) model- and (d) altimeter-derived standard deviation of high-frequency sea level anomaly (m).

The overall agreement between the model and the observation over the Gulf of Mexico from multiple sources justified the choice of model parameters, model configuration, and lateral and surface boundary conditions.
CHAPTER 4. WAVE DYNAMICS IN THE GULF OF MEXICO

In this chapter, we explore the dynamics of surface gravity waves in the Gulf of Mexico, their seasonal variability, and how they are affected by near-surface ocean currents.

4.1. Significant Wave Height

Significant wave height is one of the most important metrics used to describe ocean surface gravity waves. It is also a measure of the energy waves carry. Together with the wind speed, significant wave height is a factor that mariners should consider when sailing on the sea. In SWAN, significant wave height is calculated as 4 times the square root of the variance of the wave displacement.

Our simulations showed that the significant wave height displayed a strong seasonal cycle over the Gulf of Mexico (Figures 4.1a-4.1d). Mean significant wave height was largest during winter (DJF) (Figure 4.1a) and was smallest during summer (JJA) (Figure 4.1c) over the majority of the Gulf. The seasonality of mean $H_s$ generally followed that of the wind speed ($U_{10}$) (Figure 2.2). The westward direction of the wind over the Gulf caused an evident zonal gradient in mean $H_s$: the mean $H_s$ values were generally the largest in the western part of the GoM and reached as much as 1.8 m in winter (Figure 4.1a). Mean $H_s$ was generally small in the Bay of Campeche, the West Florida Slope, and the West Florida Shelf because of the limited fetch. For instance, the mean $H_s$ dropped below 0.5 m on the West Florida Shelf during summer (Figure 4.1c). Over the majority of the GoM, waves traveled westward. Waves were directed southwestward in the Bay of Campeche in all seasons (Figures 4.1a-4.1d) and northwestward on the West Florida Shelf in spring and summer (Figures 4.1b-4.1c).
4.2. Current Effects on Waves (CEW)

4.2.1. Current Effects on Significant Wave Height

The effects of GoM currents, especially the Loop Current and the mesoscale eddies, on wave heights were illustrated in two snapshots of $H_s$ in the coupled simulation and $H_s$ difference between the coupled and the SWAN-only simulations together with the current field, wave direction, and wind direction vectors (Figure 4.2). Currents have a number of effects on $H_s$. First, analogous to deep-water waves moving into shallow water, when waves travel over an opposing current, the speed of wave energy propagation decreases, and the convergence of wave action density increases the wave height, and vice versa (Booij et al., 1999). Second, an opposite direction of currents and the wind increases the relative wind speed (Equation 2.5), and the wind energy
input to the waves leads to larger amplitude waves, and vice versa. Furthermore, the fact that wave refraction by a highly spatially variable current field converges and diverges the wave energy also results in a change in the wave height. Finally, the aforementioned mechanisms alter the wave energy carried by swell from its generation area, and this alteration causes wave heights to change in the regions that swells travel to.

In the first snapshot (Figures 4.2a-4.2c), the Loop Current extended far north into the GoM, close to the Louisiana-Texas and Mississippi-Alabama continental shelves. The $H_s$ map (Figure 4.2a) displayed a zonal gradient similar to that of the mean $H_s$ (Figure 4.1). The largest current-induced increase in $H_s$, accounting for about 35% of the total $H_s$, occurred along the eastern branch of the Loop Current, where the currents oppose both the waves (Figure 4.2b) and the wind (Figure 4.2c). Significant wave heights were also higher over the northern and eastern areas of the warm-core anticyclonic eddy west of the Loop Current. In these regions, currents have an eastward component that opposes the waves (Figure 4.2b) and the wind (Figure 4.2c). The largest current-induced decrease in $H_s$ occurred over the northern part of the cyclonic eddy in the western Bay of Campeche and the southern area of the anticyclonic eddy west of the Loop Current. The decrease was about 25% of the total $H_s$. This was due to the waves’ travelling over a following current (Figure 4.2b) as well as to the decreased relative wind speed when the currents were in the same direction as the wind (Figure 4.2c). Whereas throughout most of the GoM the sign of the $H_s$ difference could be explained by examining the directions of currents, waves, and the wind, there were some exceptions. For instance, the $H_s$ was larger in the coupled simulation than in the SWAN-only simulation in regions between the two northward and southward branches of the Loop Current, where neither the wind nor the waves opposed the currents. This may have been due to the convergence of wave rays by refraction and/or the increase in swell energy reaching this region.
from where the waves and wind opposed the currents. In the second snapshot (Figures 4.2d-4.2f), the Loop Current retreated far to the south and left the GoM through the Florida Strait right after entering the GoM. The $H_s$ map in this snapshot (Figure 4.2d) also followed that of the mean $H_s$ (Figure 4.1). An evident increase in wave heights occurred northwest of Cuba. The other area of large increase in wave heights was the northern part of the newly detached anticyclonic eddy north of the Loop Current, where currents flowed eastward in association with the northern and eastern margin of that anticyclonic eddy. The greatest current-induced decrease in $H_s$ was apparent on the opposite side (western and southern margins) of the same anticyclonic eddy and on the western part of an anticyclonic eddy in the northwestern GoM.

Figure 4.2. (a) $H_s$ (m) in the coupled simulation, (b) difference in $H_s$ (m) between the coupled and the SWAN-only simulations (color), surface currents direction and magnitude (black arrows) and wave direction (grey arrows) when the Loop Current extended far north. (c) Same as (b) except that the grey arrows indicate wind direction. (d), (e), and (f) same as (a), (b), and (c), respectively, except for when the Loop Current turned east just north of Cuba.
4.2.2. Current Effect on Wind-Generated Energy

The difference in the energy generation due to the wind between the coupled and the SWAN-only simulations (Figure 4.3) was consistent with the misalignment between the currents and the wind in Figure 4.2. The two snapshots in Figures 4.3a and 4.3b are for the same instances as those in Figures 4.2a and 4.2d, respectively. When the Loop Current extended far north (Figure 4.3a), the greatest current-induced increase in the energy generation due to the wind occurred in the eastern branch of the Loop Current, where the currents oppose the wind. This increase in energy generation contributed to the higher $H_s$ in those regions (Figure 4.2b). The greatest current-induced decrease in the energy generation due to the wind was in the western section of the anticyclonic eddy in the western GoM, where $H_s$ was substantially smaller in the coupled simulation. When the Loop Current retreated far south (Figure 4.3b), the positive difference in energy generation due to the wind was largest north of Cuba, whereas the greatest negative difference occurred in the southern and western sections of the anticyclonic eddy in the northwestern GoM.

In addition to its local effect, the modified wind energy generation could influence regions away from its origin through the energy carried by swell. The difference in $H_s$ was mostly positive in the northern GoM close to the Mississippi-Alabama Shelf, the eastern LATEX Shelf, and the WFS (Figure 4.2). However, the fact that a difference in energy generation due to the wind was not evident suggests possible effects of refraction and transport of wave energy from elsewhere.
To separate the effect of modified wind energy input (indirect effect) and that of current-induced shoaling and refraction (direct effects), we calculated the long-term mean $H_s$ difference between the coupled and SWAN-only simulations (Figure 4.4a) as well as that between the semi-coupled and SWAN-only simulations (Figure 4.4b). The fact that the direct effects (Figure 4.4b) were responsible for a high fraction of the decrease in $H_s$ in the northern Bay of Campeche (Figure 4.4a) may have been due to the waves’ travelling in the direction of currents (Figure 4.2). On the other hand, the fact that the $H_s$ difference was considerably smaller in the semi-coupled simulation than in the coupled simulation northwest of Cuba indicated a significant influence on $H_s$ of current-induced modulation of wind energy input to waves. Interestingly, although the direct effects of currents resulted in a negative $H_s$ difference in the semi-coupled simulation in some areas in the Sigsbee Plain (Figure 4.4b), the combined effects resulted in slightly larger wave heights in the coupled simulation (Figure 4.4a). These slightly larger wave heights were likely due to the increased wind energy input to the waves elsewhere that reached this region as swell energy. In the southern Bay of Campeche, the direct effects were responsible for about half of the $H_s$ difference between the coupled and SWAN-only simulations.
Figure 4.4. Difference in the annual climatology of $H_s$ (m) (a) between the coupled and the SWAN-only simulations and (b) between the semi-coupled and the SWAN-only simulations

4.2.3. Ray Tracing Experiment

To have a better understanding of the effect of refraction on the $H_s$ differences shown in Figure 4.2, ray tracing analysis was used to demonstrate how the waves are refracted by the strongly sheared currents in the GoM, including the Loop Current and the eddies. The ray equations are (e.g., Liu et al., 1989)

$$\frac{d\vec{r}}{dt} = \vec{U}_c + \vec{C}_g$$

(4.1a)

and

$$\frac{d\vec{K}}{dt} = -\vec{K} \cdot \nabla \vec{U}_c$$

(4.1b)

where $\vec{r} = (x, y)$ is the location of the wave ray, $\vec{C}_g = (C_{g,x}, C_{g,y})$ is the wave group velocity vector, and $\vec{K} = (K_x, K_y)$ is the wavenumber vector. The absolute frequency is $\omega = \sigma + \vec{K} \cdot \vec{U}_c$, where the intrinsic frequency $\sigma$ is determined by the deep-water dispersion relationship $\sigma^2 = g |\vec{K}|$, where $g$ is the gravitational acceleration. Both current shear and bottom topography induce gradients in wave propagation speed and refraction. Because we were looking primarily at refraction in the open ocean, we neglected bottom topography, which may have an important effect
on wave refraction in the coastal ocean (e.g., Kukulka et al., 2017). The same equations have been used to examine wave refraction by different currents, including Gulf Stream meanders (Wang et al., 1994), circulation in the Gulf of Alaska (Liu et al., 1994), submesoscale frontal currents off California, and the currents at the Loop Current edge in the Northern GoM off the Alabama coast (Romero et al., 2017).

The ray equations were integrated with three different GoM current fields. The first current field was the mean of the currents in the GoM and included primarily the mean Loop Current (Figures 4.5a-4.5c). In the second (Figures 4.5d-4.5f) and third (Figures 4.5g-4.5i) scenarios, the current fields were the same as those in the snapshots in Figures 4.2a and 4.2d, respectively, and were intended to illustrate the effect of high-frequency circulations in the GoM, such as eddies, in addition to the Loop Current. Simulations were carried out for wave rays of three representative wave periods, namely, $T=4, 8,$ and $12$ s. Wave rays of the first two periods ($T=4$ and $8$ s) represent wind sea, and wave rays with a $T=12$ s represent swell. All rays that originated from the eastern GoM and had an initial westward propagation direction, consistent with the dominant westward wave direction in the GoM (Figure 4.1). Because the rays had a westward component ($K_x<0$), it was the sign of the meridional shear of the zonal current that determined the northward/southward refraction of the rays: if the zonal current had a positive meridional gradient, the rays were refracted northward and vice versa (Equation 4.1b). In the case of the climatological currents (Figures 4.5a-4.5c), wave rays that originated between $\sim24^\circ$N and $26^\circ$N were first refracted southward by the eastern branch of the Loop Current and then northward by the western branch of the Loop Current. The rays emanating north of $\sim26^\circ$N were only able to travel through the anticyclonic circulation associated with the retroflection of the Loop Current and were refracted.
southward by current shear. This led to the convergence of rays in the northwestern GoM. The current-induced refraction decreased with increasing wave period.

The refraction pattern was more complex when the instantaneous current fields including high-frequency variabilities, such as eddies, were used in the ray tracing experiment. In the first instantaneous current field, the Loop Current extended close to the Louisiana-Texas and Mississippi-Alabama continental shelves, and there was a large-scale anticyclonic eddy west of the Loop Current (Figures 4.5d-4.5f). The refraction by the Loop Current was stronger than that shown in Figures 4.5a-4.5c because the instantaneous Loop Current was stronger and narrower than the mean state of the Loop Current. Rays from the south of the Loop Current (~23°N) were refracted to their right (northward) towards the northern GoM coast. These rays originated where the wind and waves opposed the currents (Figure 4.3a), and transport of part of the additional wave energy to the northern Gulf led to a substantial current-induced increase of wave height in the northern Gulf, where the current was weak and was not preferentially opposing the wave and the wind. Some with $T=4$ even reached the northern WFS. In contrast, rays from just north of the Loop Current (~28°N) were refracted southward because of the negative meridional shear of the zonal current and reached the coast of the Bay of Campeche. In this snapshot, rays reach a wider area in the GoM coasts than when the mean Loop Current is used. In the snapshot, in which the Loop Current made a right turn north of Cuba into the Florida Strait (Figures 4.5g-4.5i), only the paths of rays originating south of ~24°N were affected by the Loop Current compared to the scenario with a northward extension of the Loop Current (Figures 4.5d-4.5f), in which all the rays were refracted by the Loop Current. Rays from ~24° to 28°N were refracted by the newly detached anticyclonic eddy north of the Loop Current, which significantly affected the paths of rays, even for the fast-travelling swells with $T=12$ s (Figures 4.5i). The rays travelling through the northern
and the southern edges of the anticyclonic eddy were refracted to their left (southward) because of the negative meridional shear of the zonal current from the radius of maximum velocity towards the eddy boundary. The rays travelling through the core of the eddy were refracted to their right (northward) because of the positive meridional shear of the zonal current from the center of the eddy toward the radius of maximum velocity. This was consistent with a previous study on wave refraction by eddies (Mathiesen, 1987).

The southward refraction of the rays by that eddy directed more rays to the Bay of Campeche (Figures 4.5g-4.5i) than was the case when the Loop Current extended northward (Figures 4.5d-4.5f). There were a number of other anticyclonic and cyclonic eddies in this snapshot. The two anticyclonic eddies in the central and northwestern GoM had a similar but weaker effect on rays compared to the anticyclonic eddy north of the Loop Current because of their weaker current shear field. The cyclonic eddies in the western and southwestern GoM refracted the rays to their left (southward) when the rays traveled through their core and to their right (northward) when the rays traveled through their northern and southern edges. Sensitivity experiments with initial ray propagation directions 10° north/south of west showed the same qualitative effects of the Loop Current and eddies on the path of the wave rays (Figures 4.6 and 4.7).

Similar to Figure 4.5, in both cases, the rays were affected greatly by the Loop Current in the first snapshot and by the anticyclonic eddy northwest of the Loop Current in the second snapshot.
4.2.4. Climatology of Current Effect on Significant Wave Height

To have a broader overview of the long-term current effects on the surface wave dynamics over the GoM, the 10-year (2001–2010) mean seasonal climatology of $H_s$ obtained from the SWAN-only simulation was subtracted from the seasonal climatology obtained from the coupled simulation (Figures 4.8a–d). Whereas the largest current-induced change in $H_s$ in the snapshots was more than 0.25 m (Figure 4.2), the mean significant wave height in the GoM could increase by about 0.2 m and decrease by about 0.15 m at difference locations in the GoM in the presence of currents. The largest positive differences occurred at the eastern flank of the Loop Current.
where it flowed towards the Florida Strait and was moving in the opposite direction of both the waves and the wind. Whereas the currents did not preferentially oppose or follow the waves or the wind in the WFS (Figure 4.2), the fact that they had a positive effect on the mean $H_s$ difference implied that swell energy was transported from regions with increased wave energy in the presence of currents and that there was a convergence of the wave energy because of current-induced refraction (Figure 4.5d).

Figure 4.6. Ray tracing experiment for wave rays initially propagating at 10° north of west for (a)-(c) mean state of the GoM currents, (d)-(f) when the Loop Current extends far north, and (g)-(i) when the Loop Current turns east just north of Cuba. Panels on the left correspond to $T=4$ s, middle panels correspond to $T=8$ s, and panels on the right correspond to $T=12$ s. Colors and black arrows show the surface current speed (m/s) and direction, respectively.
Figure 4.7. Ray tracing experiment for wave rays initially propagating 10° south of west for (a)-(c) mean state of the GoM currents, (d)-(f) when the Loop Current extends far north, and (g)-(i) when the Loop Current turns east just north of Cuba. Panels on the left correspond to $T=4$ s, middle panels correspond to $T=8$ s, and panels on the right correspond to $T=12$ s. Colors and black arrows show the surface current speed (m/s) and direction, respectively.

The fact that the mean $H_s$ values were generally higher in the coupled simulation south of the Louisiana-Texas and Mississippi-Alabama continental shelves may have been due to the refraction of wave rays from the south (Figure 4.5). The regions west of the Loop Current were mostly affected by the anticyclonic eddies, or rings, detached from the Loop Current. Drifter observations have shown that, after detaching from the Loop Current, the anticyclonic eddies travel in a mean west-southwestward path in the central GoM basin (Hamilton et al., 1999). This behavior has been confirmed in a self-organizing map analysis of satellite altimetry data (Weisberg and Liu, 2017) and was also apparent in our results (Figure 3.5). This behavior causes the wind and the
waves, which are mainly directed westward (Figures 2.2 and 4.1), to oppose (follow) the currents in the northern (southern) section of the anticyclonic eddies in the western GoM and therefore causes the waves to have greater (smaller) height southeast of the LATEX Shelf (northern Bay of Campeche) in the presence of currents. The convergence of wave rays in the northwestern GoM (Figures 4.5a-4.5c) can also contribute to greater $H_s$, whereas the shadow zone in the western GoM approximately between 22°N and 25°N (Figures 4.5a-4.5c) can contribute to smaller $H_s$ in the presence of currents (Figure 4.8a-4.8d). The positive (negative) $H_s$ difference in regions in the southern (northern) Bay of Campeche could also be a result of the wind and the waves opposing (following) the currents associated with the cyclonic eddies in Figure 4.2 that frequently occur in the Bay of Campeche (Vázquez De La Cerda et al., 2013).

Figure 4.8. Seasonal climatology of difference in $H_s$ (m) between the coupled and the SWAN-only simulations for (a) DJF, (b) MAM, (c) JJA, and (d) SON
In total, the mean significant wave heights were modulated by currents by as much as ±15%. There was no evident seasonality in the $H_s$ difference, whereas $H_s$ displayed a strong seasonal cycle, probably because the seasonal variability in the strength of the Loop Current was not as significant (Rousset and Beal, 2011) as that of the wind speed and $H_s$. 
CHAPTER 5. WAVE EFFECT ON THE ATMOSPHERE AND THE OCEAN

Waves mediate the interaction between the ocean and the atmosphere. In this section, we assessed the significance of the role that waves play in the two sides of the ocean-atmosphere interface in the GoM by examining parameters including, swell fraction, wave age ($W_{age}$), Stokes drift ($u_{st}$), and Langmuir number ($La$).

5.1. Swell

Swells are the non-local wave component and play an important role in the ocean-atmosphere momentum flux because they travel faster than wind seas (e.g., Grachev and Fairall, 2001). Although their amplitude is not generally large, swells typically carry a considerable fraction of the wave energy over large distances (Fan et al., 2014) because their energy decay scales can exceed 20,000 km (Ardhuin et al., 2009). The swell fraction of the wave energy over the GoM displayed a distinct spatial and temporal variability (Figures 5.1a-5.1d). Mean swell fraction was the largest during summer (Figure 5.1c) and was the smallest during winter (Figure 5.1a). Mean swell fraction was as high as 0.8 almost all year around in the southwestern GoM right off the coast of Mexico, mainly because of the dominant westward direction of the wind and the waves (Figures 2.2 and 4.1). The exception to the westward increase in the mean swell fraction was the northeastern GoM during summer (Figure 5.1c), when the mean swell fraction exceeded 0.5 because the wind is weak ($U_{10}$ < 4 m/s) in that region during summer. The sea west of the Yucatan Peninsula is fetch-limited and therefore displayed a consistently low mean swell fraction (~0.3). The average fraction of the swell energy in the wave energy spectrum over the GoM exceeded 0.3 and was less than 0.8.
Figure 5.1. Seasonal climatology of swell fraction in the coupled simulation for (a) DJF, (b) MAM, (c) JJA, and (d) SON

5.2. Wave Age

Here, wave age, as a dimensionless parameter (Equation 1.1) was computed to quantify the relative prevalence of wave-driven wind and wind-driven wave regimes. Wave age can also be used to quantify other processes such as wave breaking and bubble-mediated air-sea gas exchange. When the inverse wave age \( \left( U_{10} \cos \Delta \theta / c_p \right) \) is larger than 0.83, waves are growing mainly by absorbing momentum from the wind, but when the inverse wave age is smaller than 0.15, there are waves fast enough to transfer momentum back to the air. The in-between range, i.e., \( 0.15 < U_{10} \cos \Delta \theta / c_p < 0.83 \), indicates a mixed sea state composed of both swells and wind seas (Hanley et al., 2010). The maps of 10-year inverse wave age climatology in the GoM showed that
the majority of the GoM was in a mixed sea state throughout the year (Figure 5.2a-5.2d). Mean inverse wave age exceeded 0.8 in parts of the LATEX Shelf and the northern WFS during fall and winter and over a large area west of the Yucatan Peninsula during spring and summer. This high inverse wave age suggested a strong coupling of waves to the wind. This strong coupling is the combined result of strong winds, weak swell, and limited fetch. In the southwestern GoM off the coast of Mexico, a more developed sea and low \( U_{10} \) resulted in a small inverse wave age, as low as 0.05, consistent with a swell-dominated sea (Figure 5.1). During summer, a large pool with relatively low inverse wave age (~0.3) extended over the northeastern GoM, approximately collocated with the region with weak wind \( (U_{10}<4 \text{ m/s}) \) (Figure 2.2c) and large swell fraction (Figure 5.1c).

To quantify how often the wave-driven wind regime and the wind-driven wave regime occur in the GoM, the frequencies of occurrence of \( U_{10}\cos\Delta\theta/c_p < 0.15 \) (Figures 5.3a-5.3d) and \( U_{10}\cos\Delta\theta/c_p > 0.83 \) (Figures 5.3e-5.3h) were also calculated. The fact that there was a relatively high frequency (more than 25% of the time) of occurrence of \( U_{10}\cos\Delta\theta/c_p < 0.15 \) during summer over the northeastern GoM (Figure 5.3c) implied a greater prevalence of the wave-driven wind regime. This region coincided with that exhibiting low \( U_{10} \) values (Figure 2.2c) and high swell fractions (Figure 5.1c). The wave-driven wind regime was most frequent (more than 70% of the time) and occurred rather consistently in all seasons in the southwestern GoM, off the coast of Mexico. This pattern was caused mainly by the presence of fast-travelling swell originating from the east (Figure 5.1) and relatively weak winds (Figure 2.2). The fact that the frequency of occurrence of \( U_{10}\cos\Delta\theta/c_p < 0.15 \) dropped below 0.05 over a large part of the GoM during spring, especially south of \( \sim25^\circ\text{N} \), suggested that waves were strongly coupled to the wind.
Figure 5.2. Seasonal climatology of inverse wave age in the coupled simulation for (a) DJF, (b) MAM, (c) JJA, and (d) SON.

Whereas the frequency of occurrence of \( U_{10} \cos \Delta \theta / c_p < 0.15 \) shows what fraction of the time the wind is driven by waves, frequency of occurrence of \( U_{10} \cos \Delta \theta / c_p > 0.83 \) shows how frequent the wind-driven wave regime occurs. The spatial distribution of frequency of occurrence of \( U_{10} \cos \Delta \theta / c_p > 0.83 \) (Figures 5.3e-5.3h) indicated that the wind-driven wave regime was most frequent in the LATEX Shelf and west of the Yucatan Peninsula (more than 60% of the time). The frequency of occurrence of the wind-driven wave regime was greater in fall and winter than in spring and summer and displayed a meridional gradient with a stronger coupling of waves to the wind in higher latitudes. The \( U_{10} \) maps (Figure 2.2) displayed similar meridional variation during...
the same seasons. The fact that the wind-driven wave regime was less prevalent during summer (~10% of the time) was consistent with the winds’ being weakest during summer (Figure 2.2c).

Figure 5.3. Seasonal climatology of frequency of occurrence of $U_{10}\cos \Delta \theta/c_p < 0.15$ in the coupled simulation for (a) DJF, (b) MAM, (c) JJA, and (d) SON, and seasonal climatology of frequency of occurrence of $U_{10}\cos \Delta \theta/c_p > 0.83$ in the coupled simulation for (e) DJF, (f) MAM, (g) JJA, and (h) SON

5.3. Stokes Drift

Stokes drift contributes to the horizontal transport at the ocean surface (e.g., Kenyon, 1969; Weisberg et al. 2017). It can play an important role when swell is strong and is moving in a direction different from the current direction. Mean Stokes drift was generally greater in the northwestern GoM, the Campeche Bank, and north of Cuba compared to other regions in the GoM (Figures 5.4a-5.4d). The seasonal variation in the Stokes drift magnitude in the GoM was to some extent similar to that of $U_{10}$ (Figure 2.2). Mean Stokes drift was the largest and exceeded 0.09 m/s in winter (Figure 5.4a), when the wind was the strongest (Figure 2.2a). Mean Stokes drift was the smallest and dropped below 0.04 m/s during summer, when the wind was weak ($U_{10}$<4 m/s), particularly on the west Florida slope, the WFS, and the LATEX Shelf (Figure 5.4c). The fact that the Stokes drift ranged from 0.03 m/s to 0.10 m/s in the GoM agreed with the calculations by
Clarke and Gorder (2018) using observations from buoys in the GoM. Extreme meteorological conditions can cause stronger Stokes drifts. For instance, Stokes drift exceeded 0.25 m/s during the passage of Hurricane Isaac (2012) in the GoM (Curcic et al., 2016). Mean Stokes drift was dominantly westward, whereas it was directed southwestward in the Bay of Campeche and Campeche Bank throughout the year. It was southwestward on the WFS in fall and winter (Figures 5.4d and 5.4a) and northwestward in summer and spring (Figures 5.4b-5.4c). The Stokes drift mostly aligned with the wind, especially in spring and fall (Figures 5.4b and 5.4d), although a few regions displayed slight misalignments. In winter, the Stokes drift-wind misalignment was the greatest south of the Louisiana-Texas and Mississippi-Alabama shelves (Figures 5.4a), where the Stokes drift was directed to the north of the wind. Although high-frequency wind seas make a higher contribution to Stokes drift, the fact that the Stokes drift in summer (Figure 5.4c) was small enough to be affected by the low-frequency swells in the northeastern GoM (Figure 5.1c) resulted in a misalignment between Stokes drift and the wind.

5.4. Langmuir Number

The significance of wave-driven Langmuir turbulence was assessed using the turbulent Langmuir number \((La)\) (Equation 1.2). Belcher et al. (2012) have shown that more than 90% of the turbulent kinetic energy in the OSBL is from the waves when \(La<0.3\). The fact that the mean \(La\) was largest in winter (Figure 5.5a) and smallest in summer (Figure 5.5c) implied that there was a higher relative contribution of wind-driven mixing in winter than in summer. The peak \(La\) values \((La>0.3)\) occurred in the northern GoM during winter over the regions with strong winds \((U_{10}>8\) m/s) (Figure 2.2a). The fact that low wind speed (Figure 2.2c) and large swell fraction (Figure 5.1c) during summer were favorable for low values of the Langmuir number \((La<0.3)\) over a large part of the GoM implied that wave forcing made a greater contribution during summer, particularly
in the western GoM. In fall (Figure 5.5d), the fact that La values were greater over the eastern part of the GoM, where values exceeded 0.33, highlighted the greater relative importance of wind-driven turbulence compared to the western GoM.

Figure 5.4. Seasonal climatology of Stokes drift magnitude (m/s) (color) and direction (black arrows) in the coupled simulation and wind direction (gray arrows) for (a) DJF, (b) MAM, (c) JJA, and (d) SON
We also examined the frequency of occurrence of $La<0.30$ (Figures 5.6a-5.6d). The importance of wave-driven Langmuir turbulence was variable spatially and temporally in the GoM. The lowest mean frequency of occurrence of $La<0.30$ was observed during winter (~0.2), when strong winds ($U_{10}>8$ m/s) occurred. In spring, a pool with a relatively high frequency of occurrence (>0.5) of $La<0.30$ over the northeastern GoM (Figure 5.6b) coincided with low $U_{10}$ values (Figure 2.2b) and implied a greater role for wave-driven turbulence within the OSBL. The fact that low wind speed (Figure 2.2c) and large swell fraction (Figure 5.1c) during summer were favorable for frequency of dominance of wave-driven turbulence (as high as 0.6) over the majority of the Gulf (Figure 5.6c) implied a greater contribution of wave forcing.
Figure 5.6. Seasonal climatology of frequency of occurrence of $La<0.30$ in the coupled simulation for (a) DJF, (b) MAM, (c) JJA, and (d) SON
CHAPTER 6. DISCUSSION AND CONCLUSION

The Gulf of Mexico is a semi-enclosed sea with a relatively small amount of swell traveling into it from the Atlantic Ocean and the Caribbean Sea (Fan et al., 2014). This small amount of swell causes the spatial and temporal variability in the surface gravity waves in the GoM to be strongly tied to that of the overlying wind field, especially in the eastern GoM. The exception is summer, when the swell fraction exceeded 0.5 south of the LATEX Shelf. The swell fraction was generally larger in the western GoM in all seasons because the prevailing westward direction of the wind caused the western GoM to be a more developed and swell-dominated sea. The east-west gradient in swell fraction was consistent with a previous study using the ERA-40 global dataset (Semedo et al., 2011). High swells along the western GoM coasts can contribute to coastal erosion through re-suspension and transport of sediments away from the nearshore zones (Smith et al., 2010). Although the GoM is not as vast as the major oceanic basins, such as the Southern Ocean or the Pacific Ocean, and the waves in the GoM have a limited fetch and are driven by weaker wind, our results showed that a wave-driven wind regime could still occur frequently in the GoM, especially in the northeastern GoM in summer. Wave-driven turbulence can play an important role in mixing in the OSBL in the Gulf of Mexico. The high spatial and temporal variability in frequency of occurrence of $La<0.3$ in the GoM suggested that the wind forcing and the wave forcing were not at a constant ratio in the GoM. This necessitates the explicit inclusion of a parameterization for wave-driven mixing (e.g., Harcourt, 2015; Sinha et al., 2015; Reichl et al, 2016) in the region, which may improve the simulated mixed layer depth and sea surface temperature estimates in ocean models.

Current-induced modulation of significant wave height reached as much as ±35% in instantaneous wave fields. The differences in wave solutions between the simulations with and
without currents effects were manifested not only in wave characteristics, such as significant wave height, but also in the role of waves in the ocean-atmosphere interface. The currents generally increased the frequency of occurrence of a wave-driven wind regime in the eastern GoM (the Loop Current and east of the Loop Current) and increased the frequency of occurrence of a wind-driven wave regime in the western GoM (west of the Loop Current) (Figures 6.1a-6.1h).

Figure 6.1. Seasonal climatology of difference in the frequency of occurrence of $U_{10}\cos\Delta \theta / c_p < 0.15$ between the coupled and SWAN-only simulations for (a) DJF, (b) MAM, (c) JJA, and (d) SON, and seasonal climatology of difference in the frequency of occurrence of $U_{10}\cos\Delta \theta / c_p > 0.83$ between the coupled and SWAN-only simulations for (e) DJF, (f) MAM, (g) JJA, and (h) SON.

Compared to the SWAN-only simulation, the frequency of occurrence of $La<0.3$ was generally larger in the eastern and northern GoM in the coupled simulation (Figures 6.2a-6.2d). In contrast, the fact that north of the Yucatan channel and in the Bay of Campeche, the frequency of occurrence of $La<0.3$ was generally smaller in the coupled simulation compared to the SWAN-only simulation implied a greater contribution of wind-driven turbulence within the OSBL in these regions in the presence of currents.
Figure 6.2. Seasonal climatology of difference in the frequency of occurrence of $\La < 0.3$ between the coupled and SWAN-only simulations for (a) DJF, (b) MAM, (c) JJA, and (d) SON

In summary, the most important conclusions of this study were:

(1) Ocean currents altered the mean significant wave heights by as much as $\pm 15\%$ in the Gulf of Mexico, and the current-induced modulation of significant wave heights could reach up to $\pm 35\%$ during individual snapshots.

(2) Swell fraction of the wave energy generally increased from the east to the west in the GoM and exceeded 0.8 in the southwestern GoM, off the coast of Mexico. Swell fraction was the largest in summer and smallest in winter.
(3) The Gulf of Mexico was generally in a mixed state in terms of air-sea momentum flux. However, a wave-driven wind regime was prevalent and occurred more than 70% of the time in the southwestern GoM, off the coast of Mexico. The wind-driven wave regime, in contrast, was most prevalent on the West Florida Shelf and west of the Yucatan Peninsula.

(4) The direction of surface Stokes drift closely followed that of the 10-m wind, with slight misalignment in the northeastern GoM, especially in winter and summer.

(5) The wave-driven Langmuir turbulence made a greater contribution in the ocean surface boundary layer mixing in the Gulf of Mexico in spring and summer than in fall and winter. The spatial and temporal variability in the frequency of occurrence of $La<0.30$ suggested that the contributions of wave-driven and wind-driven mixing in the OSBL in the GoM were variable.
REFERENCES


Tracy, B., Devaliere, E.-M., Nicolini, T., Tolman, H. L., and Hanson, J. L. (2007). Wind sea and swell delineation for numerical wave modeling. *Proc. 10th Int. Workshop on Wave Hindcasting and Forecasting and Coastal Hazard Symp.* Oahu, HI, U.S. Army Engineer Research & Development Center, P12


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