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Field Investigation of Wave and Surge Attenuation in Salt Marsh Vegetation and Wave Climate in a Shallow Estuary

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FIELD INVESTIGATION OF WAVE AND SURGE ATTENUATION IN SALT MARSH VEGETATION AND WAVE CLIMATE IN A SHALLOW ESTUARY

A Dissertation

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 Doctor of Philosophy in The Department of Civil and Environmental Engineering

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December 2012
To my parents.
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ABSTRACT

This research investigates and quantifies the effectiveness of salt marsh vegetation in reducing storm-induced waves and surge, and the potential for wetland erosion due to wave action, using field measurements on the Louisiana coast. To quantify wave attenuation and wave energy dissipation by vegetation (*Spartina alterniflora*), wave data were measured along a transect using pressure transducers during two tropical storms. Measurements showed that incident waves attenuated exponentially over the vegetation. The linear spatial wave height reduction rate increased from 1.5% to 4% /m as incident wave height decreased. The bulk drag coefficient estimated from the field measurements decreased with increasing Reynolds ($R_e$) and Keulegan-Carpenter ($K_C$) numbers.

The vegetation-induced wave energy dissipation did not linearly follow incident energy, and the degree of non-linearity varied with the dominant wave frequency. The estimated drag coefficient is shown to be frequency-dependent and is parameterized by a frequency-dependent velocity attenuation parameter inside the canopy. The spectral drag coefficient predicts the frequency-dependent energy dissipation with better accuracy than the integral coefficient.

The probability distribution of zero-crossing wave heights attenuated by vegetation was observed to deviate from the Rayleigh distribution and follow the theoretically derived one-parameter Weibull distribution which depends on local wave conditions only. Empirical relationships are developed to estimate the shape parameter from the local wave parameters.

Field data collected during Tropical storm Ida (2009) and Lee (2011) showed that the surge attenuated at different rates in two estuaries of different topography. Surge reduction by vegetation was more effective on a large marsh.

To quantify the potential for wave action to cause erosion of coastal wetlands, directional wave measurements were collected over a seven-month period. Marsh retreat rates estimated in the study area, using the wave power calculated from the field measurements are on the same order of magnitude of the recent marsh loss monitoring data.

The empirical relationships of vegetation drag coefficient and wave height probability distribution function can be used to improve coastal modeling and to estimate characteristic wave heights for the design of coastal defense structures fronted by large swaths of salt marsh vegetation.
CHAPTER 1: INTRODUCTION

1.1 Marine Coastal Wetlands: Importance and Issues

Marine coastal ecosystems, which include salt and brackish marshes (i.e., coastal wetlands), coral reefs, mangroves, and seagrass beds, are some of the most productive and threatened ecosystems in the world. These systems provide important ecological and economic value (e.g., Halpern et al., 2008; Loltz et al., 2006). However, these systems are in peril. An estimated 50% of marshes, 35% of mangroves, 30% of coral reefs, and 29% of seagrass beds have been either lost or degraded worldwide (Barbier et al., 2011 and references therein). This loss has resulted in a 33% decline in the number of viable fisheries; 69% decline in the provision of nursery habitats such as oyster reefs, seagrass beds, and wetlands; and 63% decline in filtering and detoxification services provided by suspension feeders, submerged vegetation, and wetlands (Worm et al., 2006).

In the world’s major deltaic plains, loss of land and associated wetlands has been estimated to be 95 km$^2$/year over the past 14 years (Coleman et al., 2008). The Mississippi River delta in Louisiana has experienced dramatic wetland loss. Between 1956 and 2006, annual land loss rates ranged from as little as 34 km$^2$/year to as much as 104 km$^2$/year. The average annual land loss rate over this time period was approximately 70 km$^2$/year (Barras et al., 2003). Coastal wetland loss in Louisiana accounts for 80% of the coastal wetland loss in the entire continental United States. The value of this loss to public use is projected to exceed 37 billion USD by 2050 (LCWCRTF, 1998).

On the Louisiana coast, the causes of wetland loss are complex, and both natural and anthropogenic in origin. Natural causes include subsidence from sediment compaction and dewatering, eustatic sea-level rise, growth faulting, isostatic adjustments, halokinesis and erosion due to daily waves and hurricane waves and surge. Anthropogenic causes include channelization of the Mississippi River, canal dredging through the wetlands, and fluid withdrawal (Day et al., 2000; Gagliano, 2003; Morton et al., 2006).

One of the important causes of coastal wetland loss is erosion along the marsh edges resulting from wave action. Analysis by Penland et al. (2000) showed that 26% of the wetland loss in the Mississippi river delta from 1932 to 1990 can be attributed to erosion due to wind waves; the second highest cause of loss is activities related to oil and gas industry, to which 36% of the loss was attributed.

Wind waves also influence sediment re-suspension in the nearshore area, and have been shown to play an important role in the morphological evolution of intertidal regions (Jaramillo et al., 2009; Kineke et al., 2006; Sheremet et al., 2005; Defina et al., 2007; Fagherazzi et al., 2007). Kirby (2000) noted that the shape of the mudshore profile is controlled by tidal currents and, particularly, by wave climate. The erosive action of waves on coastal marshes increases as more of the marsh is converted to open-water, thereby increasing the fetch and wave forces on exposed marsh edges.
On the Louisiana-Mississippi coast, coastal marshes are typically protected by barrier islands. When the barrier islands disappear, so do the marshes because they are exposed to increased wave-induced damage and erosion. Previous studies (e.g., Roland and Douglass, 2005) have found a strong correlation between the level of wave energy and the survival of coastal marshes.

In addition to the continuous action of wind waves, coastal wetlands also experience frequent surge and stronger wave forces resulting from tropical storms and hurricanes. In the last 50 years, the Louisiana-Mississippi coast has been impacted by 14 major hurricanes including Katrina (2005), Rita (2005), Gustav (2008) and Ike (2008). According to some estimates, this region is more than twice as likely to see major hurricanes than the Texas and Florida coasts (Resio, 2007). Hurricanes Katrina and Rita converted 562 km² of wetlands in coastal Louisiana to open water (Barras, 2006). The impact of the devastation caused by the hurricane surge and waves on human life and property along the coast has been enormous. For example, in 2005, after Hurricane Katrina, more than a quarter of a million people were displaced, more than 1,500 people lost their lives, and the property damage exceeded $100 billion (Graumann et al., 2005).

When considering mitigating hurricane impacts, it is generally acknowledged that coastal wetlands provide a natural first line of defense against damage by storm surge and waves (e.g., Lopez, 2009). Recently, Gedan et al. (2011) took a comprehensive look at existing studies to highlight the critical role of wetlands in attenuating storm waves. By one estimate, in the US, coastal wetlands provide $23.2 billion in storm protection services annually (Costanza, 2008).

1.2 Current Knowledge, Needs and Research Goals

Federal and State agencies have committed significant financial resources to maintaining and improving surge/wave reduction and ecological benefits of coastal wetlands through restoration and protection efforts (CPRA, 2012). The goal of this research is to examine the role of coastal wetland vegetation in reducing storm-induced surge and waves, and the physical sustainability of the wetlands in the presence of waves. These topics are studied using data from field investigations carried out in the unique environment of coastal Louisiana.

1.2.1 Wave and Surge Propagation over Marsh Vegetation

To protect communities from storm surge and waves, traditionally, levees and floodgates have been employed. In many situations, this solution has proven costly, and unsustainable, causing unintended ecosystem consequences by disturbing the deltaic processes (Day et al., 2007). There has been renewed interest in capitalizing on the potential of natural coastal wetlands to reduce the impacts of storm surge and waves. Wetland vegetation dissipates wave energy through increased bottom friction and drag within the water column. It also reduces wave set-up that adds to the total storm level (Dean and Bender, 2006). Most studies of wave attenuation in coastal wetlands have been undertaken in controlled laboratory settings (e.g., Augustin et al., 2009; Cox et al., 2003; Kobayashi, 1993), while a few have been carried out in saltmarshes (Möller et al., 1999, 2006) and lake environments (e.g., Lövstedt and Larson, 2010). Field investigations are sparse, making it difficult to reliably interpret laboratory results and extrapolate them to field conditions for practical applications. Research is needed to provide field
measurements of bulk parameters of surge and wave attenuation, and collective resistance to wave forces by wetland vegetation for coastal engineering applications (e.g., Irish et al., 2008). These data can be used to develop a more realistic and physically-based parameterization of vegetation-dependent bottom drag coefficient. The bottom drag coefficient is one of the key parameters of the storm surge and wave models that are currently used for restoration planning, natural resource management and emergency response. The current modeling practice (e.g., Bunya et al., 2010) is to account for wetland frictional effects by specifying Manning’s n coefficients using land-cover definitions from the USGS GAP data (Hartley et al., 2000; Villea, 2005). Little information is available about the drag coefficient for waves under field conditions.

To fulfill these needs, field measurements of waves passing through wetland vegetation are carried out. The collected data sets are used to answer several important questions such as:

What is the nature and extent of wave attenuation offered by marsh vegetation? How do wave characteristics change as waves travel through vegetated marsh under storm conditions? The characteristics examined are the frequency distributions of dissipation, spectral energy and width, and the wave height distribution. To account for energy dissipation through vegetation, existing spectral wave models have used the Mendez and Losada (2004) formulation which assumes a Rayleigh distribution. This assumption needs to be validated under field conditions.

Marsh vegetation also plays a role in tropical storm surge reduction. The potential of wetlands to dampen storm surge has been expressed by empirical rules of thumb based on observation, e.g., storm surge could be reduced by 1 m over an inland length of 14.5 km. However, use of these rules of thumb has been called outdated (USACE 2006). Recent studies point out that such constant rates do not account for transient forcing and local topography (Resio and Westerink, 2008). There have been numerical studies to understand the wave and surge attenuation potential of coastal wetlands (e.g., Wamsley et al., 2009; Wamsley et al., 2010). Vegetation has also been proposed to reduce wave set up (Dean and Bender, 2006). Field data sets, however, are scarce in the current literature. Moreover, existing data sets are limited to relatively small waves (Moller, 2006; Smith et al., 2010). With the help of field measurement, the extent of surge attenuation provided by the coastal marsh is quantified and underlying mechanisms explored in this study.

1.2.2 Wave Climate in Shallow Muddy Bays

Much of the Louisiana coastal wetlands line the periphery of bays and are subjected to the erosive force of the waves generated in this shallow water environment. These forces have not been studied extensively in Louisiana bays. There is a lack of long-term field measurements. One of the goals of the research is to quantify the characteristics of wave environment inside bays in terms of magnitude of wave heights and peak periods, and to examine the wind wave growth. To this end, a program of long-term wave measurement is carried out in Terrebonne Bay and Breton Sound. The existing empirical formulas to predict fetch-limited wind wave growth (Young and Verhagen, 1996) have been derived using data sets from a large, shallow lake in Australia. In the present study, the performance of these formulas in Louisiana bays, with characteristic soft muddy bottoms, is evaluated. The extent of any discrepancy and its causes and implications are investigated. Previously, such discrepancies have been associated with uncertainties in depth, wind variability (e.g., Kahma and Calkoen, 1992), fetch geometry (Donelan et al., 1992) or tidal currents (Battjes et al., 1987). Some studies (e.g., Ardhuin et al.,
2007) have highlighted deficiencies in current formulations when applied to mixed swell-sea conditions.

In spite of the presence of barrier islands, many Louisiana bays are believed to receive offshore swell energy. Swell components will be identified and partitioned in the measured bimodal wave spectra by implementing an appropriate partitioning scheme (e.g., Voorrips et al., 1997, Hanson and Phillips, 2001, Portilla et al., 2009). Related research questions to be pursued are: How often do offshore swells penetrate into the bays, considering the presence of barrier islands? What are the swell characteristics in terms of wave heights, peak periods? To what extent do the swells attenuate as they propagate northwards in the bays? Current research (e.g., Elgar and Raubenheimer, 2008, Sheremet et al., 2011) suggests that the wave dissipation in shallow, muddy environments is strongly coupled to bed-sediment reworking by waves.

1.3 Objectives

The goal of the present research study is to investigate attenuation of waves and surge through coastal marsh vegetation in the field setting for applications related to coastal restoration and risk reduction from tropical cyclones. The objectives fall in two general areas as follows.

The first area is the investigation of wave and surge propagation over coastal marsh during tropical storms. The following objectives aim to fill the knowledge gaps identified.

1. Measure waves during tropical storms at a suitable wetland site with an array of wave gages. Measure biomechanical vegetation properties.
2. Analyze wave spectra to quantify wave height attenuation.
3. Use existing vegetation-induced wave energy dissipation models to estimate bulk drag coefficients.
4. Parameterize bulk drag coefficient with respect to Keulegan-Carpenter number and Reynolds number.
5. Analyze characteristics of frequency-dependent energy dissipation. Develop methodology to improve modeling of spectral dissipation of energy caused by vegetation.
6. Measure surge levels over the duration of tropical storms by placing water level sensors along shore-normal transects. Quantify surge level and propagation speed reduction and investigate effects of resistance by vegetation.

The second area is investigation of the general wave climate inside a shallow bay that has experienced rapid erosion. The specific objectives are:

7. Measure wave climate at a location inside Terrebonne Bay in terms of wave heights and peak periods over several months.
8. Examine the characteristics of wind wave generation.
9. Assess adequacy of existing wave growth formulations.
10. Estimate potential rates of marsh shoreline retreat.
1.4 Organization of the Dissertation

Chapter 1 introduces the importance of coastal wetlands and their interaction with the hydrodynamic environment. Specifically, attention is drawn to the storm reduction benefits derived from these systems, and the potential for their erosion. It then outlines the current state of knowledge regarding these issues, identifies knowledge gaps and describes the objective of the present research with some supporting literature. Additional, detailed literature review is presented in the respective chapters. The research covers two general areas and the chapters are organized accordingly.

Research Area 1: Wave and surge attenuation by salt marsh vegetation during tropical storms

   Chapter 2: Integral wave height attenuation

   Chapter 3: Frequency-dependent energy dissipation

   Chapter 4: Wave height probability distribution

   Chapter 5: Surge attenuation by salt marsh vegetation

Research Area 2: Wave climate in a shallow estuary and erosion potential

   Chapter 6: Wave climate and erosion potential

Finally, Chapter 7 summarizes all findings with recommendations for further research on the problems related to wave and surge attenuation by salt marsh.

The dissertation is organized in the “journal-type” format. Each of the chapters 2-6 represents a prepared or in-preparation manuscript with individual introductions and list of references. Chapters 1 and 7 tie together all the other chapters.

1.5 References


CPRA (2012), Louisiana’s Comprehensive Master Plan for a Sustainable Coast. Coastal Protection and Restoration Authority of Louisiana. Baton Rouge, LA.


CHAPTER 2: WAVE ATTENUATION BY SALT MARSH VEGETATION DURING TROPICAL STORM LEE (2011)

2.1 Introduction

Coastal wetlands have been recognized as a natural defense against damage from storm surge and waves (e.g., Costanza et al., 2008; Dixon et al., 1998; Gedan et al., 2011; Lopez, 2009). There has been increased interest among scientists, engineers, and policy makers in utilizing coastal wetlands to supplement traditional structural measures used to mitigate coastal flooding from storm surge and waves (e.g., Borsje et al., 2011; CPRA, 2012). To assess the effectiveness of coastal wetlands in wave reduction, an improved understanding of wave transformation over vegetation under storm conditions is needed. Existing literature on wave propagation over wetland vegetation consists of several theoretical and experimental studies. Summaries of these studies can be found in Irish et al. (2008) and Anderson et al. (2011).

Dalrymple et al. (1984) presented the first theoretical model of wave energy dissipation assuming plants as rigid cylinders that exert drag force on the monochromatic waves. Kobayashi et al. (1993) presented an approach based on continuity and momentum equations demonstrating exponential wave height decay. The Dalrymple et al. (1984) formulation was extended by Mendez et al. (1999) and Mendez and Losada (2004) for irregular waves. Chen and Zhao (2012) examined existing wave energy dissipation formulations and proposed two new models. The first model was based on the model of energy dissipation of random waves by bottom friction developed by Hasselmann and Collins (1968). The second model was based on the joint probability distribution of wave heights and wave periods. Lowe et al. (2005a) developed a theoretical model of monochromatic wave flow structure inside a model canopy of rigid cylinders based on momentum balance, and demonstrated that wave orbital excursion was the single relevant parameter affecting flow attenuation inside the canopy. Lowe et al. (2007) extended this model to random wave conditions, and evaluated its performance in the field by submerging the artificial rigid cylinder canopy on a reef under random waves. They confirmed that the shorter-wave velocity components penetrate the canopy more efficiently, and result in more energy loss over the same distance, compared with the longer-wave velocity components.

In a controlled laboratory environment, wave propagation through vegetation has been studied by Augustin et al. (2009), Chakrabarti et al. (2011), Dubi and Tørum (1996), Løvås and Tørum (2001), and Stratigaki et al. (2011), among others. Field investigations of waves over vegetation have been carried out in a variety of environments, including salt marshes (Bradley and Houser, 2009; Cooper, 2005; Möller et al., 1999; Möller and Spencer, 2002; Möller, 2006; Mullarney and Henderson, 2010), coastal mangrove forests (Mazda et al., 2006; Quartel et al., 2007), and vegetated lakeshores (Lövstedt and Larson, 2010). All of these studies show varying degrees of wave attenuation, depending on the vegetation types and wave environment. Wave attenuation by salt marshes has been reported to be anywhere from 50% (Möller et al., 1999) to 100% (Cooper, 2005) greater than that over mudflats. In coastal mangroves, wave attenuation has been reported to be 5 times more than that due to bottom friction alone (Quartel et al., 2007).
In general, waves over vegetation have been observed to decay exponentially with the distance travelled.

Some of these studies have explored specific mechanisms related to wave-vegetation interaction. For example, Bradley and Houser (2009) examined the role of oscillatory seagrass blade movement in wave attenuation. At lower orbital velocities, blades were observed to sway over the entire wave cycle while at higher orbital velocities, the blades extended in the direction of flow for the longer part of the cycle, becoming streamlined, which resulted in reduced drag and therefore lesser attenuation. Mullarney and Henderson (2010) derived and field-tested an analytical model for the wave-induced movement of single-stem vegetation treated as an Euler-Bernoulli problem for a cantilevered beam. During field tests, vegetation stem motion was observed to lead water motion. The phase difference of motions decreased with increase in wave frequency. For moderately flexible stems, the model predicted total wave energy dissipation equivalent to about 30% of the dissipation for an equivalent rigid stem. Riffe et al. (2011) applied this flexible vegetation model to demonstrate improvement in the predicted wave energy dissipation when vegetation motion is simulated. Lövstedt and Larson (2010) examined wave attenuation and transformation of wave height distribution by reeds in a shallow lake. In their study, a commonly assumed Rayleigh distribution for random variation in wave height was observed to change significantly only under conditions of longer wave propagation distances and higher waves.

Although several field studies have quantified the rate of wave height attenuation and demonstrated the utility of wetlands as a measure for reducing impacts of waves, they have been carried out in low-energy environments with exception of Möller and Spencer, (2002). Smith et al. (2011) describe the challenges of measuring storm induced waves in coastal wetlands based on their attempt during Hurricane Gustav (2008), and emphasize the need of such measurements. Table 2.1 shows ranges of wave heights under which some of the more recent field studies were performed. The table also lists some laboratory studies that have developed empirical relationships for drag coefficients. The validity of extrapolating these results to a high-energy environment is uncertain, limiting the utility of the current knowledge. Note that the parameters, drag coefficient, \( C_D \), in Table 2.1 represents the “bulk” value over the measurement transect (vegetation patch) of a given study rather than the drag coefficient of an idealized isolated, cylinder.

The objective of the present study is to collect and analyze comprehensive field data to investigate wave attenuation over coastal marsh in a high-energy environment, such as that produced by a tropical storm. The wave data are used to quantify the rate of wave attenuation and the vegetation-induced bulk drag coefficient. Behavior of the wave height decay rate and the bulk drag coefficient is analyzed with respect to changing wave parameters and surge heights (degree of submergence). The dataset reveals the presence of bimodal spectra, consisting of low-frequency ocean swell in addition to the wind sea, providing an opportunity to examine differences in the bulk drag coefficients associated with these two wave systems.
Table 2.1: Wave and vegetation parameters and empirical relations of $C_D$ in recent studies.

<table>
<thead>
<tr>
<th>Study and Location</th>
<th>Waves (Random unless otherwise stated)</th>
<th>Vegetation</th>
<th>Transect</th>
<th>Empirical Relationship</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kobayashi et. al. (1993) based on Asano et al. (1988); Laboratory</td>
<td>Monochromatic $h=0.45$,0.52 m $H=0.056-0.194$ m $T_p=0.7-2$ s</td>
<td>Artificial kelp $h_v=25$ cm $b_v=5.2$ cm Thickness=0.003 cm $N_c=1100-1490$ /m$^2$</td>
<td>6 m (gages 2 m apart)</td>
<td>$C_D = 0.08 + \left( \frac{2200}{R_e} \right)^{2.4}$ $2,000&lt;R_e&lt;18,000$ $C_D=0.09-1.3$</td>
</tr>
<tr>
<td>Mendez et al. (1999) based on Asano et al. (1988); Laboratory</td>
<td>Same as above</td>
<td>Same as above</td>
<td>Same as above</td>
<td>Without swaying $C_D = 0.08 + \left( \frac{2200}{R_e} \right)^{2.2}$ $2,000&lt;R_e&lt;15,500$ $C_D=0.1-1.6$ With swaying $C_D = 0.40 + \left( \frac{4600}{R_e} \right)^{2.9}$ $2,300&lt;R_e&lt;20,000$ $C_D=0.3-6.9$</td>
</tr>
<tr>
<td>Möller et al. (1999); North Norfolk coast, England</td>
<td>$h=0.52-1.39$ m $H_z=0.24$ m (mean) $T_z=2.8$ s</td>
<td>Mixed Limonium sp., Aster sp., Atriplex sp., Salicornia sp., and Spartina.</td>
<td>180 m (gage spacing variable)</td>
<td>Not available</td>
</tr>
<tr>
<td>Möller and Spencer (2002); East Essex coast, England</td>
<td>$h=0.12-1.04$ m $H_z=0.05-0.30$ m (mean) $T_z=3.0$ s</td>
<td>Seasonal, mixed Salicornia sp., Suaeda sp., Puccinellia sp.</td>
<td>10-163 m (gage spacing variable)</td>
<td>Not available</td>
</tr>
<tr>
<td>Mendez and Losada (2004) based on Dubi (1995); Laboratory</td>
<td>$h=0.4-1$ m $H_{m0}=0.06-0.24$ m $T_p=1.26-4.42$ s</td>
<td>Artificial kelp $h_v=20$ cm $b_v=2.5$ cm $N_c=1200$ /m$^2$</td>
<td>9.3 m (gages 1.15 m apart)</td>
<td>$C_D = 0.47e^{-0.052K_c}$ $3 \leq K_c \leq 59$ $C_D=0.02-0.4$ Or $C_D = Q^{0.3}e^{-0.013BQ}$ $7 \leq Q \leq 172$ $Q = K_c^{0.76}$</td>
</tr>
<tr>
<td>Möller (2006); East Essex coast, England</td>
<td>$h=0.12-0.84$ m $H_{m0}=0.037-0.28$ m (mean) $T_z=1.1-3.3$ s</td>
<td>Mixed Salicornia sp., Spartina sp.</td>
<td>10 m (gages 10 m apart)</td>
<td>Not available</td>
</tr>
<tr>
<td>Bradley and Houser (2009); Northwest Florida coast, USA</td>
<td>$h=0.95-1.05$ m $H_{m0}=0.07-0.09$ m $T_p=1.4$ s</td>
<td>Thalassii testudinum $h_v=25-30$ cm $b_v=3-3.7$ mm $N_c=1100$ /m$^2$</td>
<td>43 m (gages 5-15 m apart)</td>
<td>$C_D = 0.1 + \left( \frac{925}{R_e} \right)^{3.16}$ $200&lt;R_e&lt;800$ $C_D=1.7-126.5$ $C_D = 126.45 K_c^{-2.7}$ $1&lt;K_c&lt;6$</td>
</tr>
<tr>
<td>Lövstedt and Larson (2010); Lake Krankesjön, Sweden</td>
<td>$h=0.36-1.37$ m $H_{rms}=0.01-0.06$ m $T_z=0.5-1.2$ s</td>
<td>Phragmites australis $h_v=h$ (emergent) $b_v=8.4$ mm $N_c=20-80$ /m$^2$</td>
<td>5.3-14.1 m (gages 1-2 m apart)</td>
<td>Not available</td>
</tr>
</tbody>
</table>
Table 2.1: (continued).

<table>
<thead>
<tr>
<th>Study and Location</th>
<th>Waves (Random unless otherwise stated)</th>
<th>Vegetation</th>
<th>Transect</th>
<th>Empirical Relationship</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mullarney and Henderson (2010); Puget Sound, Washington, USA</td>
<td>h=0.9 m H= not available T_p=2, 2.13 s</td>
<td>Schonoplectus americanus h_v=0.45, 0.81 m b_v=1.6, 2.7 mm 2 stems studied</td>
<td>Not applicable</td>
<td>Not available</td>
</tr>
<tr>
<td>Paul and Amos (2011); Isle of Wight, England</td>
<td>h=1 m H_m=0.13-0.18 m T_p=3.1-5.0 s (mean)</td>
<td>Zostera noltii h_v=12-16 cm b_v= Not available N_v=1980-4636 /m²</td>
<td>300 m (gages 30-95 m apart)</td>
<td>C_D = 0.06 + (\frac{153}{R_e})^{1.45} 100&lt;R_e&lt;1,000 C_D=0.13-1.9</td>
</tr>
<tr>
<td>Sánchez-González et al. (2011); Laboratory</td>
<td>h= 0.3-0.8 m H_m= 0.03-0.13 m T_p= 1.25-2.5 s</td>
<td>Artificial Posidonia sp. h_v= 10 cm b_v= 3 mm N_v=40,000 /m²</td>
<td>9 m (gages 3 m apart)</td>
<td>C_D = 6.7 K_C^{-0.73} 10&lt;K_C&lt;170 C_D=0.16-1.2</td>
</tr>
<tr>
<td>This study; South Louisiana coast; USA</td>
<td>h= 0.2-1.1 m H_m= 0.2-0.4 m T_p= 3-9 s</td>
<td>Spartina alterniflora h_v= 20 cm b_v= 8.5 mm N_v=420 /m²</td>
<td>28 m (gages 12-16 m apart)</td>
<td>C_D = 2 (\frac{1300}{R_e} + 0.18) 600&lt;R_e&lt;3,200 C_D=1.2-4.3</td>
</tr>
</tbody>
</table>

This paper is organized as follows. In Section 2, the formulations used to describe wave attenuation and spectral energy dissipation in the collected data, are presented. Section 3 describes the study site, experimental set up, vegetation properties and data processing methods. In Section 4, observations from the experiment and the analysis results are presented in three sub-sections with relevant discussion. Section 4.1 contains an overview of observed integral wave parameters and identifies unique characteristics of data pertaining to the presence of vegetation. In Section 4.2, spatial variation in wave height attenuation is quantified and the attenuation rates are examined in relation to the incident wave and hydrodynamic characteristics. Section 4.3 presents estimated bulk drag coefficients and their variation with respect to the non-dimensional parameters of wave regime. In Section 5, discussion on the validity of the rigid stem assumption, and lessons learned from the storm wave field study are presented. Finally, in Section 6, the summary and conclusions are presented.

### 2.2 Modeling Wave Transformation over Vegetation

Waves propagating through vegetation (e.g., seagrass, salt marsh, mangroves) dissipate energy by interacting with the vegetation. Assuming normally incident monochromatic waves that follow linear wave theory, the wave energy balance equation can be written as follows,
\[
\frac{\partial EC_g}{\partial x} = -S_v
\]  

(2.1)

where, \( E = (1/8)H^2 \) is the wave energy density, \( H \) is the wave height, \( C_g = nc \) is the group velocity, \( c = \sqrt{(g/k)\tanh (kh)} \) is the phase speed, \( k \) is the wave number, \( h \) is the still water depth, \( g \) is the acceleration due to gravity, and coefficient \( n \) is given by \( n = (1/2)[1 + (2kh/\sinh 2kh)] \). The cross-shore coordinate is represented by \( x \), and \( S_v \) (m\(^2\)/s) is the time averaged rate of energy dissipation due to vegetation per unit horizontal area.

This balance equation assumes a rigid bed and neglects all other source terms (local wave generation, white-capping, depth-limited breaking, and bed friction losses) relative to the losses due to vegetation induced drag. For the data analyzed in this paper, the magnitudes of these secondary source terms are estimated in Section 3.3.

2.2.1 Energy Dissipation Models

To estimate wave energy losses caused by vegetation that can be treated as rigid, the stems are represented by rigid obstructing cylindrical elements that impart drag forces on the flow (e.g., Dalrymple et al., 1984; Kobayashi et al., 1993; Mendez and Losada, 2004; Lowe et. al., 2005a; Lowe et. al., 2007; Luhar et al., 2010; Myrhaug and Holmdal, 2011; Chen and Zhao, 2012). The first such formulation was proposed by Dalrymple et al. (1984) for monochromatic waves. In this approach, small-diameter, rigid cylinders obstruct the flow, causing energy dissipation. The forces induced by the vegetation stems are expressed in a manner similar to Morison et al. (1950). For rigid stems, drag forces become dominant compared to the inertial forces due to accelerating fluid. Further, the drag forces due to pressure differences only (form drag) are considered as they are much larger than those arising from friction. The time-averaged (represented by over-bar) rate of energy dissipation per unit horizontal area can then be expressed as,

\[
S_v = \int_{-h}^{-(h-sh)} \frac{1}{2g} b_v N_v C_D u_z |u_z| u_z dz
\]  

(2.2)

where \( s \) is the ratio of vegetation height (\( h_v \)) to the still water depth (\( h \)), \( b_v \) is the stem diameter, \( N_v \) is the vegetation density, \( z \) is the vertical coordinate with origin at the still water level and pointing upwards, \( C_D \) is the bulk drag coefficient, and \( u_z \) is the horizontal water velocity at \( z \). More precisely, \( u_z \) is the fluid velocity relative to the horizontal velocity of the stem, but the motion of the vegetation is considered negligible in this analysis (see Section 3.3).

Assuming linear wave theory, integration of the equation above leads to,

\[
S_v = \frac{2}{3\pi} \frac{C_D b_v N_v}{g} \left( \frac{\sigma}{2} \right)^3 \frac{\sinh^3 ksh + 3 \sinh ksh}{3k \sinh^3 kh} H^3
\]

(2.3)
where $s$ is the ratio of vegetation height ($h_v$) to the still water depth ($h$), $b_v$ is the stem diameter, $N_v$ is the vegetation density, $C_D$ is the bulk drag coefficient, and $\sigma$ is the wave angular frequency.

The monochromatic wave expression above was extended by Mendez and Losada (2004) to random waves, assuming a uni-directional, narrow-banded incident spectrum, as follows,

$$
\langle S_v \rangle = \frac{1}{2\sqrt{\pi g}} \frac{C_D b_v N_v}{g} \left( \frac{\sigma_r}{2} \right)^3 \sinh^3 k_r s h + 3 \sinh k_r s h \frac{3 k_r \sinh^3 k_r h}{k_r^3} H_{rms}^3
$$

(2.4)

where the symbol $\langle \quad \rangle$ denotes the expected value of a random variable, subscript $r$ indicates that the parameters are representative and $H_{rms}$ is the root-mean-square wave height based on the Rayleigh probability density function. As representative parameters, Mendez and Losada (2004) used spectral peak values, while implementation of this formulation in the SWAN (Simulation of WAves in Nearshore areas) model uses spectral mean values. SWAN is a third-generation wave model that solves the wave action balance equation to describe the evolution of the wave spectrum over time, and geographical and spectral spaces (Booij et al., 2004).

In these integrated formulations of dissipation, as an approximation, it is assumed that the drag coefficient is independent of the wave height and wave period. The uncertainties resulting from this approximation are accounted for by the estimated bulk drag coefficients (Mendez and Losada, 2004).

A more generalized model for energy dissipation of random waves due to rigid vegetation was proposed by Chen and Zhao (2012). According to Chen and Zhao (2012), the expected value of the wave energy dissipation rate is given by,

$$
\langle S_v \rangle = \int S_{ds,v}(\sigma) d\sigma
$$

(2.5)

where,

$$
S_{ds,v}(\sigma) = \frac{1}{2} \frac{C_D b_v N_v}{g} \left( \frac{\sigma}{\sinh k h} \right)^2 \left( \int_{-h}^{-(h+sh)} U_{rms}(z) \cosh^2[k(h+z)] dz \right) S_e(\sigma)
$$

(2.6)

and

$$
U_{rms}(z) = \sqrt{2 \int \frac{\sigma^2 \cosh^2 k(h+z)}{\sinh^2 k h} S_e(\sigma) d\sigma}
$$

(2.7)

The spectral energy density is denoted by $S_e(\sigma)$. 

15
2.2.2 Wave Height Attenuation

In the literature, wave height attenuation has been generally quantified as the percentage reduction, in the representative wave height as the wave propagates over a vegetated field along a given length (e.g., Bradley and Hauser, 2009; Lövstedt and Larson, 2010; Möller, 2006; Quartel et al., 2007). It is expressed as,

\[ r = \frac{H_{in} - H_{out}}{H_{in}} \cdot 100 \]  

(2.8)

where \( H_{in} \) is the wave height entering the measurement transect and \( H_{out} \) is the wave height leaving the transect of length \( \Delta x \) along the direction of wave propagation.

Though calculation of the reduction rate, \( r \), offers a compact way of indicating the role and effectiveness of vegetation in wave damping, it is rather inconvenient for universal comparisons because its value depends on several parameters related to vegetation, as well as hydraulic regime. The important parameters affecting the reduction rate are the type of vegetation (grassy, reed-like, leafy, shrubs, or trees), vegetation density, and biomechanical properties (stiffness, height, and stem diameter). All these parameters have seasonal and spatial variation. Among the hydraulic parameters, the reduction rate may depend on water depth at the time of measurement and magnitude of the wave heights and wave periods. Thus, to improve the practical utility of the percentage wave height reduction rate, it should be qualified with the important parameters mentioned above.

Wave height attenuation has also been characterized as an exponential decay process (Asano et al., 1993; Cox et al., 2003; Kobayashi et al., 1993; Möller et al., 1999) expressed as,

\[ H_{out} = H_{in} e^{-k_H x} \]

(2.9)

where \( k_H \) is the decay rate and \( x \) is the distance along the direction of wave propagation from the location of the first gage (where \( H_{in} \) is measured) to the location where \( H_{out} \) is sought. Universal application of \( k_H \) suffers from the same drawbacks that apply to the reduction rate parameter, \( r \). However, as shown in this paper (Section 4.2), some of these dependencies can be quantified. Note that the reduction rate \( r \) in Eq. (2.8) is equivalent to \( k_H \) when \( k_H x \ll 1 \).

Assuming constant water depth and monochromatic waves (Dalrymple et al. (1984) Eq. (2.1) can be integrated to express wave attenuation as,

\[ H_{out} = \frac{H_{in}}{1 + \beta_1 x} \]

(2.10)
where

\[ \beta_1 = \frac{4}{9\pi} C_D b_y N_v H_{in} k \left( \frac{\sinh^3 ksh + 3\sinh ksh}{(\sinh 2kh + 2kh)\sinh kh} \right) \] (2.11)

Similarly, for random waves over constant depth, wave attenuation can be expressed as (Mendez and Losada, 2004),

\[ H_{rms, out} = \frac{H_{rms, in}}{1 + \beta_2 x} \] (2.12)

where

\[ \beta_2 = \frac{1}{3\sqrt{\pi}} C_D b_y N_v H_{rms, in} k \left( \frac{\sinh^3 ksh + 3\sinh ksh}{(\sinh 2kh + 2kh)\sinh kh} \right) \] (2.13)

When attenuation is low (\( \beta_1 x \) or \( \beta_2 x \ll 1 \)), the exponential decay rate parameter, \( k_H \), in Eq. (2.9) can be shown to be equivalent to \( \beta_1 \) or \( \beta_2 \) by using the approximation \( e^{k_H x} \approx 1 + k_H x \) such that \( H_{out}/H_{in} = 1/(1 + k_H x) \).

In the analysis presented herein, only \( k_H \) (Eq. (2.9)) is estimated and examined with respect to wave parameters. Parameter \( \beta_2 \) was not estimated because the presence of slope between our gages violates the assumptions of Eq. (2.12). However, this formulation is presented for completeness.

### 2.2.3 Determination of Bulk Drag Coefficients and Decay Rates

The drag coefficient is one of the unknown parameters in the models of wave energy dissipation caused by vegetation. For a single rigid stem in an oscillatory flow, the drag coefficient is a function of orbital velocity at a given depth, which in turn is a function of wave height and wave period. Additionally, in the case of flexible vegetation, stems can sway, reducing the relative velocity between stem and the orbital velocity. In a patch of vegetation, wakes formed by the neighboring stems can interact and affect the magnitude of the drag (Folkard, 2011; Ghisalberti and Nepf, 2002). These factors influence estimates of bulk drag coefficients determined from field measurements.

In most existing experimental studies, the bulk drag coefficients have been estimated from the measurements and then related to non-dimensional parameters such as the Reynolds number, \( R_e \), and the Keulegan-Carpenter number, \( K_C \). Table 2.1 summarizes existing empirical relationships along with relevant features of the studies. The common methods used to estimate \( C_D \) are described below as applied to an example set of data from three wave gages deployed along a straight line in the direction of wave propagation.
1. Using measured wave energy flux at two gages and inverting Eq. (2.3) to estimate $C_D$ for each burst. In our example with three gages, this gives us two estimates (one each between gage 1 and 2, and between gage 2 and 3) of $C_D$ for each burst. This is similar to the method followed by Bradley and Houser (2009) and Paul and Amos (2011).

2. Using measured integral wave heights and fitting Eq. (2.10) to the set of synoptic wave heights at all three gages using $C_D$ as the single variable. This results in a single $C_D$ value for each burst. Such an approach was utilized by Mendez and Losada (2004). Note that this method assumes horizontal bathymetry.

3. Using measured wave energy spectra and applying the formulation of Chen and Zhao (2012) (Eq. (2.6)) between consecutive gages. In our example, this results in two estimates of $C_D$ for each burst triplet. This method is adopted in the present study. Method 2 could not be used because it is only valid for wave attenuation by vegetation on a horizontal bottom.

Further, to determine a single $R_e$ or $K_C$ for each burst and therefore for each $C_D$, one can use measurements at the first (windward) gage or use the average of the measurements at all the gages considered. Depending on the distances between the deployed gages, overall length of the measurement transect, and intensity of wave energy and attenuation, different methods of analysis could produce different $C_D$ estimates and empirical relationships. Also, to determine the $R_e$ or $K_C$ for each burst, one can consider time-averaged, maximum orbital velocity at the bed, $u_b$, or at the canopy height, $u_c$. The length scale can be the stem diameter ($b_p$), stem height ($h_p$) or wave excursion length. Most existing studies have used stem diameter for the length scale. In this study, the $R_e$ or $K_C$ are based on $u_b$ at the first gage and stem diameter is used for the length scale.

### 2.3 Data and Methods

#### 2.3.1 Study Area and Experimental Setup

Wave data were collected over a two-day period (September 3-4, 2011) at a salt marsh wetland in Terrebonne Bay on the Louisiana coast of the Gulf of Mexico (Fig. 2.1), west and south of the Mississippi River during Tropical Storm Lee. Situated in the Mississippi River delta, Terrebonne Bay and the coastline extending for about 300 km east and west of Terrebonne Bay is one of the most productive and fragile marsh systems in the world. Due to natural and anthropogenic stressors/forces, between 1956 and 2006, the Louisiana coast has lost land at the rate of approximately 70 km$^2$/yr. This represents 80% of the total coastal wetland loss in the continental United States (Barras et al., 2003).

Terrebonne Bay is a shallow estuary bounded by the natural levees of Bayou Terrebonne on the east, and the Houma Navigation Canal on the west. Salt marshes line the upper portion of the bay, where vegetation communities include smooth cordgrass ($Spartina alterniflora$) and salt marsh meadow ($Spartina patens$). On the south, the bay is bordered by a series of narrow, low-lying barrier islands, the Isles Dernieres and the Timbalier Islands. The wave environment in the bay is generally comprised of locally generated seas, but offshore swell waves also propagate inwards through the gaps in the barrier island chain, or when the barrier islands are flooded by a tropical storm surge. The region has a micro-tidal environment (tidal range < 0.5 m) and depths
in the bay vary from 1 to 3 m. The southern fetch from the measurement site varies from 10 to 24 km. The region experiences annual winter cold weather fronts and surge and waves from tropical cyclones.

![Study area location. Terrebonne Bay, Louisiana.](image)

The marsh site selected for the field study is a vegetated platform wetland with a shallow bay on the windward (south) side. On the leeward (north) side the marsh extends for a distance of about 500 m, beyond which lies open water of the bay. A field topographic survey along a north-south transect shows a very low berm near the southern edge from where the marsh floor gently slopes inland with an average slope of 0.0062 within the measurement transect.

The southern marsh edge, where the incident waves first landed, has an approximate east-west alignment. The shore-normal direction has a bearing of 20° northwest to southeast. Five wave gages (pressure transducers W0 through W4) were deployed along a north-south transect nearly perpendicular to the marsh edge (Fig. 2.2a). Gage W0 was located in the open water on the up-wave (south) side of the marsh about 45 m away at a depth of about 1.4 m below the
mean sea level, to measure incoming wave energy. Gage W1 was the most southern gage on the marsh that encountered incident waves first. This gage was placed more than 16 m inwards (north) of the marsh edge to avoid the breaking zone created by waves breaking at the marsh edge. The post-cyclone survey of the site showed vegetation and surface damage within 8 to 10 m of the edge. The remaining three gages, W2, W3 and W4, were further inland (north); gage W4 being the farthest north at 43.8 m from the first marsh gage W1. For a maximum of 20° error in the alignment, the measurements would overestimate the travel distances between the gages by about 6% (1 − cos 20°) introducing error by the same amount in the estimates of drag.

All gages were self-logging pressure sensors that sampled continuously at 10 Hz over the duration of the storm. The sensors were encased in a heavy metal base to ensure stability under passing waves.

2.3.2 Vegetation Properties

The dominant vegetation at the site is *Spartina alterniflora*. This plant typically has a thick stem, with tapering flexible narrow blades (Fig. 2.2b). Vegetation properties were measured 11 days after the storm. Stem population density ($N_v$), stem height ($h_v$), total plant height ($h_{vt}$), stem diameter ($b_v$), and Young’s Modulus ($E_v$) were measured at one location each between gages W1-W2 and W2-W3. The population density is the number of stems in a one meter square area. The stem height is defined as the length between the plant base and the location of the topmost blade along the stem. The total plant height is defined as the length between the plant base and the tip of the plant with all blades aligned along the stem. The representative diameter of the plant was measured at one-fourth the stem height from the bottom. The Young’s Modulus was determined from measuring force required to bend the stem in the field from one-fourth the stem height by 45° angle and applying the Euler-Bernoulli beam theory. All above parameters were measured for 14 plants at each location. The mean and standard deviation of the measurements collected from the site between gages W1 and W2, and W2 and W3 are listed in Table 2.2.

<table>
<thead>
<tr>
<th>Plant parameters</th>
<th>Between W1 and W2</th>
<th>Between W2 and W3</th>
</tr>
</thead>
<tbody>
<tr>
<td>Population density, $N_v$ (m$^{-2}$)</td>
<td>424</td>
<td>420</td>
</tr>
<tr>
<td>Stem height, $h_v$ (m)</td>
<td>0.21±0.04</td>
<td>0.23±0.06</td>
</tr>
<tr>
<td>Total plant height, $h_{vt}$ (m)</td>
<td>0.62±0.05</td>
<td>0.63±0.11</td>
</tr>
<tr>
<td>Stem diameter, $b_v$ (m)</td>
<td>8.0±1.1</td>
<td>7.5±1.3</td>
</tr>
<tr>
<td>Young’s modulus, $E_v$ (MPa)</td>
<td>80±27</td>
<td>79±32</td>
</tr>
<tr>
<td>Second moment of inertia of stem, $I_v$ (m$^4$)</td>
<td>2.01E-10</td>
<td>1.55E-10</td>
</tr>
<tr>
<td>Flexural rigidity, $E_vI_v$ (N-m$^2$)</td>
<td>0.017±0.009</td>
<td>0.013±0.007</td>
</tr>
</tbody>
</table>
Fig. 2.2. (a) Close up aerial view of the study site showing wave gage configuration. The line W1-W3 (28 m, drawn to scale) shows transect alignment. (b) A Spartina alterniflora plant collected from the site for measurements. (c) Profile view of the experimental set up.

In the analysis presented in this paper, the vegetation is treated as rigid, based on our observations and on the measured biomechanical properties of the vegetation and integral wave parameters. To ascertain the validity of this treatment, a non-dimensional stiffness parameter, $\psi$, as defined in Mullarney and Henderson (2010), is calculated using the following expression.

$$\psi = \frac{E_v (b_v/2)^3 T_z}{\rho C_D h_v^3 u_c}$$

(2.14)
Using the mean vegetation properties and wave period, and the maximum velocity at the canopy height, $u_c$, the non-dimensional stiffness for our data is in the range of 17 to 38. If the peak wave period is used, then $\psi$ ranges between 32-50. Comparatively, the two stems (*Schoenoplectus americanus*) characterized as moderately flexible by Mullarney and Henderson (2010) to demonstrate effect of stem motion on wave energy dissipation, have non-dimensional stiffness values of 0.27 and 0.71; two orders of magnitude less than the present field measurements. Note that the non-dimensional stiffness calculations in Mullarney and Henderson (2010) were based on velocities measured at 0.25 m and 0.32 m depth above bed, while our velocity is based on the canopy height of 0.21 m. As we have shallow water waves, the vertical variation in velocity is negligible. Also, as seen in Fig. 2.2b, our *Spartina alterniflora* plants have a thick stem with several flat long flexible leaf blades. The blades have been observed to easily align with the flow under even moderate waves, offering no form drag. Streamlined vegetation has been observed to cause little dissipation (e.g., Elwany et al., 1995).

### 2.3.3 Wave Data Reduction

As a first step in processing the wave data, all measurements recorded while the water depth was less than 0.4 m were eliminated from further consideration, because the wave energy was found to be negligible at these levels (significant wave heights less than 0.04 m at W3). Thus, the study represents submerged vegetation conditions only.

In the last gage segment, between W3 and W4, the characteristic exponential energy dissipation due to vegetation was observed during only 5 bursts. Therefore, the entire dataset from gage W4 is not used in this analysis.

The wave energy spectra and the integral wave parameters were calculated using standard spectral analysis. The measured continuous pressure time series was first divided into consecutive segments or bursts of 15 minutes. For each burst, the spectral density of pressure, $S_p$, was calculated using Welch’s periodogram method (e.g., Bendat and Piersol, 2000). Each burst (9000 samples) was divided into segments containing 256 samples with 50% overlap, windowed with Hanning window, and ensemble averaged giving 70 degrees of freedom. The pressure spectra were transferred to wave energy spectra, $S_e$, using linear wave theory. Excessive amplification of noise through the transfer function was generally observed above 0.7 Hz with a distinct local spectral minimum. The amplified portion of the energy spectrum above this minimum was replaced by a $f^{-4}$ spectral tail. The final energy spectrum had a bandwidth of $\Delta f = 0.01$ Hz.

The integral wave parameters are defined in terms of spectral moments calculated as: significant wave height, $H_{mo} = 4\sqrt{m_0}$; root-mean-squared (RMS) wave height, $H_{rms} = 2\sqrt{2 m_0}$; mean wave period, $T_z = \sqrt{m_0/m_2}$; and spectral width, $\nu = \sqrt{(m_0 m_2/m_1^2 - 1)}$ where $m_0, m_1,$ and $m_2$ are the zero-th, first and second moment of the wave spectrum (0.03-0.7 Hz), respectively. The spectral energy above 0.7 Hz is generally less than 5% of the total energy, so excluding it does not significantly affect the analysis results.

The wave energy losses due to vegetation were considered dominant compared to the other source terms. To ascertain the validity of this assumption, the relative magnitude of source
terms of the local wave generation and the losses due to bottom-friction, white-capping, and depth-limited breaking were evaluated. The wave records with significant potential for these source terms were removed from further analysis as described below.

The existing formulations of wave generation are based on longer fetches than those analyzed in this study (16.5 m between W1 and W2; 11.5 m between W2 and W3). Considering the finite-depth conditions, the magnitude of wind generated wave energy within our study transect was estimated using Young and Verhagen (1996) non-dimensional formulations. This energy was less than 1% on the first day (average wind speed of 16.0 m/s) and less than 10% on the second day (average wind speed of 18.6 m/s). Wave records during which potential, local wind generated energy was greater than 7% of the total spectral energy were removed from further analysis.

Following Madsen (1994), energy dissipation of random waves due to bottom friction was computed (see also Lowe et al., 2005b). For the wave records analyzed, this dissipation was less than 7% of the measured energy dissipation.

The magnitude and frequency scale of white-capping is one of the least understood processes. For finite-depth conditions, Babanin et al. (2001) proposed breaking probability as a function of wave parameters based on the extensive Lake George (Australia) dataset. For our wave records, when the peak spectral steepness exceeded the proposed threshold of 0.055, the breaking probabilities were usually less than 0.03. The few wave records with considerable breaking probabilities (> 0.15) were removed from the analysis.

In the absence of a video documentation, it is reasonable to use $\gamma = H_m \sigma / h = 0.6$ as the limit for the depth limited breaking (Thornton and Guza, 1982). Due to the down-sloping bathymetry in our case, this limit is likely to be slightly higher (Raubenheimer et al., 1996), reducing the likelihood of breaking even more. In our reach between wave gages W2 and W3, the depth-limited breaking is not a concern because $\gamma$ is <0.3. At gage W1, there were six wave records with $\gamma > 0.5$, which are removed from the analysis. We chose 0.5 as the breaking limit for $\gamma$ to be conservative.

Based on above analyses, fifteen wave records where source terms other than vegetation-induced dissipation were deemed to be of significance, were removed from further analysis.

2.4 Observations and Results

2.4.1 Characteristics of the Measured Waves

Our study site experiences flooding only during high tide conditions. Such flooding is generally very shallow, with depths less than 10 cm. However, high winds and associated surge during Tropical Storm Lee on September 3rd and 4th of 2011, caused significant marsh flooding and provided an opportunity to examine wave transformation over vegetation. Tropical Storm Lee made landfall in south-central Louisiana (See Fig. 5.3 for the storm track). The slow moving storm (2 mph, 3.2 km/hr, with sustained winds of 35 mph, 56.3 km/hr on September 2nd) produced surge above 1 m at the study site. The greater water depths enabled higher incident waves to propagate over the marsh vegetation. Fig. 2.3 shows the integral wave parameters at the study site during the 2-day sampling period. The magnitude of incident waves to the marsh (gage
W0) was related to the water depth (Fig. 2.3a). On the first day, when the depth of water on the marsh steadily rose to about 0.9 m, significant wave heights correspondingly increased to about 0.79 m in the open water and 0.39 m on the marsh. On the second day of the storm, the water on the marsh rose to a slightly lower maximum of about 0.7 m, and the measured significant wave heights were up to 0.65 m in the open water, and 0.22 m on the marsh. The waves reduced sharply in height as they landed on the marsh edge and further reduced in height as they propagated over the vegetation. The wave attenuation is quantified and discussed further in Section 4.2. Note that the incident waves (Gage W1) are clearly depth-limited.

The two series of peak wave periods shown in Figures 2.3c and 2.3d indicate persistent presence of low-frequency swell in addition to wind sea. The two wave systems can be clearly identified in typical energy spectra observed at the four marsh gages at 6:45 AM on September 3, 2011 (Fig. 2.4). The local spectral energy minima (generally found to be around 0.17 Hz) between the peaks of the two wave systems was used to partition the energy into wind sea and swell. Most incident spectra were bimodal, with a low-frequency swell component and a mid-frequency wind sea component. In most cases, these spectral signatures were retained as waves propagated over the marsh vegetation.

To understand the characteristics of the wave environment further, the temporal evolution of some of the derived parameters are shown in Fig. 2.5. The parameters are relative wave height, \( \gamma = H_{m_0}/h \), relative depth, \( h/L \), and Ursell number, \( H_{m0}L^2/h^3 \), where \( L \) is the wave length based on the mean wave period, \( T_z \).

During the observation period, the relative wave height, \( \gamma \), was less than 0.6. The ratio, \( h/L \), was less than 0.2, indicating relatively shallow water depths during the observation period. Energy spectra became broader as waves travelled over vegetation, as indicated by increased spectral width (see Fig. 2.4 also). This tendency to broaden with propagation became stronger on the second day, when water depth was smaller. Ursell number is a measure of wave non-linearity in shallow water, with higher values indicating higher non-linearity. Waves appear to be rather non-linear at W1, but the non-linearity is quickly reduced as the waves propagated into the marsh.

### 2.4.2 Observed Wave Height Attenuation

The vegetation-induced wave energy losses along the wave gage transect result in corresponding attenuation in wave height. Fig. 2.6 shows the spatial variation of observed RMS wave height (\( H_{rms} \)) along the study transect for the entire dataset. When computing the mean, the waves are grouped by ranges of plant submergence ratio, \( s = h_p/h \) at gage W1. Using the submergence ratio from one gage for a given burst instead of one for each gage, ensures that each \( s \)-group consists of the same set of waves as they propagated along the transect.

The data show that the mean incident RMS wave heights varied from 0.09 m to 0.24 m depending on the submergence (or depth). Within about 45 m (the distance to the last gage), mean wave height was reduced to 0.02-0.09 m. The reduction in wave height was sharper in the first reach (W1-W2) compared to the subsequent reach (W2-W3).
Fig. 2.3. Wave environment at the study site during Tropical Storm Lee. (a) Water depth measured by wave gages (5-min averaged from the continuous record), (b) Spectral significant wave height, (c) Peak period of the low-frequency swell, and (d) Peak period of the wind sea portion of the spectra.
Fig. 2.4. Wave energy spectra recorded at four marsh gages on September 3, 2011 at 6:45 UTC.

Fig. 2.5. Wave environment at the study site during Tropical Storm Lee. (a) Relative wave height, (b) Relative water depth, (c) Spectral width, and (d) Ursell number.
Several researchers (Table 2.1) have analyzed wave height attenuation caused by vegetation and evaluated the effect of various parameters on the rate of attenuation. The general observations are that the wave height reduction increased with vegetation patch width and stem density, and decreased with increasing water depth, but no clear relationship to wave period has been reported. Some of these studies (e.g., Möller, 2006) have also presented relationships between wave height attenuation and parameters such as relative wave height, and have identified threshold values beyond which attenuation did not show any increase. In our study, the wave height attenuation was also calculated in terms of percentage wave height reduction rate \( r \) within a reach (/m) as defined by Eq. (2.8). The reduction rate varied from 1.5% to 4% /m, depending on the incident wave height (or water depth). The reduction rate decreased with increasing wave height. It should be noted that if the dissipation in the measurement transect is not dominated by vegetation drag (when gage is much farther downwave outside the realm of exponential decay), then the attenuation rate could be skewed by the distance over which it is calculated. The attenuation expressed by Eq. (2.8) implies linear wave height decay, while in reality attenuation is closer to being exponential as stated before. Consequently, reduction rate values may vary depending on the spatial location where they are calculated. This may partly explain the wide ranges of percentage wave height attenuation rates reported in the literature, such as 0.34% /m (Möller et al., 1999), 0.77% /m (Bradley and Houser, 2009) and 4.0-5.0% /m (Lövstedt and Larson, 2010).

To examine the exponential nature of wave height decay, the decay rates \( k_H \) were obtained by fitting Eq. (2.9) to the data. For each burst, Eq. (2.9) was fitted to the set of RMS wave heights from gages W1, W2, and W3 to obtain a single \( k_H \) value. This process was repeated for all bursts, resulting in a set of \( k_H \) values that ranged from 0.022 to 0.051 /m, with...
excellent fit for each burst (lowest $R^2=0.98$). The rates are the same order of magnitude as those from some previous studies. For example, for random wave dissipation over artificial kelp in a laboratory flume, Dubi and Tørum (1996) estimated the decay parameters up to 0.011 /m when the depth was 1 m. The range of exponential decay rates reported by Kobayashi et al. (1993) is from 0.015 to 0.101 /m for monochromatic waves over artificial vegetation in a laboratory flume. The decay rates observed by Bradley and Houser (2009) for three wave records were 0.007, 0.015 and 0.008 /m for random waves in the field with incident wave heights of 7 to 10 cm.

To understand the nature of this variation further, the $k_H$ values were plotted against incident wave height, Reynolds number, $R_e (= \rho b_v u_b/\mu)$, and Keulegan-Carpenter number, $K_C (= u_b T_z/b_v)$ (Fig. 2.7). Here $u_b (= H_{rms} \omega_z/2 \sinh k_z h)$ is the maximum near-bed orbital velocity in the absence of vegetation given by the linear wave theory, and subscript $z$ indicates the mean value of the parameter, using the measurements at the first (windward) of the two bounding gages. Fig. 2.7 shows that larger waves decayed at a slower rate than smaller waves. The decay rates for the low-frequency swell were lower than those for the high-frequency wind sea regardless of wave height. This is consistent with observations of Lowe et al. (2007), who found that, under random wave conditions, high-frequency (>0.2 Hz) waves penetrated more effectively into a model canopy and were dissipated at a greater rate.

2.4.3 Bulk Drag Coefficient

The wave conditions during our study consisted of random waves with broad spectra. This precluded application of the Dalrymple et al. (1984) model, which was developed for monochromatic waves. Therefore, the random wave models by Mendez and Losada (2004) and Chen and Zhao (2012) were used. To apply the Mendez and Losada (2004) model, the left hand side of Eq. (2.4) was calculated by dividing the difference of measured wave energy flux between adjacent pairs of wave gages (i.e., W1-W2 and W2-W3) by the distance between them. The right hand side was calculated using the average values of integral wave parameters at the same two gages. In applying Chen and Zhao (2012) formulation, the left hand side of Eq. (2.6) was calculated as stated above and the right hand side was calculated by numerical integration along the vertical stem height and along spectral frequency.

Fig. 2.8 shows the estimated values of bulk drag coefficients, $C_D$, using these two methods, plotted against the Reynolds number, $R_e$. The $R_e$ is calculated based on the velocity at the first of the two bounding gages. The $C_D$ estimates by the two methods are very similar. Note that the relatively low $R_e$ values are due to the small stem diameter (8.0 mm) used as the characteristic length. To derive the empirical relationship between $C_D$ and $R_e$, the form suggested by Tanino and Nepf (2008) is fitted ($R^2 = 0.93$) to yield,

$$C_D = 2 \left( \frac{1300}{R_e} + 0.18 \right) \quad 600 < R_e < 3,200 \quad (2.15)$$

The solid volume fraction for our data is $\phi = N_s \pi b_v^2 / 4 = 0.021$. Considering the range of $R_e$ and $\phi$, the coefficients in Eq. (2.15) are of the same order as those listed in Tanino and Nepf (2008, Table 2 for Petryk 1969). The $R_e$ in that table is based on pore velocity, but is the same order of magnitude as the $R_e$ in this study.
Fig. 2.7. Variation of exponential wave height decay rate with RMS wave height, Reynolds number, and Keulegan-Carpenter number. Independent variables are based on measurements at gage W1.

Fig. 2.8. Variation of bulk drag coefficient estimated by two models, with Reynolds number. The Reynolds number is calculated using measurements from the windward gage of each pair and stem diameter.
In Fig. 2.8, the estimated $C_D$ ranges from 1.2 to 4.3. It decreases as $R_e$ increases. The $C_D$ values are larger than those for an isolated cylinder, specifically for $R_e < 1500$. This is consistent with the findings of Koch and Ladd (1977) and Tanino and Nepf (2008). Table 2.1 also shows that, in several previous studies, $C_D$ values much larger than 1 have been reported. In Fig. 2.8, smaller $C_D$ values (<2.0) were found in the reach between gages W1 and W2, where $R_e > 1500$.

Variation of $C_D$ with the Keulegan-Carpenter number, $K_C$, is shown in Fig. 2.9 with the following regression equation,

$$C_D = 70.0 \, (K_C)^{-0.86} \quad 25 < K_C < 135$$

(2.16)

![Fig. 2.9. Variation of estimated drag coefficient, with Keulegan-Carpenter number. The Keulegan-Carpenter number is calculated using measurements from the windward gage of each pair and stem diameter.](image)

As stated before, the recorded wave spectra showed the presence of low-frequency swell in addition to wind sea. The frequency distributed form of the Chen and Zhao (2012) model (Eq. (2.6)) allows calculation of a band-averaged $C_D$. Fig. 2.10 shows average $C_D$ values calculated for the swell (0.03-0.17 Hz) and the wind sea (0.17-0.7 Hz) bands, plotted against the $K_C$ number for the entire spectrum. Each $K_C$ value represents one spectrum. The estimated $C_D$ for longer-period waves is generally smaller than that for the wind sea at each spectrum ($K_C$ number). This is consistent with the theoretical analysis proposed by Lowe et al. (2007) for orbital velocity attenuation within a canopy. Our bulk $C_D$ scales as $C_d \alpha^3$, where $C_d$ is the empirical drag coefficient (assumed to be 2.5 in Lowe et al., 2007) and $\alpha$ is the ratio of the orbital velocity inside the canopy to that outside the canopy. As demonstrated by Lowe et al. (2005b, 2007), the orbital velocity of the longer-period waves inside the canopy is considerably less (smaller $\alpha$) than that of the shorter-period waves. This results in smaller $C_D$ values for the longer-period...
waves (swell) compared to the shorter-period wind sea component of a spectrum. For practical applications of spectral wave modeling, determining yet another $C_D - K_C$ regression equation may be of limited value, while a more detailed investigation into frequency-dependence of $C_D$ is warranted. This is a subject of a separate research paper.

### 2.5 Discussion

The primary objective of this paper is to present the unique data and analysis of wave attenuation by vegetation under storm-induced high-energy waves, and review the results in relation to the few past studies, which were mostly conducted under low-energy wave conditions. Based on the published data, a direct comparison of the results, especially the $C_D - K_C$ regression curves, from existing studies (Table 2.1) is not possible. This is due to the different approaches followed in each study. For example, Mendez et al. (1999) estimated $C_D$ as a single calibration parameter fitted to several gages along a transect, while Bradley and Houser (2009) and Paul and Amos (2011) estimated $C_D$ between pairs of gages for each burst. Also, some studies have calculated $R_e$ or $K_C$ based on the maximum orbital velocity at the canopy height, while the others have used near-bed velocity as a reference. Further, this velocity can be either apparent or pore velocity. Some studies have used the average $R_e$ or $K_C$ along the gage transect, while others have used the value at the first gage. The representative wave period has been selected to be either the peak period or mean period. Among the relationships listed, Mendez et al. (1999) and Kobayashi et al. (1993) were developed under laboratory monochromatic waves, while Bradley and Houser (2009) and Paul and Amos (2011) were developed from field studies of random waves. Moreover, the type of vegetation differs among studies. Nevertheless, the various relationships do show the nature and strength of the dependency on the selected variables.

The results from our study could be extended to reed-like vegetation under similar wave conditions and submergence. In the case of vegetation where above-ground stem is absent or is characterized by significant foliage, our results cannot be applied. The presence of foliage contributes to the drag provided by the plant, especially at lower velocities when it is not streamlined or compressed (Wilson et al., 2008). The vegetation submergence is also important, as the mechanism of turbulence exchange changes from longitudinal to vertical with increasing submergence (Nepf and Vivoni, 2000). Another important factor is the plant flexibility. While empirical results from rigid, reed-like vegetation studies can be used for similar vegetation, for flexible vegetation, models that capture vegetation motion (e.g., Mullarney and Henderson, 2010) must be implemented.

Field data collection of attenuation of storm-induced high-energy waves by salt marsh vegetation poses several challenges which partly explain the lack of such data prior to this study. The first of these challenges is finding a site where a healthy stand of vegetation exists that has a reasonable chance of occurrence of sufficient water depth and waves. This is difficult because, along the coastal locations where high-energy waves are routinely present, salt marsh vegetation does not survive. Therefore, one has to look for a site that has established vegetation, has a fair chance of inundation, and has a favorable fetch to produce high-energy wave conditions when high winds occur. Due to the changing forecast of the storm-track, several candidate salt marsh sites needed to be considered. The ultimate deployment site can usually be decided only a couple
of days prior to cyclone landfall. Our 2011 Tropical Storm Lee field experiment was successful only after failed attempts in 2009 (Tropical Storm Ida) and 2010 (Tropical Storm Bonnie) when our study sites fell on the left side of the storm-track and did not experience either surge or waves along our wave gage transect. Due to the anti-clockwise wind field of the tropical storms in the northern hemisphere, the coastal water is pushed out of the wetlands on the left side of the storm.

Second, the sustained high winds (>20 m/s) and wave forces associated with the storms make it difficult to deploy any upright instrumentation such as video cameras, wave staffs, meteorological stations or acoustic Doppler velocity profilers on wetlands. Therefore, one is limited to the use of bottom-mounted pressure transducers with a short window for rapid deployment prior to cyclone landfall. The presence of high winds also makes it necessary to ascertain that the source terms of wave generation and white-capping (even if the fetch between the gages is short) are negligible compared to the wave energy loses due to vegetation. This is less of a concern in low-energy wave-vegetation studies.

Finally, the relatively rapidly changing (compared to tidally varying) storm-induced hydrodynamic environment results in simultaneous changes in surge, wave heights, and wave periods. The opportunity of controlling one variable to examine others is thus unavailable under storm conditions. Comparatively, when wave attenuation through vegetation is studied during high tide inundation, the depths are relatively stable while wave parameters could be changing.

Though the findings of this study are applicable to any coastal area with Spartina-type salt marsh vegetation, it provides critical wave attenuation and drag information applicable to the vast marshes of the Northern Gulf of Mexico. Due to the catastrophic land-loss and the ongoing navigational, and oil and gas industry impacts of national importance, protection of coastal

Fig. 2.10. Estimated drag coefficients for the long-period (swell) and the short-period (wind sea) waves of measured spectra. Each spectrum is represented by a single Keulegan-Carpenter number.
wetlands in this region to reduce storm damages has become critically important, warranting science based solutions (e.g., Day et al., 2007; CPRA, 2012). Numerical models of wave and surge employed in the design and protection of coastal infrastructure and for resource management will benefit from the improved understanding of vegetation-induced wave attenuation in these wetland systems.

### 2.6 Summary and Conclusions

The phenomenon of wave energy dissipation by salt marsh has been investigated in several laboratory studies but relatively few field studies. Further, the existing field studies were carried out in a low-energy wave environment, limiting their applicability to the high-energy wave field, such as that produced during a tropical cyclone. The present study fills this gap in the knowledge of this important phenomenon by providing analysis of cyclone generated waves over salt marsh consisting of Spartina alterniflora. This is the first comprehensive field dataset acquired and analyzed over salt marsh vegetation under tropical storm wave conditions.

The magnitude of the wave height reduction rate, \( r \), commonly expressed by Eq. (2.8) was found to vary considerably (1.5 to 4% /m) depending on the magnitude of incident wave height. Since the incident wave height was different for each pair of gages, \( r \) was spatially variable, making it unreliable as a general indicator of the effectiveness of vegetation in wave damping. Also, \( r \) values based on observations from different vegetation types or wave conditions cannot be compared.

Consistent with the previous studies, the storm waves were observed to attenuate exponentially over vegetation (decay rates of 0.022-0.051 /m). The larger waves attenuated at a smaller rate than the smaller waves. The wave height attenuation rate was also observed to be dependent on the magnitude of the dominant frequency of the wave systems. The low-frequency waves (swell) attenuated at a lower rate than the high-frequency wind sea waves. This is consistent with Lowe et al. (2007), who observed efficient attenuation of higher frequency random waves by a rigid model canopy array.

The bulk drag coefficient \( C_D \) (1.2-4.3) was estimated along the study transect using two formulations, namely Mendez and Losada (2004) and Chen and Zhao (2012). This coefficient does not represent the drag coefficient of an isolated rigid cylinder but rather a bulk drag coefficient that is temporally and spatially averaged over the vegetation patch. It accounts for uncertainties associated with processes that are not explicitly defined in the equations, such as wake interference due to other vegetation, frictional losses due to vegetation blades in addition to stems, and, most importantly, reduced velocity inside a canopy. Consistent with the previous studies, \( C_D \) was observed to decrease with increasing \( R_e \) and \( K_C \) numbers. The coefficients of the empirical relationship between \( C_D \) and \( R_e \) developed in this study are consistent with those reported in the literature. For Spartina type vegetation, this relationship can be applied within the \( R_e \) range of 600-3,200. The Spartina spp. is found along the margins of the most continents in the temperate zone (e.g., Chapman, 1960). Comparison of published empirical relationships between \( C_D \) and non-dimensional numbers, such as of \( R_e \) and \( K_C \) requires caution, because the methods employed in various studies to estimate \( C_D \) differ, as do the definitions of \( R_e \) and \( K_C \).
For a given wave spectrum, the $C_D$ was observed to be smaller for the longer-period waves than for the shorter-period waves. The data presented in this Chapter have been analyzed to quantify variation of the bulk drag coefficient across frequency scales. The results of these analyses are presented in the next chapter.

### 2.7 References


CPRA (2012), Louisiana’s Comprehensive Master Plan for a Sustainable Coast. Coastal Protection and Restoration Authority of Louisiana. Baton Rouge, LA.


CHAPTER 3: SPECTRAL DISTRIBUTION OF WAVE ENERGY DISSIPATION BY SALT MARSH VEGETATION

3.1 Introduction

Wave propagation through vegetation is an important physical process along many coastal regions of the world, and along the shores of large inland lakes. Waves approaching vegetated shores lose energy due to obstructing vegetation. This reduces shoreline erosion and is of engineering significance for shoreline protection. The role and importance of coastal wetlands as a natural defense system against storm waves is generally acknowledged (e.g., Costanza et al., 2008; Dixon et al., 1998; Gedan et al., 2011; Lopez, 2009). Utilization of coastal wetlands to augment structural measures for mitigation of coastal flooding due to storm surge and waves is promoted in several regions of the world (e.g., Borsje et al., 2011; CPRA, 2012).

A body of literature exists quantifying reduction rates of integral wave heights due to vegetation (for summary, see Anderson et al., 2011; Jadhav and Chen, 2012). Theoretical models based on energy conservation, have been proposed for application to both monochromatic waves (Dalrymple et al., 1984), and for narrow-banded random waves (Mendez and Losada, 2004). Kobayashi et al. (1993) presented an approach based on continuity and momentum equations, that assumed exponential decay of integral wave height. Chen and Zhao (2012) proposed a vegetation-induced dissipation model based on the formulation of Hasselmann and Collins (1968) for energy dissipation of random waves by bottom friction. All these models assume rigid vegetation. A number of recent studies have underscored the importance of accounting for the stem and blade motion of flexible vegetation, and have proposed models that account for it (Bradley and Houser, 2009; Mullarney and Henderson, 2010; Riffe et al., 2011). Wave attenuation has been studied in a controlled laboratory environment (Augustin et al., 2009; Dubi and Tørum, 1996; Løvås and Tørum, 2001; Stratigaki et al., 2011), in field conditions involving salt marshes (Bradley and Houser, 2009; Cooper, 2005; Jadhav and Chen, 2012; Möller et al., 1999; Möller and Spencer, 2002; Möller, 2006; Riffe et al., 2011), coastal mangrove forests (Mazda et al., 2006; Quartel et al., 2007), and vegetated lakeshores (Lövstedt and Larson, 2010). Most of these studies primarily focused on the attenuation of integral wave heights or wave energy, and estimation of integral bulk vegetation drag coefficients. As a step beyond integral dissipation characteristics, Lowe et al. (2005) developed an analytical model to predict the magnitude of the in-canopy velocity of waves propagating over a model canopy made up of rigid cylinders. Lowe et al., (2007) extended this model to random waves and predicted that high frequency components of wave energy would dissipate more efficiently inside the canopy. The model was verified with measurements taken from an artificial rigid cylinder canopy submerged on a barrier reef (random wave conditions) for 2 hours and assuming a constant drag coefficient.

In the case of natural vegetation under random waves generated by a tropical cyclone, there are no published studies that examine in detail the frequency-based characteristics of wave energy dissipation and drag coefficient, though some studies have illustrated such characteristics with an example (Bradley and Houser, 2009; Paul and Amos, 2011). The present study investigates the spectral characteristics of wave energy dissipation due to natural vegetation, and
the relationship between dissipation and the incident wave energy spectrum, using comprehensive field data. The study also identifies spectral variation of the vegetation drag coefficient. We hypothesize that the frequency-varying spectral drag coefficient will predict spectral distribution of energy dissipation more accurately than an integral drag coefficient. To test the hypothesis, a new method is developed to parameterize the spectral drag coefficient over the entire range of measured wave spectra. The spectral and integral drag coefficients are then both used to estimate energy dissipation losses, and these estimates are compared to the observed dissipation to assess the validity of the hypothesis.

The following section describes the spectral energy dissipation model proposed by Chen and Zhao (2012) which is used to estimate drag coefficients and introduces the velocity attenuation factor. Sections 3 and 4 describe the field program and the wave conditions. Section 5 contains data analysis, where spectral characteristics of the observed energy dissipation are examined. In Section 6, spectral variation of estimated drag coefficient is demonstrated, and the spectral behavior of the mean velocity attenuation parameter is quantified. The mean velocity attenuation parameter and average drag coefficients are then applied to predict energy dissipation and compared with the existing prediction methods in Section 7. Finally the results are discussed, followed by a summary and conclusions.

### 3.2 Spectral Energy Dissipation Model

Assuming the linear wave theory holds, the evolution of random waves propagating through vegetation can be expressed with the following wave energy balance equation,

$$\frac{\Delta (E_j C_{g,j})}{\Delta x} = -S_{ds,j}$$  \hspace{1cm} (3.1)

where subscript \( j \) represents the \( j \)th frequency component of a wave spectrum, \( E \) is the spectral wave energy density, \( C_g = nc \) is the group velocity, \( c = \sqrt{(g/k) \tanh (kh)} \) is the phase speed, \( k \) is the wave number, \( h \) is the still water depth, \( g \) is the acceleration due to gravity and coefficient \( n \) is given by \( n = (1/2)[1 + (2kh/\sinh 2kh)] \). The cross-shore coordinate is given by \( x \) pointing landward and \( S_{ds} \) is the energy dissipation due to vegetation per unit horizontal area. All other source terms are considered negligible compared to the vegetation induced losses.

The spectral wave energy dissipation due to vegetation is obtained by using a reorganized form of the model proposed by Chen and Zhao (2012). Their model treats vegetation as rigid, cylindrical elements that impart drag forces on the flow. Further, only the drag forces due to pressure differences are considered, as they are much larger than those arising from friction in the hydraulic regimes encountered in the field conditions.

In this model, the spectral energy dissipation due to vegetation is expressed by,

$$S_{ds,j} = \frac{1}{2} \frac{C_{v,j} b_v N_e}{g} \left( \frac{\sigma_j}{\sinh k_j h} \right)^2 \left( \sum_{z=-h}^{-h+sh} U_{z,rms}(z) \cosh[k_j(h+z)] \Delta z \right) E_j$$  \hspace{1cm} (3.2)
where $C_{D,j}$ is a bulk drag coefficient, $b_v$ is the stem diameter, $N_v$ is the vegetation population density, $\sigma_j$ is the wave angular frequency, $s$ is the ratio of vegetation height, $h_v$, to the still water depth, $h$, and $U_{rms}$ is the root-mean-squared (RMS) velocity given by,

$$U_{z,rms} = \sqrt{\frac{2}{\sum_{j=1}^{N_f} \frac{\sigma_j^2 \cosh^2 k_j(h + z)}{\sinh^2 k_j h} E_j \Delta \sigma}} \quad (3.3)$$

where $N_f$ is the total number of frequency components of a spectrum.

Eq. (3.2) is based on the quadratic representation of the shear stress induced by the vegetation. We parameterize the shear stress due to vegetation drag at elevation $z$ (positive upwards with origin at the still water level) due to $j^{th}$ component of the spectrum as,

$$\tau_{z,j} = -\frac{1}{2} \rho b_v N_v C_d \alpha_j u_{z,j} |\alpha_j u_{z,j}| \Delta z \quad (3.4)$$

where $\rho$ is the density of water, $\alpha_j u_{z,j}$ is the vegetation-affected velocity at elevation $z$, and $C_d$ is the drag coefficient corresponding to this velocity. The velocity attenuation parameter, $\alpha$, is defined as the ratio of the vegetation-affected velocity, $u'_z$, to the velocity in the absence of vegetation, $u_z$, at elevation $z$ inside the canopy:

$$\alpha_{z,j} = \frac{u'_{z,j}}{u_{z,j}} \quad (3.5)$$

This parameter is similar to Lowe et al. (2005) but not exactly the same. In Lowe et al. (2005), a similar parameter is defined as the ratio of the velocity within canopy to that above canopy. These two definitions of the velocity attenuation parameter are related by depth factor resulting from the depth-dependent decay of orbital velocity.

Similar to the definition of $\alpha$ (Eq. (3.5)), the ratio of the vegetation-affected RMS velocity at an elevation $z$, $U'_{z,rms}$, to the RMS velocity in the absence of vegetation, $U_{z,rms}$, at elevation $z$ inside the canopy is defined as,

$$\alpha_{z,r} = \frac{U'_{z,rms}}{U_{z,rms}} \quad (3.6)$$

Using these definitions, Chen and Zhao (2012) formulation is reorganized and the spectral distribution of energy dissipation is expressed as (See Appendix for details),

$$S_{ds,j} = \frac{1}{2} \bar{C}_d b_v N_v \alpha_{z,j}^2 \left( \frac{\sigma_j}{\sinh k_j h} \right)^2 \left( \sum_{-h}^{-h+sh} U_{z,rms} \cosh^2 [k_j(h + z)] \Delta z \right) E_j \quad (3.7)$$
where $\bar{C}_D$ is the spectrally-averaged, or integral, drag coefficient. To facilitate solution of Eq. (3.7), $\alpha$ is assumed to be independent of depth, and a normalized form of $\alpha$ is introduced as,

$$\alpha_{n,j} = \frac{\alpha_j}{\alpha_r} \quad (3.8)$$

Note that while $\alpha_j$ is always less than 1, $\alpha_{n,j}$ can be greater than 1. Using $\alpha_{n,j}$, Eq. (3.7) can then be re-written as,

$$S_{ds,j} = \frac{1}{2} \frac{\bar{C}_D b v N_v}{g} \alpha_{n,j} \left( \frac{\sigma_j}{\sinh k_j h} \right)^2 \left( \sum_{n = -h}^{h} U_{z, rms} \cosh[k_j(h + z)] \Delta z \right) E_j \quad (3.9)$$

The spectrally variable drag coefficient is then expressed as,

$$C_{D,j} = \bar{C}_D \cdot \alpha_{n,j}^2 \quad (3.10)$$

Integrated over the entire spectrum, the time-averaged rate of energy dissipation per unit area is given by,

$$S_v = \sum_{j=1}^{j=N_f} S_{ds,j} \Delta \sigma \quad (3.11)$$

### 3.3 Study Area and Field Program

The study site was a salt marsh wetland in Terrebonne Bay on the Louisiana coast of the Gulf of Mexico (Fig. 3.1) west of the Mississippi River bird-foot delta. The shallow (depth, 1-3 m), micro-tidal (diurnal tidal range < 0.5 m) bay is bordered by salt marsh to the north, and a series of narrow, low-lying barrier islands to the south. The waves in the bay consist of frequent low-energy offshore swell and locally generated seas which intensify during the passages of annual winter cold fronts and tropical cyclones.

During Tropical Storm Lee (September 3-4, 2011), three wave gages (pressure transducers W1 through W3) were deployed on a vegetated platform marsh along a north-south transect (28 m long) approximately perpendicular to the salt marsh edge (Fig. 3.1). The shore-normal has a bearing of 20° northwest to southeast. For a maximum of 20° error in the alignment, the measurements would overestimate the travel distances between the gages by about 6% ($1 - \cos 20^\circ$) introducing corresponding error in estimates of energy dissipation. Waves approached from the south and propagated from Gage W1 to W3 through vegetation. Gage W1 was located more than 16 m inwards of the marsh edge to avoid the waves breaking at the marsh edge. The self-logging pressure sensors sampled continuously at 10 Hz over the 2-day duration of the storm.
The dominant vegetation at the site is *Spartina alterniflora*, having a thick stem and thin, tapering flexible narrow blades. The average measured vegetation properties were: \( N_v = 422 \) stems/m\(^2\), \( h_v = 0.22 \) m (stem height), \( h_{vt} = 0.63 \) m (total plant height), \( b_v = 8.0 \) mm, and \( E_v = 80 \) MPa \( (E_v I_v = 0.015 \) N-m\(^2\)) where \( E_v I_v \) is the flexural rigidity and \( I_v \) is the second moment of inertia of a stem. Based on our observations and the estimated non-dimensional stiffness parameter (Mullarney and Henderson, 2010), the vegetation can be treated as rigid (see analysis in Jadhav and Chen, 2012).

The time series of continuous pressure measurement from wave gages were analyzed using standard spectral techniques (e.g., Bendat and Piersol, 2000). Each burst (9000 samples) was divided into segments containing 256 samples with 50% overlap, windowed with Hanning window, and ensemble averaged giving 70 degrees of freedom. The resulting energy spectra had bandwidth, \( \Delta f \), of 0.01 Hz, with 95% of the spectral energy between 0.03 and 0.7 Hz. Thus each spectrum had 69 frequency components \( (N_f \text{ in Eq. (3.3) and Eq. (3.11))}. \) The integral wave parameters are defined as: significant wave height, \( H_{mo} = 4\sqrt{m_0} \); mean wave period, \( T_z = \sqrt{m_0/m_2} \); and spectral width, \( \nu = (m_0 m_2/m_1^2 - 1) \) where \( m_0, m_1, \) and \( m_2 \) are the zero-th, first and second moment of the wave spectrum, respectively.

The wave energy loss due to vegetation was considered dominant compared to the other source terms. To ascertain the validity of this assumption, the relative magnitude of source terms for the local wave generation and the losses due to bottom-friction, white-capping, and depth-limited breaking were evaluated. The wave records with significant potential for the magnitude of these source terms to be dominant, were removed from further analysis (for details see Jadhav and Chen, 2012).


### 3.4 Overview of Wave Conditions

A total of 177 wave records (59 records at each of the 3 gages) were analyzed in this study. Table 3.1 lists summary statistics of water depth, zero-moment wave height, mean period and some derived parameters characterizing the wave conditions. The statistics in Table 3.1 describe only the analyzed data, not the entire measured data set. As stated in the previous section, the wave records that violated assumptions of Eq. (3.1) were removed from analysis. With the diurnal tide augmented by the storm surge, the water depth rose from about 0.1 m to 0.8 m and then fell along with the tide. Only the measurements collected when water depth was greater than 0.4 m were used in the analyses, because wave energy levels were insignificant when water depth was less than 0.4 m.

The incident significant wave heights \( H_{m0} \) on the marsh varied from 0.05 to 0.39 m and were directly proportional to the depth of flood water. The recorded wave spectra were largely bimodal (Fig. 3.2) with distinct low-frequency swell (7-10 s) and wind sea components (2-4.5 s).

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Gage W1</th>
<th>Gage W2</th>
<th>Gage W3</th>
</tr>
</thead>
<tbody>
<tr>
<td>Depth, ( h ) (m)</td>
<td>0.40-0.82 (0.55)</td>
<td>0.57-1.0 (0.72)</td>
<td>0.57-1.01 (0.72)</td>
</tr>
<tr>
<td>Significant Wave Height, ( H_{m0} ) (m)</td>
<td>0.15-0.40 (0.24)</td>
<td>0.07-0.28 (0.14)</td>
<td>0.04-0.21 (0.09)</td>
</tr>
<tr>
<td>Peak Wave Period, ( T_p ) (s)</td>
<td>2.5-4.7 (4.0)</td>
<td>1.2-4.5 (2.3)</td>
<td>1.3-4.5 (2.6)</td>
</tr>
<tr>
<td>Relative Wave Height, ( H_{m0}/h )</td>
<td>0.36-0.49 (0.41)</td>
<td>0.12-0.29 (0.18)</td>
<td>0.08-0.22 (0.12)</td>
</tr>
<tr>
<td>Relative Depth, ( h/L_z )</td>
<td>0.07-0.13 (0.10)</td>
<td>0.09-0.16 (0.13)</td>
<td>0.10-0.16 (0.12)</td>
</tr>
<tr>
<td>Spectral Width, ( \nu )</td>
<td>0.45-0.58 (0.51)</td>
<td>0.44-0.64 (0.5)</td>
<td>0.43-0.65 (0.53)</td>
</tr>
<tr>
<td>Ursell Number, ( H_{m0}L_z^2/h^3 )</td>
<td>29-81 (48)</td>
<td>9-16 (11)</td>
<td>6-10 (8)</td>
</tr>
</tbody>
</table>

Fig. 3.2. Wave energy spectra recorded on September 3, 2011 at (a) 6:45 UTC and (b) 12:30 UTC.
3.5 Observed Spectral Wave Energy Dissipation Characteristics

Measured spectra showed significant wave energy reduction over vegetation, as evidenced by the reduction in wave heights (Table 3.1). Energy reduction with respect to frequency was calculated between pairs of wave gages (W1-W2 and W2-W3) based on the measured wave energy density spectra, using Eq. (3.1). Ensemble averages of all analyzed energy density spectra, along with the ensemble average of the energy dissipation are shown in Fig. 3.3 for reaches W1-W2 (between gages W1 and W2) and W2-W3 (between gages W2 and W3). The energy density and dissipation are normalized by, $m_0$, the zero-th moment of the individual spectrum measured at the windward gage of the pair of gages bounding the reach. Fig. 3.3 shows that the magnitude of energy dissipation varies with the frequency. Higher dissipation was observed at the frequencies adjacent to the spectral peak in both reaches. Most of the wind sea energy dissipated in the leading vegetation reach, W1-W2. Significant portions of swell energy propagated beyond the leading reach and dissipated in reach W2-W3.

![Fig. 3.3. Ensemble average of all normalized energy density and energy dissipation spectra in (a) reach W1-W2 and (b) reach W2-W3. Spectra normalized by the zero-th moment ($m_0$) of the energy spectrum measured at the windward gage of the pair of gages.](image)

Fig. 3.3 also shows that the dominant loss near the spectral peak is less pronounced in the second reach, W2-W3, where a substantial portion of the total energy loss occurs at frequencies higher than the peak. This is illustrated in Fig. 3.4 where energy reduction in the dominant wave frequencies, i.e., swell and wind sea band (0.03-0.36 Hz), as a percentage of the total (0.03-0.7 Hz) energy reduction is plotted as a function of Keulegan-Carpenter number, $K_C$. The $K_C$ number is defined as, $K_C = U_{rms}T_d/b_v$, where $U_{rms}$ is the root-mean-square orbital velocity at the bed,
considering the entire spectrum. In reach W1-W2, wave energy reduction in the swell and wind sea bands accounted for 55 to 70% of the total reduction, while in reach W2-W3, this percentage was only 40 to 55%. Thus, in reach W2-W3 the energy reduction was more evenly distributed between dominant and higher frequencies. This is partly due to modification of the spectral shape as a result of the non-linear transfer of energy to the higher frequencies as waves propagated from gage W1 to W2.

![Graph showing wave energy reduction](image)

Fig. 3.4. Wave energy reduction in the swell and wind sea band (0.03-0.36 Hz) as a percentage of the total (0.03-0.7 Hz) energy reduction.

Across the frequencies above the peak, the spectral distribution of energy dissipation was observed to gradually taper off. The rate of such tapering with respect to spectral frequency is shown in Fig. 3.5 using normalized dissipation ($S_{ds}(f)/E(f)$) for 3 ranges of $K_C$ numbers. The choice of the range of $K_C$ for ensemble averaging is inconsequential and is made for the purpose of creating three ranges of $K_C$ signifying ranges of hydrodynamic conditions. Variation of the frequency exponent over all spectra with respect to $K_C$ number is shown in Fig. 3.6. Larger $K_C$ numbers generally represent waves in reach W1-W2. Waves in this reach were more energetic, with more peaked spectra and larger concentration of energy in the swell-sea band (0.03-0.36 Hz). The smaller values of $K_C$ numbers represent relatively low energy waves with much broader spectra. Fig. 3.5 shows that at frequencies above the peak, and at higher $K_C$ numbers, the normalized energy dissipation has a stronger dependence on frequency.

The current standard modeling practice assumes that the distribution of energy dissipation generally follows the incident wave energy density spectrum (e.g., Suzuki et al., 2011). To assess the validity of this assumption, the following hypothesis was tested using our field study measurements:

$$S_{ds}(f) = a \cdot E(f)^b$$  \hfill (3.12)
where \(a\) and \(b\) are determined by regression analysis. For a given reach (W1-W2 or W2-W3), each incident energy spectrum, \(E(f)\), and the corresponding dissipation spectrum, \(S_{ds}(f)\), were divided into three frequency bands, representing swell (0.03-0.16 Hz), wind sea (0.16-0.32 Hz) and high frequency (0.32-0.7 Hz). These divisions correspond to the local spectral energy minima observed around 0.16 Hz and 0.32 Hz in the recorded bimodal spectra (Fig. 3.2). For each of these three frequency bands, a coefficient pair \((a, b)\) was determined by fitting Eq. (3.12) to the field data. Thus, for each spectrum (wave record), three coefficient pairs were obtained. Coefficient pairs where the fit of Eq. (3.12) to the field data resulted in an \(R^2\) (coefficient of determination) less than 0.8, were excluded from the analysis.

![Fig. 3.5](image1.png)

Fig. 3.5. Frequency distribution of the the ensemble-averaged normalized energy dissipation rate. Curves represent ensemble averages of all measured spectra in reaches W1-W2 and W2-W3. The thin smooth solid lines represent a least-square fit to the data points above spectral peaks.

![Fig. 3.6](image2.png)

Fig. 3.6. Frequency exponent (from Fig. 3.5) versus Keulegan-Carpenter number for all spectra. Only data points with \(R^2 > 0.8\) are shown.
The exponent $b$ is a measure of linearity (linear when $b=1$) of the relationship between energy dissipation, $S_{ds}(f)$ and incident spectrum, $E(f)$. The probability of occurrence of $b$ is plotted in Fig. 3.7 for the three frequency bands, within three ranges of $K_C$ numbers. The $K_C$ number is based on the entire spectrum. Note that a $K_C$ value of about 60 segregates first pair of gages, W1-W2, and the second pair, W2-W3. Fig. 3.7 shows that the relationship between $S_{ds}(f)$ and $E(f)$ is not consistently linear ($b \neq 1$) across the frequency scales. The relationship tends to be most linear in the wind sea band across the entire range of $K_C$ numbers, with slightly narrower distribution in the middle $K_C$ number range. The relationship between energy dissipation and incident spectrum becomes slightly more nonlinear in the swell frequency band. The coefficient $b$ tends to increase at smaller $K_C$ numbers (which are more common in the second reach, W2-W3). In the high-frequency band ($f>0.32$ Hz) the relationship between $S_{ds}(f)$ and $E(f)$ is linear for waves with $K_C < 47$, and gradually becomes nonlinear with increasing $K_C$ number. Note that the energy spectra and hence, the energy dissipation, in this high-frequency range is also affected by non-linear triad interactions.

![Fig. 3.7. Probability of occurrence of exponent $b$ (Eq. (12)) with respect to ranges of Keulegan-Carpenter number.](attachment:image.jpg)

Parameter $a$ in Eq. (3.12) was confirmed to be equal to the ratio of the integrated energy dissipation to the total wave energy, $S_v/m_b$, where $m_b = \int E^b df$ and $S_v = \int S_{ds} df$.

### 3.6 Estimates of Integral and Frequency-Dependent Bulk Drag Coefficients

The integral energy dissipation formulations (e.g., Mendez and Losada, 2004) assume the drag coefficient is independent of frequency and determine its single value, $\bar{C}_D$, for the entire spectrum, which is assumed to be narrow-banded. The variation of drag coefficient with the hydrodynamics has been typically related to the Reynolds ($R_e$) and Keulegan-Carpenter ($K_C$)
numbers using empirical relationships. Several studies have developed empirical formulations for integral estimates of $C_D$ (Bradley and Houser, 2009; Jadhav and Chen, 2012; Kobayashi et al., 1993; Mendez et al., 1999; Mendez and Losada, 2004; Paul and Amos, 2011; Sánchez-González et al., 2011). The empirical relationships are a valuable tool for predicting integral wave heights. For the data presented in this paper, the integral drag coefficients correlate well to the $K_c$ number ($R^2 = 0.95$) (Fig. 3.8), resulting in the following empirical formula:

$$C_D = 70 K_c^{-0.86} \quad 25 < K_c < 135$$  \hspace{1cm} (3.13)

Note that this $C_D$ represents the “bulk” value over the field study transect (vegetation patch), rather than the drag coefficient of an idealized, isolated, cylinder (e.g., Tanino and Nepf, 2008). The $C_D$ in Fig. 3.8 was estimated using Eq.(3.2).

Using the same equation, and allowing the drag coefficient to vary with frequency for each spectrum, produces a frequency distributed drag coefficient. Fig. 3.9 shows such distributions that are ensemble averaged over the three $K_c$ ranges. It is clear from these plots that the drag coefficient varies with the frequency, and a single integral drag coefficient value over the entire spectral frequency scale does not adequately represent the spectral evolution. This is most notable for the range containing the smallest $K_c$ numbers, where the drag coefficient varies by a factor of 6. Therefore, in studies of wave spectral evolution dominated by energy losses due to vegetation, a spectrally varying drag coefficient will more accurately predict wave energy dissipation.

![Fig. 3.8. Estimated integral bulk drag coefficient and its variation with the Keulegan-Carpenter number.](image)

Eq. (3.13) and Eq. (3.10) can be used to compute the frequency varying drag coefficient, $C_D$, when $C_D$ and $a_n$ are known. For a given spectrum (with its $K_c$), $C_D$ can be determined using Eq.(3.13). To calculate $a_n$, the following procedure was followed. Using the measured energy spectra, Eqs. (3.1) and (3.9) were numerically solved to compute $a_{n,j}$ for each frequency component of a spectrum. All $a_{n,j}$ profiles were then ensemble-averaged, producing the single $\bar{a}_n$ curve shown in Fig. 3.10. Across the spectrum of frequencies, $\bar{a}_n$ gradually increases up to the region of the peak, and then slightly decreases. The $\bar{a}_n$ values for frequencies above about 0.4 Hz are not considered reliable, due to the influence of non-linear energy transfer, and greater amplification of noise resulting from the pressure response function at those frequencies. The $\bar{a}_n$
values for frequencies above 0.4 Hz are therefore excluded from the following analysis. Multiplying the integral $\bar{C}_D$ obtained from Eq. (3.13) by values of $\bar{\alpha}_n$ (Fig. 3.10), provides values that can be used in Eq. (3.10) to calculate frequency-dependent values of $C_D$, that can be used to predict the frequency-dependent energy dissipation in wave spectra.

Fig. 3.9. Spectral variation of the bulk drag coefficient. All individual spectral distributions are ensemble-averaged based on $\kappa_c$ ranges.

Fig. 3.10. Spectral variation of ensemble-averaged velocity attenuation parameter, $\alpha_s$, based on all 118 measured profiles. Dashed lines represent $\pm 1$ standard deviation.
### 3.7 Prediction of Energy Dissipation using Estimated Drag Coefficients

To estimate energy dissipation due to vegetation in practical applications, selection of the appropriate drag coefficient is necessary. This section compares two approaches for selecting drag coefficients to determine which approach results in the better prediction of wave spectra in the presence of rigid-type vegetation. In the first, simple approach (existing standard practice), an integral drag coefficient, $\bar{C}_D$ (such as would be calculated using Eq. (3.13)) is specified and then spectral dissipation is calculated using Eq. (3.2). In the second approach, the frequency-dependent variable drag coefficient, $C_D$, is specified (Eq. (3.10)) and used in Eq. (3.2) to calculate spectral dissipation.

Fig. 3.11 shows comparison plots of the measured and predicted energy dissipation using these two approaches, for one wave record. The frequency-dependent $C_D$ predicts the frequency distribution of energy dissipation with better accuracy than the integral $\bar{C}_D$.

To quantitatively assess the predictive accuracy associated with the different approaches, over the entire dataset, the error between the measured and the predicted energy dissipation was calculated for each record and was ensemble averaged (Fig. 3.12a). In the frequency range with the dominant energy (0.03-0.36 Hz), the energy dissipation predicted by the frequency varying $C_D$ has much less error than that predicted by the integral $\bar{C}_D$. The improvement is especially significant in the vicinity of the spectral peak frequencies, where the largest dissipation is encountered. Additionally, Fig. 3.12b shows that, by employing the frequency-varying $C_D$, the model is able to predict total dissipation, $S_v$ (Eq. (3.11)) reliably.

The error in the prediction of $S_v$ is generally less than 5%. The mean error in the predicted $H_{mo}$ for the dominant frequency range (0.03-0.36 Hz) at gages W2 and W3 using the two methods ($\bar{C}_D$ and $C_D$) are (6.5% and 8.2%) and (-5.0% and -2.3%), respectively. At W2, the frequency-dependent $C_D$ method may appear slightly worse than the $\bar{C}_D$ method, however, the true advantage of the $C_D$ method is in the improved prediction of the frequency distribution of energy dissipation, as seen in Figs. 3.11a,b. This is reflected in the much better improvement in the estimate of mean period with errors being (-9.0% and 4.1%) and (-2.6% and 1.5%) at gages W2 and W3, respectively. Likewise the spectral width estimates are better when using $C_D$ compared to $\bar{C}_D$ with errors being (-25.1% and -5.4%) and (-9.2% and 2.1%) at gages W2 and W3, respectively.

### 3.8 Discussion

The Chen and Zhao (2012) formulation for energy dissipation caused by rigid vegetation has been reorganized by introducing the velocity attenuation parameter, $\alpha$. In this study, $\alpha$ is defined as the ratio of vegetation-attenuated orbital velocity inside the canopy at a given elevation, to the orbital velocity in the absence of vegetation at the same elevation. This is similar to the velocity attenuation parameter of Lowe et al. (2005), which was defined as the ratio of the velocity inside canopy to that outside canopy. These two versions of the velocity attenuation parameter are related by a factor which results from the decay of orbital velocity with respect to depth. To illustrate the equivalence of these two parameters, $\alpha$ was calculated using
the Tropical Storm Lee field data and compared to the velocity attenuation parameter values reported in Lowe et al. (2007, Fig. 3.5a). To this end, when calculating $\alpha$, the drag coefficient corresponding to the use of the velocity inside a canopy, $C_d$, was set to a fixed value of 2.5, as in Lowe et al. (2007). Fig. 3.9 shows that relatively stable value of the drag coefficient was observed for wave records with $K_c > 85$, so only those wave records were used for this comparison. The values of $\alpha$ plotted in Fig. 3.13 are the result of ensemble averaging 118 (59 wave records at each of the 2 gages, W2 and W3) $\alpha$ profiles. Comparison of Fig. 3.13 with Fig. 5a of Lowe et al. (2007) shows that, in both cases, the velocity attenuation parameter decreases gradually over the longer waves with the maximum values associated with shorter period waves.

The values of $\alpha$ associated with wave periods shorter than 2 s are unreliable due to observed non-linear energy transfers in that frequency band, and possible amplification of noise in the data analysis.

Fig. 3.11. Comparison of observed and predicted spectral energy dissipation using average and spectral drag coefficient for a sample wave record on September 3, 2011 at 12:30 UTC. (a) Dissipation between W1-W2 and (b) Dissipation between W2-W3. Dissipation based on $C_d$ values shown in (c) for W1-W2 and (d) for W2-W3.
Fig. 3.12. (a) Ensemble average of percentage error between the observed and estimated spectral energy dissipation using integral and spectrally variable drag coefficients. (b) Comparison of predicted and observed total energy dissipation.

Fig. 3.13. Variation of ensemble-averaged $a_j$ with wave period $T_j$. Dashed lines represent ±1 standard deviation.

Because the formulations for energy dissipation given in Eqs. (3.8) and (3.9) are based on the velocities at the same elevation inside a canopy, the results can be applied to cases involving shallow water and emergent vegetation. Further, Eq. (3.9) consists of explicit integration over discrete vertical increments and can be conveniently adopted when vertical variations of vegetation properties and hydrodynamics are important (e.g., Neumeier and Amos, 2006).

The velocity attenuation factor, $a$, is directly proportional to the normalized energy dissipation \( S_{dy}(f) / E(f) \) as is evident from Eq.(3.7). In the special case of shallow water, this equation simplifies to,
\[ \alpha_n^2 \propto \frac{S_{ds}}{E} \]  

(3.14)

The equivalence of \( \alpha_n^2 \) and \( S_{ds}/E \) is seen in the similarities between Fig. 3.5 and Fig. 3.10 in the dominant energy band. As shown in this study, the magnitude of the velocity attenuation factor is expected to decrease with increasing excursion (i.e., \( K_C \) number). The lower \( \alpha_n \) value reduces the normalized dissipation at the higher \( K_C \) numbers in Fig. 3.5, causing a steeper decline of the frequency distributions as shown.

In the prediction of drag-induced energy dissipation, the drag coefficient is an important input parameter, and attempt to universalize it remains a challenge. Consistent estimates of drag coefficients based on a range of wave and vegetation conditions will improve predictability of \( C_D \) as more data become available. Several complex processes are involved in the wave energy dissipation induced by vegetation drag. Laboratory studies of hydrodynamics around a single rigid circular cylinder in oscillating flows, in which force is modeled as a summation of inertial and drag forces by a Morrison-type equation (Morrison et al., 1950), contribute to understanding of these processes. Even in this simple form, under controlled conditions, the drag coefficients vary with time, Reynolds number, relative motion of the fluid, relative roughness, variable flow separation, wake interference, ambient turbulence, etc. (Sarpkaya, 1976). Additionally, in wavy flows (as opposed to simple oscillatory flows), velocity decays exponentially with depth and the orbital motion induces 3D flow effects and rotating vortices, further complicating the processes. Although Stokes’ solution exists for force coefficients in un-separated and laminar oscillating flows, such information must be obtained using experimental studies for separated flows, which are present in the field conditions (Sarpkaya, 1976). In the case of natural vegetation, the necessity of deriving drag coefficients from field studies is underscored by the fact that, to effectively model field conditions, these coefficients need to represent a stem array rather than a single cylinder (Tanino and Nepf, 2008). If the vegetation is flexible, then the consideration of the stem motion becomes essential (Mullarney and Henderson, 2010).

### 3.9 Summary and Conclusions

Random wave spectra were measured over salt marsh vegetation to study vegetation induced energy dissipation along a marsh transect with two reaches. The waves in the leading reach of the transect were more energetic, highly nonlinear, occurred in shallower water, and exhibited greater energy dissipation compared to the subsequent reach, where waves were less energetic, significantly less nonlinear, and exhibited less energy dissipation. Waves propagating over salt marsh vegetation dissipate energy due to drag induced by the stems. The magnitude of energy dissipation was observed to vary with the wave frequency. The greatest energy dissipation was observed near the incident spectral peak frequencies, with energy dissipation gradually decreasing with frequency above the peak. The rate of this decrease was greater for waves with larger \( K_C \) numbers and lower for waves with decreasing \( K_C \) numbers. Upon entering the vegetation, the low-frequency swell (<0.16 Hz) dissipated less in the leading reach of the measurement transect than the wind sea (0.16-0.32 Hz), carrying energy further and continuing the dissipation process in the subsequent reach of the transect. On the other hand, the majority of the wind sea energy dissipated in the leading reach of the transect. Across a spectrum, energy
dissipation did not linearly follow incident energy density and the degree of non-linearity varied with the frequency scale. The relationship of the spectral dissipation to energy density tended to be less nonlinear in the wind sea than the swell band, but the relationship became slightly more nonlinear and consistent (across bands) for waves with larger $K_c$ numbers. In general, the relationship was slightly more nonlinear in the swell band than the wind sea band.

The normalized wave energy dissipation ($S_{ds}(f)/E(f)$) was observed to be greatest near the spectral peak frequencies. The magnitude of the normalized dissipation was directly related to the frequency in the band below the peak, and inversely related to the frequency in the band above the peak of the wave energy density spectrum.

The vegetation induced drag coefficient was shown to vary with frequency. The distribution increased gradually up to the spectral peak and then remained generally uniform. The magnitude of the peak of this distribution was directly related to the magnitude of the corresponding $K_c$ number of the waves. The frequency-dependent drag coefficient was parameterized by introducing a normalized velocity attenuation parameter, $\alpha_n$. The spectral profiles of $\alpha_n$ were ensemble-averaged and a single $\bar{\alpha}_n$ curve was developed. This single curve along with the integral drag coefficient allowed for a prediction of the frequency-dependent drag coefficient. It was demonstrated that the frequency-dependent drag coefficient predicted the spectral distribution of energy dissipation with better accuracy than the integral drag coefficient.

The methodology and drag coefficient parameterization presented in this paper has been verified using the same dataset on which it is based. This validates the parameterization of the spectral bulk drag coefficient using a single velocity attenuation curve. This parameterization approach needs to be further tested using other, independent, datasets.

### 3.10 References


CHAPTER 4: PROBABILITY DISTRIBUTION OF WAVE HEIGHTS ATTENUATED BY SALT MARSH VEGETATION DURING TROPICAL CYCLONE

4.1 Introduction

Coastal wetlands reduce shoreline erosion impacts and provide natural defense against storm waves because wetland vegetation obstruct and dissipate waves (e.g., Costanza et al., 2008; Dixon et al., 1998; Gedan et al., 2011; Lopez, 2009). Utilization of coastal wetlands to augment structural measures for mitigation of coastal flooding due to storm surge and waves is promoted in several regions of the world (e.g., Borsje et al., 2011; CPRA, 2012). To design sea-defense structures against the extreme conditions, quantification of wave statistics is required. In deep water the waves are relatively linear and Gaussian, allowing a theoretical statistical description of the wave parameters. As the waves propagate shoreward, nearshore processes of depth-limited breaking and shoaling change the wave height distribution. If the waves then propagate over salt marsh vegetation, where obstructing vegetation dissipates wave energy, the wave height distribution is further changed.

Attenuation of integral wave heights and characteristics of energy dissipation due to vegetation have been studied in the laboratory and field conditions (for summary, see Anderson et al., 2011; Jadhav and Chen, submitted). However, to date, only one study has examined transformation of wave height distribution due to vegetation (Lövstedt and Larson, 2010). By measuring waves in reeds in a shallow lake, they observed that the distribution of wave heights was significantly different from the commonly assumed Rayleigh distribution for random variation in wave heights when longer wave propagation distances and higher waves are present. Their study was carried out in a low-energy environment (root-mean-square wave height=0.01-0.06 m and mean wave period=0.5-1.2 s). For the design of sea-defense works, studies involving storm-induced high energy wave conditions are essential. The need for and challenges of such field measurements were described by Smith et al. (2011).

The present study reports on a unique data set documenting propagation of waves through salt marsh vegetation during a tropical storm. We investigate the characteristics and transformation of wave height and wave period distribution due to natural vegetation, using wave measurements collected under tropical storm conditions. We demonstrate that the distribution of wave heights attenuated by vegetation deviates from the Rayleigh distribution routinely applied in deepwater analysis. Drawing on the literature describing wave distributions in the surf zone, we develop a modified Weibull probability density function (pdf) and estimate its parameters using the measured data. For prediction purposes, we further develop relationships between the Weibull parameters and the properties of the local wave field. The methodology can be used to predict characteristic wave heights such as the mean, root-mean-square or the average of the certain number of highest waves. A reliable estimate of storm wave height probability distribution is essential to the design of coastal structures for storm protection and planning of operational activities.
4.2 Wave Height Distribution Model

The short-term wave height statistics for deep water are well described by the Rayleigh probability density function (Longuet-Higgins, 1952).

\[
p(H) = \frac{2H}{H_{rms}^2} \exp \left[ -\left( \frac{H}{H_{rms}} \right)^2 \right] \quad 0 \leq H < \infty \tag{4.1}
\]

where, \( H \) is the local wave height and \( H_{rms} \) is the local root-mean-square wave height. In terms of non-dimensional wave height, \( \xi = H/H_{rms} \), the Rayleigh pdf, \( p \), and the cumulative distribution function (cdf), \( F \), are written as below.

\[
p(\xi) = 2\xi \exp(-\xi^2) \quad 0 \leq \xi < \infty \tag{4.2}
\]

\[
F(\xi) = 1 - \exp (-\xi^2) \tag{4.3}
\]

This distribution is based on the assumption that the waves are narrow-banded and linear, and that the water surface elevation follows Gaussian distribution. However, as waves propagate into shallow nearshore waters, the distribution of wave heights deviates from the Rayleigh distribution (see e.g., Dally, 1990; Ebersole and Hughes, 1987; Hameed and Baba, 1985; Mase, 1989). Further, when salt marsh vegetation is present, waves undergo dissipation due to drag offered by vegetation, which also causes changes in the wave height distribution (Lövstedt and Larson, 2010).

In this section, we derive an expression for the vegetation-transformed wave height distribution. In the derivation, each wave in the incident distribution is transformed in accordance with the theories of wave height attenuation due to vegetation. Treatment of random waves as a collection of individual regular waves is a method that is used to examine wave height distributions in the surf zone (Battjes and Groenendijk, 2000; Dally, 1990; Mase and Iwagaki, 1982; and Mendez et al., 2004).

Waves propagating through rigid vegetation dissipate energy due to the drag produced by the vegetation. Assuming normally incident linear waves, and treating vegetation as rigid obstructing cylindrical elements that impart drag forces on the monochromatic waves, Dalrymple et al. (1984) expressed wave attenuation as follows,

\[
H = \frac{H_o}{1 + \beta_1 H_o} \tag{4.4}
\]

where

\[
\beta_1 = \frac{4}{9\pi} C_D b_v N_v k \frac{\sinh^3 ksh + 3\sinh ksh}{(\sinh 2kh + 2kh)\sinh kh} \tag{4.5}
\]
and, $H_o$ is the incident wave height, $H$ is the local attenuated wave height, $k$ is the wave number, $h$ is the still water depth, $s$ is the ratio of vegetation height ($h_v$) to the still water depth ($h$), $b_v$ is the stem diameter, $N_v$ is the vegetation density, and $C_D$ is the bulk drag coefficient. The cross-shore coordinate is represented by $x$. Note that $\beta_1$ has the units of $[L^{-1}]$. Eq. (4.4) is the solution of the wave energy balance equation on a flat bottom topography where the source term due to vegetation induced energy dissipation is dominant.

Assuming that the incident wave heights, $H_o$, exhibit the Rayleigh distribution, the incident wave height pdf is expressed as,

$$p_o(H_o) = \frac{2H_o}{H_{rms,o}^2} \exp \left[ -\left( \frac{H_o}{H_{rms,o}} \right)^2 \right] \quad 0 \leq H_o < \infty \quad (4.6)$$

where, $H_{rms,o}$ is the root-mean-squared incident wave height. When transformed, the pdf becomes,

$$p(H) = p_o(H_o) \left| \frac{\partial H_o}{\partial H} \right| \quad (4.7)$$

Using Eq. (4.4), $H_o$ can be expressed in terms of $H$ as,

$$H_o = \frac{H}{1 - \beta_1 H} \quad (4.8)$$

This gives us,

$$\left| \frac{\partial H_o}{\partial H} \right| = \frac{1}{(1 - \beta_1 H)^2} \quad (4.9)$$

Substituting Eq. (4.6) and Eq. (4.9) into Eq. (4.7), we get,

$$p(H) = \frac{2H}{(1 - \beta_1 H)^3} \frac{1}{H_{rms,o}^2} \exp \left[ -\left( \frac{H}{(1 - \beta_1 H) H_{rms,o}} \right)^2 \right] \quad 0 \leq H < 1/\beta_1 \quad (4.10)$$

Eq. (4.10) is a model of the vegetation-transformed wave height distribution, developed from the Rayleigh distribution and the wave height decay model. The model depends on the history of the waves and the incident root-mean-square wave height, $H_{rms,o}$, and thus can be referred to as a “propagation model” as opposed to a “local model” (Battjes and Groenendijk, 2000). Local models assume that the wave height distribution is primarily determined by the local wave parameters, irrespective of the history of the incident waves. The propagation-type model described by Eq. (4.10) can be converted to a local model, if $H_{rms,o}$ can be expressed in
terms of the local $H_{rms}$. Following Mendez and Losada (2004), for narrow-banded waves attenuating through rigid vegetation, $H_{rms,o}$ can be expressed in terms of local $H_{rms}$, as follows,

$$H_{rms,o} = \frac{H_{rms}}{1 - \beta_2 H_{rms}} \quad (4.11)$$

where

$$\beta_2 = \frac{1}{3 \sqrt{\pi}} C_{D,r} b_v N_v k_r \frac{\sinh^3 k_r sh + 3 \sinh k_r sh}{(\sinh 2k_r h + 2k_r h) \sinh k_r h} x \quad (4.12)$$

and, subscript $r$ indicates the representative value. To obtain a local model, substitute Eq. (4.11) into Eq. (4.10) to get,

$$p(H) = \frac{2H}{(1 - \beta_1 H)^3} \frac{(1 - \beta_2 H_{rms})^2}{H_{rms}^2} \exp \left[ -\frac{H (1 - \beta_2 H_{rms})}{(1 - \beta_1 H) H_{rms}} \right] \quad 0 \leq H < 1/\beta_1 \quad (4.13)$$

To simplify this expression, we can define a local non-dimensional parameter,

$$\kappa = \beta_1 H_{rms} \quad (4.14)$$

and, also a parameter,

$$\phi = 1 - \beta_2 H_{rms} \quad (4.15)$$

Substituting Eq. (4.14) and Eq. (4.15) into Eq. (4.13), we get,

$$p(H) = \frac{2H}{(1 - \frac{\kappa}{H_{rms}} H)^3} \frac{\phi^2}{H_{rms}^2} \exp \left[ -\phi^2 \left( \frac{H}{(1 - \frac{\kappa}{H_{rms}} H) H_{rms}} \right) \right] \quad 0 \leq H < H_{rms}/\kappa \quad (4.16)$$

Expressing this pdf equation in terms of non-dimensional wave height, $\xi = H/H_{rms}$, gives,

$$p(\xi) = \frac{2\phi^2 \xi}{(1 - \kappa \xi)^3} \exp \left[ -\phi^2 \left( \frac{\xi}{1 - \kappa \xi} \right) \right] \quad 0 \leq \xi < 1/\kappa \quad (4.17)$$

The cdf is given by,

$$F(\xi) = 1 - \exp \left[ -\phi^2 \left( \frac{\xi}{1 - \kappa \xi} \right) \right] \quad 0 \leq \xi < 1/\kappa \quad (4.18)$$
This distribution, described using the shape parameter $\kappa$ and scale parameter $\phi$, is similar in form to the Weibull distribution (Kies, 1958; Nadarajah and Kotz, 2006; Phani, 1987). In this case, however, the parameter $\phi$ is not independent and can be shown to be a function of $\kappa$, by eliminating $H_{rms}$ between Eq. (4.14) and Eq. (4.15) as,

$$\phi = 1 - \frac{\beta_2}{\beta_1} \kappa$$  \hspace{1cm} (4.19)

Using the expressions for $\beta$ values from Eq. (4.5) and Eq. (4.12), we get

$$\phi = 1 - \frac{3\sqrt{\pi}}{4} \kappa = 1 - 1.33\kappa$$  \hspace{1cm} (4.20)

Note that the hyperbolic terms in the $\beta$ expressions would cancel out only for very narrow banded wave field. Nevertheless, Eq. (4.20) suggests the form of the dependence that can be fitted to the estimated parameters. Incidentally, this sets the upper limit of $\kappa$ to $1/1.33=0.75$. Mendez et al., (2004) have obtained the same form of distribution (Eq. (4.17)) for the wave height distribution on a planar beach due to shoaling and breaking. However, their relationship between $\phi$ and $k$ was obtained by numerical curve fitting as $\phi = (1 - \kappa^a)^b$ with $a = 0.944$ and $b = 1.187$. Note that, for $\kappa \ll 1$, and $a \approx 1$, this equation approximates to $\phi = 1 - 1.187\kappa$.

The distribution in Eq. (4.17) has only one independent parameter, $\kappa$. In Section 5, $\kappa$ is estimated by the maximum-likelihood method, and then correlated with the ambient wave parameters. To aid in this exercise, the dependencies of $\kappa$ are examined by expressing it in terms of wave parameters as follows. From Eq. (4.14) and Eq. (4.5), parameter $\kappa$ is given by,

$$\kappa = \frac{4}{9\pi} C_D b_v N_v k \frac{\sinh^3 ksh + 3\sinh ksh}{(\sinh 2kh + 2kh)\sinh kh} (x) H_{rms}$$  \hspace{1cm} (4.21)

For shallow water this can be approximated as,

$$\kappa \approx \frac{1}{3\pi} C_D b_v N_v s \frac{H_{rms}}{h} x$$  \hspace{1cm} (4.22)

The shape parameter, $\kappa$ is thus directly proportional to the drag coefficient, vegetation characteristics ($b_v, N_v, s$) and the ratio of local characteristic wave height to depth.

$$\kappa \propto C_D, b_v, N_v, s, \frac{H_{rms}}{h}$$  \hspace{1cm} (4.23)

Considering, the inverse relationship of $C_D$ to $K_c$ (e.g., Jadhav and Chen, submitted),

$$\kappa \propto \frac{1}{K_c}, b_v, N_v, s, \frac{H_{rms}}{h}$$  \hspace{1cm} (4.24)
In Section 5, an empirical equation to estimate the distribution parameter, $\kappa$, is developed in terms of the variables on the right hand side of Eq. (4.24) by fitting to the measured data.

The characteristic wave heights of the proposed distribution can be obtained by the following expression,

$$\xi_{1/q} = q \int_{\xi_q}^{1/\kappa} \xi p(\xi) d\xi$$

(4.25)

where, $\xi_{1/q}$ is the mean of the highest $1/q$ normalized wave heights and $\xi_q$ is the normalized wave height at the exceedance probability of $1/q$. The term $\xi_q$ is computed using Eq. (4.17) as,

$$\xi_q = \frac{\sqrt{\log (q)}}{\phi + \kappa \sqrt{\log (q)}}$$

(4.26)

### 4.3 Field Data Collection

The experimental set up is described in Section 3.3. All gages were self-logging pressure sensors that sampled continuously at 10 Hz over the duration of the storm. The time series of continuous pressure measurements from wave gages were segmented into 15-min bursts or wave records. The wave records were analyzed using Fourier analysis and wave-by-wave analysis. This paper primarily uses wave-by-wave analysis. The Fourier analysis was performed using standard spectral techniques and is described in Jadhav and Chen, submitted. All measurements recorded while the water depth was less than 0.4 m were eliminated from further consideration, because the wave energy was found to be negligible under those conditions. Thus, the study represents submerged vegetation conditions only.

The wave energy loss due to vegetation was considered dominant compared to the other energy loss source terms. To ascertain the validity of this assumption, the relative magnitude of source terms for the local wave generation and the losses due to bottom-friction, white-capping, and depth-limited breaking were evaluated. The wave records, with significant potential for the magnitude of these source terms to be dominant, were removed from further analysis (for details see Jadhav and Chen, submitted).

The zero-crossing method is used to obtain distributions of wave height and wave period. First, the water surface elevation time series corresponding to the pressure time series of each wave record is estimated using the method proposed by Nielsen (1989). In this method, a sine curve is fitted locally with the locally defined frequency. A semi-empirical transfer function is then applied to obtain a water surface elevation series. To eliminate data noise, third-neighbor points are used in the computations, instead of adjacent points. Wave height is defined as the difference between the maximum and minimum water surface elevation occurring between two consecutive zero-crossings. The corresponding wave period is defined as the time between the same zero-crossings.
4.4 Observed Wave Conditions

A total of 177 wave records (59 records at each of the 3 gages) collected over two days were analyzed in this study. Fig. 4.1 shows wave conditions during the two day period through time series plots of the hydrodynamic and bulk wave parameters. With the diurnal tide augmented by the storm surge, the water depth rose from about 0.1 m to 0.8 m and then fell along with the tide. The incident significant wave heights ($H_{mo}$) on the marsh varied from 0.05 to 0.39 m and were directly proportional to the depth of flood water. The relative depth, $h/L_z$, was less than 0.2, indicating relatively shallow water depths during the observation period. The wave length $L_z$ is based on the mean wave period, $T_z(=\sqrt{m_0/m_2})$, where $m_0$ and $m_2$ are the zero-th and second moments of the frequency spectrum, respectively. The mean wave period was around 2 s, though the observed spectral peak period (not shown) was as much as 4.5 s.

Spectral width, $\nu$, defined as $\sqrt{(m_0m_2/m_1^2 - 1)}$, ranged from 0.42 to 0.86, an indicator of the broad nature of the observed spectra. Fig. 4.1 also shows the Ursell number, $H_{mo}L_z^2/h^3$, as a measure of non-linearity. Waves were largely non-linear at W1 but the non-linearity quickly decreased as the waves propagated further into the marsh (beyond W2) and were dissipated by the vegetation.

The wave height distribution obtained from the wave-by-wave analysis was examined for all the wave records. The distribution was observed to deviate from the theoretical Rayleigh distribution (Fig. 4.2), which overestimates larger wave heights.

The observed spectra were bimodal (not shown), with a low-energy, low-frequency persistent swell (peak period, 7-10 s) in addition to the wind sea (peak period, 2-4.5 s). The impact of mixed states on the wave height distribution depends on the ratio of the wave energy in each wave system and the intermodal distance (Rodriguez et al., 2002). The sea-swell wave energy ratio is defined as $m_{0,sea}/m_{0,swell}$, while the intermodal distance is defined as $(f_{p,sea} - f_{p,swell})/(f_{p,sea} + f_{p,swell})$, where $f_p$ is the spectral peak frequency. For the wave records analyzed in this study, the ratio of wind sea energy to swell energy was 4 to 8, while the intermodal distance was 0.5 to 0.7, indicating wind sea dominated, significantly separated spectra. Under such conditions, the swell has no significant impact on the wave height distribution (Rodriguez et al., 2002). In the next section, the proposed Weibull-type distribution developed in Section 4 is parameterized.

4.5 Parameter Estimation of the Model

The derived wave height distribution model defined by Eq. (4.18) was fitted to the observed wave height distribution using the Maximum Likelihood Method. Wave records collected only during the first day of measurements were used for this exercise, and the wave measurements from the second day were used for the validation of the model (Section 6). The parameters, $\kappa$ and $\phi$, of the pdf, were calibrated using 123 wave records, each containing 250 to 300 waves. Fig. 4.3 shows an example calibration of wave records from the 3 wave gages.
Fig. 4.1. Wave conditions at the study site during Tropical Storm Lee. (a) Water depth measured by wave gages (5-min averaged from the continuous record), (b) Spectral significant wave height, (c) Mean wave period (d) Relative depth (e) Spectral width, and (f) Ursell number.
Fig. 4.2. An example of deviation of observed wave height distribution at Gage W3 during a 15-min burst (296 waves). (a) Observed wave height histogram with Rayleigh distribution (red line). (b) Observed cumulative wave height distribution (blue circles) relative to the Rayleigh distribution (red line).

Fig. 4.3. Estimated parameter $\kappa$ during for a wave record at each gage. Top panel: Histograms of observed values and $pdfs$. Bottom panel: $cdf$s on Rayleigh paper. Solid blue lines are proposed distribution and dashed lines are Rayleigh distribution. Circles in bottom panel are observed values.
The calibrated theoretically derived distribution (solid blue line) is more similar to the observed distribution than the Rayleigh distribution (dashed black line). Over the 123 records, the estimates ranged from 0.02-0.42 for $\kappa$ and from 0.51-0.98 for $\phi$. Fig. 4.4 shows the range and relationship between these two parameters. The relationship is expressed as,

$$\phi = 1.0 - 1.2\kappa$$  \hspace{1cm} (4.27)

The linear form of this relationship agrees with the theory (Eq.(4.20)). The multiplier of $\kappa$ in the above equation is slightly lower than the theoretically determined value of 1.33, which is based on the assumption of narrow-banded spectra.

Fig. 4.4. Relationship between parameters $\kappa$ and $\phi$ estimated using all wave records from the first day at gages W1, W2 and W3.

Eq. (4.24) suggests possible candidates to parameterize $\kappa$ using local wave field parameters. For a given vegetation field (constant $b_v$ and $N_v$), $\kappa$ is inversely related to the Keulegan-Carpenter number ($K_C$) and directly proportional to the relative wave height, $H_{rms}/h$. The Keulegan-Carpenter number is defined as, $K_C = u_b T_z / b_v$, where $u_b$ (= $\pi H_{rms}/T_z \sinh k_z h$) is the near-bed orbital velocity, and $k_z$ is the wave number based on the mean wave period, $T_z$. Fig. 4.5a shows $\kappa$ parameterized in this manner. From Eq. (4.27), the upper limit of $\kappa$ is 0.83 (reciprocal of 1.2). The observed data are categorized based on the range of relative wave height, $\gamma = H_{rms}/h$. The relationship between $\kappa$ and $K_C$ is expressed as,

$$\kappa = 0.83 \exp(-mK_C)$$  \hspace{1cm} (4.28)

In general, exponent $m$ is significantly correlated to $\gamma$, as shown in Fig. 4.5b, as described by the following equation.

$$m = 0.044 \gamma^{-1.11}$$  \hspace{1cm} (4.29)
Fig. 4.5. (a) Relationship between the estimated parameter $\kappa$ and the observed Keulegan-Carpener number, $K_C$, grouped by measured $\gamma = H_{rms}/h$; (b) Relationship between the estimated exponent $m$ in the left figure observed and $\gamma$. Symbol ‘x’ shows exponential wave height decay rate estimated for the same data in Jadhav and Chen, submitted.

Also, the exponent $m$ is found to be closely related to the exponential wave height decay parameter, $k_H$, as shown in Fig. 4.5b. The exponential decay of the wave height propagating through vegetation is expressed as (e.g., Jadhav and Chen, submitted; Kobayashi et al., 1993),

$$H = H_o \exp \left( -k_H x \right)$$  \hspace{1cm} (4.30)

where, $H_o$ is the wave height incident to the vegetation patch and $H$ is the attenuated wave height after the wave has travelled distance, $x$, landward. The $k_H$ values shown in Fig. 4.5b are estimated by fitting Eq. (4.30) to the wave measurements at the three gages (see for details, Jadhav and Chen, submitted). Fig. 4.5, thus, relates characteristics of wave height attenuation to the characteristics of wave height distribution. For a given $\gamma$ value, exponent $m$ can be determined from Eq.(4.29), and using known $K_C$, distribution parameter $\kappa$ can be determined from Eq. (4.28). Then Eq. (4.27) can be used to determine the remaining parameter, $\phi$. Along with the known $H_{rms}$, Eq. (4.18) describes the complete wave height distribution.

To calculate $H_{rms}$, its relationship to the variance of surface elevation or related parameters is required. Using the observed data, we developed the following empirical relationship between $H_{rms}/m_0^{1/2}$ and the Ursell number, $U_{rs}$ ($=H_{rms}L_2^2/h^3$) as shown in Fig. 4.6.

$$H_{rms}/m_0^{1/2} = 2.83U_{rs}^{0.09}$$  \hspace{1cm} (4.31)

For this purpose, in similar studies of wave height distribution in the nearshore, Battjes and Groenendijk, (2000) selected $m_0^{1/2}/h$ while Mendez et al. (2004) adopted $\kappa$ as an independent variable.
Fig. 4.6. Variation of the ratio $H_{rms}/m_o^{1/2}$ with the Ursell number based on observations from the first day.

4.6 Validation of the Model

To validate the proposed model, the wave height distribution predicted by Eq. (4.17) is compared with the observed distributions on the second day of the study period. Using known spectral moments, $m_0$, $m_2$, and water depth, $h$, Eq. (4.31) is used to calculate the local $H_{rms}$. The ratio, $\gamma = H_{rms}/h$, is used in Eq. (4.29) to determine the exponent $m$ required for applying Eq. (4.28). The shape parameter, $\kappa$, is calculated from Eq. (4.28) and the scale parameter, $\phi$ from Eq. (4.27). The $K_C$ number is calculated based on $m_0$ as stated previously. Fig. 4.7 shows comparison of the predicted and observed wave height distributions at three gages from one wave record. The proposed model accurately captures observed deviations of the wave height distribution from the Rayleigh distribution.

As an overall indicator of the model performance, the error is quantified in terms of the normalized root-mean-square error (NRMSE) defined as,

$$NRMSE = \sqrt{\frac{1}{N} \sum_{q=1}^{N} \left(\frac{H_{q,\text{pred}}}{H_{q,\text{obs}}} - 1\right)^2}$$

(4.32)

where $N$ is the total number of wave records, $H_q$ is the wave height with the probability of exceedance $q$, and the subscripts $\text{pred}$ and $\text{obs}$ denote the predicted and measured values, respectively. The characteristic wave heights are calculated by numerical integration of Eq. (4.25). Fig. 4.8 shows magnitudes of relative error for $H_{10\%}$, $H_{2\%}$ and $H_{1\%}$ considering all wave
records together. On average, the NRMSE of the proposed distribution was found to be 77%, 57%, and 50% less than the NRMSE of the Rayleigh distribution, for $H_{10\%}$, $H_{2\%}$ and $H_{1\%}$, respectively.

Fig. 4.7. An example of predicted pdf (top) and cdf (bottom) during a wave record at the three gages. Solid red lines are the predicted distribution and the dashed lines are Rayleigh distribution.

Fig. 4.8. Normalized RMS error in the various characteristics wave heights predicted for the second-day wave conditions at all gages combined.
4.7 Discussion

When the wave height distribution deviates from the Rayleigh pdf, the ratios of characteristic wave heights to $\sqrt{m_0}$ change, and the theoretically derived values cannot be used. For example, the ratio $H_{rms}/\sqrt{m_0}$ for waves attenuated by salt marsh vegetation increased from the theoretical value of $\sqrt{B} = 2.83$ as the shape of the distribution (characterized by $\kappa$) changed, as shown in Fig. 4.9. As previously seen in Fig. 4.6, this ratio increases gradually with nonlinearity. A similar relationship between $H_{rms}/\sqrt{m_0}$ and $\kappa$ was found for wave breaking on a planar beach without vegetation by Mendez et al. (2004).

It should be noted that Eq. (4.28), does not account for vegetation characteristics ($b_v, N_v, s$) explicitly, even though the theory shows such dependence (Eq. (4.24)). It is anticipated that, at higher vegetation obstruction (solid volume fraction, $\pi N_v b_v^2/4$), the measured distribution will be still lower in the low exceedance region, indicating higher shape parameter values, $\kappa$. This would mean, for the same $K_c$, the values of exponent $m$ will be smaller than those shown in Fig. 4.5b. This is possible if the curve in Fig 4.6b is shifted lower. Thus the variability of $b_v$ and $N_v$ will be evident in the multiplier of the power law relation given by Eq. (4.29). More field data collected at different sites are needed to quantify the effects on vegetation properties on the shape parameter of the theoretically derived Weibull function for wave height distribution.

![Fig. 4.9. Variation of measured ratio $H_{rms}/\sqrt{m_0}$ with shape parameter, $\kappa$, of the wave height pdf.](image-url)
4.8 Conclusions

The study reports on a unique set of data consisting of wave height distributions of tropical cyclone-induced waves attenuated by salt marsh vegetation. The data was collected along a linear transect of 3 wave gages, over two days and consisted of 177 wave records of waves propagating over rigid salt marsh vegetation. The measured wave height distributions were observed to deviate from the Rayleigh probability distribution that is commonly used for waves in deep water. The observed probability densities of the higher wave heights were reduced significantly, producing wave heights lower than those predicted by the Rayleigh distribution. Assuming Rayleigh distributed wave heights for the incident waves to the vegetation patch, a probability distribution function is derived using the existing formulations of vegetation-induced wave height attenuation. The distribution is a function of the local parameters only. The proposed distribution function is a form of two-parameter Weibull function. However, it is theoretically shown that the scale parameter can be expressed as a function of the shape parameter, effectively reducing the proposed distribution to a one-parameter type. The single (shape) parameter of the proposed distribution is estimated using measured wave height distributions on the first day. It is then parameterized in terms of the Keulegan-Carpenter number and the relative wave height; two variables suggested by the theoretical dependencies of the shape parameter.

The pdf is then used with the shape parameter determined from the derived empirical expressions to estimate wave height distributions for the wave conditions on the second day. The proposed pdf predicts the reduced probability density in the low-exceedance range. The normalized root mean square error between the measured and predicted characteristic wave heights is reduced by 50-77% compared to the Rayleigh estimates.

Based on the analysis presented in this chapter, to estimate the characteristic wave heights attenuated by salt marsh vegetation at a given location, the following steps are applied.

1. Determine the local water depth, the local spectral wave energy, and the mean wave period.
2. Calculate the Keulegan-Carpenter number and the Ursell number at the location.
3. Using Eq. (4.31) determine the local root-mean-square wave height.
4. Using Eq. (4.28) and Eq. (4.29) calculate the shape parameter.
5. Using Eq. (4.27) calculate the dependent scale parameter.
7. Calculate characteristic wave heights (e.g., $H_{1%}$) using Eq. (4.25).

It should be noted that this is the first and only study, to-date, which has quantified and parameterized the wave height distribution of waves attenuated by salt marsh vegetation. The robustness of the empirical expressions and parameterizations derived using this data set will increase as more such studies become available.
4.9 References


5.1 Introduction

Many coastal regions of the world experience tropical storms, and the resulting surge, annually. The northern coast of the Gulf of Mexico is particularly vulnerable to such events. In the last 50 years, the Louisiana coast has been impacted by 14 major hurricanes, including Katrina (2005), Rita (2005), Gustav (2008) and Ike (2008). According to some estimates, the region is more than twice as likely to see major hurricanes than the Texas and Florida coasts (Resio, 2007). Hurricanes Katrina and Rita converted 562 km² of coastal land to water in Louisiana (Barras, 2006). The impact of the devastation caused by the hurricane surge and waves to human life and property along the coast has been enormous. For example, in 2005, after Hurricane Katrina, more than a quarter of a million people were displaced, more than 1,500 people lost their lives, and the property damage exceeded $100 billion (Graumann et al., 2005).

When considering mitigating hurricane impacts, it is generally acknowledged that coastal wetlands provide a natural first line of defense against damage by storm surge and waves (e.g., Lopez, 2009). By one estimate, in the US, the coastal wetlands provide $23.2 billion in storm protection services annually (Costanza, 2008). Federal and State agencies have committed significant financial resources to maintaining and improving surge/wave reduction and ecological benefits of coastal wetland through restoration and protection efforts (CPRA, 2012).

To protect communities from storm surge and waves, traditionally, levees and floodgates have been employed. In many situations, this solution has proven costly, unsustainable, and short sighted, causing unintended ecosystem consequences by disturbing the deltaic processes (Day et al., 2007). There has been renewed interest in capitalizing on the potential of natural coastal wetlands to reduce impacts of storm surge. Research is needed to provide field measurements of surge attenuation and collective resistance by wetland vegetation for coastal engineering applications (Irish et al., 2008).

The potential of wetlands to dampen storm surge has been expressed by empirical rules of thumb based on observation, e.g., storm surge could be reduced by 1 m over an inland length of 14.5 km. However, use of these rules of thumb has been acknowledged as outdated (USACE 2006). Recent studies point out that such constant rates do not account for transient forcing and local topography (Resio and Westerink, 2008). There have been numerical studies to understand surge attenuation potential of coastal wetlands (e.g., Wamsley et al., 2009; Wamsley et al., 2010). The current literature, however, has scarce field data sets. The goal of this study is to quantify the characteristics of surge propagation over coastal marsh using field measurements.

5.2 Data and Methods

During the study period, the Louisiana coast experienced three tropical storms, Tropical Storm Ida (November 10, 2009), Tropical Storm Bonnie (July 25, 2010), and Tropical Storm Lee...
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(November 3, 2011), but no hurricanes. Several water level gages were deployed in Breton Sound during Tropical Storm Ida, and in Terrebonne Bay during Tropical Storm Lee. The data collection methods for each storm are described in separate sub-sections below.

5.2.1 Tropical Storm Ida

Surge gages were deployed in the marshes of upper Breton Sound estuary. The Breton Sound estuary covers about 270,000 km$^2$ in Plaquemines and St. Bernard parishes of Louisiana (www.lacoast.gov). It is bounded on the west by the Mississippi River, on the east by the Mississippi River Gulf Outlet (MRGO), and on the north by Bayou La Loutre. Chandeleur barrier island chain is located at about 35 km seaward of the marshes. The sound is the remnant of a Mississippi River delta lobe, the abandoned St. Bernard Delta. The prevalent vegetation communities in the marshes are smooth cordgrass (Spartina alterniflora) and saltmarsh meadow (Spartina patens). The health of the vegetation varies with elevation, exposure to the waves, and salinity regime. The plant density is seasonal, with maximum density during the summer months.

Tropical Storm Ida was a late season (4-10 November 2009) hurricane (Fig. 5.1). Ida was the first November hurricane in the Gulf of Mexico since Kate in 1985 (Avila and Cangialosi, 2009). On Monday November 9, 2009 at 12:00 PM CST (18:00 UTC), according to National Hurricane Center Advisory Number 23A (Fig. 5.1), Ida was moving NNW at 18 mph (30 km/hr), with maximum sustained winds of 70 mph (113 km/hr). The center was expected to make landfall near Dauphin Island, Alabama on Tuesday morning.

On the morning of November 9, 2009 between 10 AM to 2 PM CST, four pressure sensor gages (E,F,I, and J) were deployed by boat in the marshes near Mozambique Point in upper Breton Sound. The locations of these gages and existing USGS monitoring stations in the area are shown in Fig. 5.2. Gage J was deployed at a location as far south as it was possible to travel...
safely in the face of wind and waves. Then, moving northwards, Gages F, I and E were placed in the marshes adjoining Bayou Terre aux Boeufs. Finally, Gage G was deployed in Lake Lery.

Fig. 5.2. Locations of USGS gages and gages deployed for Ida (left). Close-up view of the locations of the gages deployed for Ida (right). Storm track (not shown) is north-south, approximately 90 km to the east of gage J.

All four gages were retrieved on November 16, 2009, several days after the surge receded. This deployment provided approximately seven days of continuously recorded water levels. As an example, Fig. 5.3 shows the location of Gage I photographed on the days of deployment and retrieval.

Fig. 5.3. Location of Gage I photographed during deployment (left) and during retrieval (right) for Tropical Storm Ida.
The sampling frequency of the pressure sensors was set to 1.67 Hz, 2 Hz and 4 Hz depending on the sensor. The gages sampled either continuously or in a burst mode for approximately seven days until the memory became full. The topographic elevations of the gages were not surveyed with respect to NAVD datum.

Table 5.1. Coordinates of the gages deployed during Tropical Storm Ida

<table>
<thead>
<tr>
<th>Gage ID</th>
<th>Northing</th>
<th>Easting</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>J</td>
<td>29 39.7065'</td>
<td>89 34.0103'</td>
<td>Southernmost; in open water at a depth of 5.25 ft</td>
</tr>
<tr>
<td>F</td>
<td>29 39.7354'</td>
<td>89 34.0438'</td>
<td>At the edge of the marsh; nearest to the open water</td>
</tr>
<tr>
<td>I</td>
<td>29 39.5458'</td>
<td>89 33.6793'</td>
<td>On the marsh</td>
</tr>
<tr>
<td>E</td>
<td>29 40.0340'</td>
<td>89 34.9766'</td>
<td>On the marsh</td>
</tr>
<tr>
<td>G</td>
<td>29 47.9250'</td>
<td>89 48.1544'</td>
<td>Northernmost; in Lake Lery</td>
</tr>
</tbody>
</table>

5.2.2 Tropical Storm Lee

Water level gages were deployed in the upper marshes of Terrebonne Bay. The bay is located on the west side of Barataria Bay which is west of the Mississippi River. Terrebonne Bay is bounded by Bayou Terrebonne on the east and the Houma Navigation Canal and Bayou Little Caillou on the west. It is bordered on the south by a series of narrow, low-lying barrier islands of the Isles Dernieres and the Timbalier Islands, approximately 15 km south of the northern marshes. The prevalent vegetation communities in these marshes are smooth cordgrass (*Spartina alterniflora*) and saltmarsh meadow (*Spartina patens*).

Tropical Storm Lee made landfall about 20 km south-southeast of Intracoastal City, Louisiana on September 4, 2011 (Brown, 2011). On September 2, six water level gages (pressure sensors) were deployed by airboat in the marshes between Bayou Little Caillou and Bayou Terrebonne in the upper Terrebonne Bay. The locations of the gages are shown in Fig. 5.4. Table 5.2 shows coordinates, topographic elevations and the distances between the gages. The gages were HOBO U20 Water Level Data Loggers (U20-001-01). All six gages sampled pressure at 1 min frequency for approximately seven days.

Table 5.2: Coordinates and inter-gage distances of surge gages deployed during Tropical Storm Lee.

<table>
<thead>
<tr>
<th>Gage ID</th>
<th>Northing</th>
<th>Easting</th>
<th>Elevation (m, NAVD88)</th>
<th>Distance from S1 (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>S1</td>
<td>N 29° 13.923'</td>
<td>W 90° 36.985'</td>
<td>0.35</td>
<td>0</td>
</tr>
<tr>
<td>S2</td>
<td>N 29° 14.349'</td>
<td>W 90° 37.538'</td>
<td>0.28</td>
<td>1.2</td>
</tr>
<tr>
<td>S3</td>
<td>N 29° 14.859'</td>
<td>W 90° 38.015'</td>
<td>0.31</td>
<td>1.2</td>
</tr>
<tr>
<td>S4</td>
<td>N 29° 15.315'</td>
<td>W 90° 38.506'</td>
<td>0.29</td>
<td>1.2</td>
</tr>
<tr>
<td>S5</td>
<td>N 29° 16.279'</td>
<td>W 90° 38.393'</td>
<td>0.29</td>
<td>1.8</td>
</tr>
<tr>
<td>S6</td>
<td>N 29° 17.052'</td>
<td>W 90° 38.241'</td>
<td>0.16</td>
<td>1.5</td>
</tr>
</tbody>
</table>
The six gages were placed along a generally north-south transect, starting from the southernmost fringing marsh. Fig. 5.5 shows a typical surge monitoring location at the onset of the storm surge and after the surge retreated.

Fig. 5.4. Tropical Storm Lee track and surge gage locations (left). A close-up view of the same surge gage locations (right).

Fig. 5.5. Typical surge monitoring gage location with (left) and without (right) the surge from Tropical Storm Lee.
5.3 Results

5.3.1 Analysis of Tropical Storm Ida Surge

Winds during Ida peaked at midnight on November 9, 2009 as shown by the record at the NOAA meteorological station (SHBL1 No. 8761305) at Shell Beach, LA (Fig. 5.6). The pressure transducers rapidly deployed on the morning of November 9 provided approximately seven days of continuous record of water levels at four locations (Fig. 5.2).

![Wind Speed](image1)

![Water Depth](image2)

Fig. 5.6. Wind recoded at Shell Beach, LA NOAA station (SHBL1 No. 8761305) (top) and water levels recorded during Tropical Storm Ida (November 2009) (bottom).

In the early hours of the monitoring period, the surge in Breton Sound marsh rose against the north, north-easterly winds. The surge receded within hours once the center of the storm moved northwards. As seen from the records, the marsh in the Breton Sound basin experienced
surge for about 12 hours. At its peak, the surge depth in the marsh (gages E and I) was about 1 m. At the northernmost gage, G, the surge peaked 14.5 hour later than the southernmost gage, J. As the research team noted during deployment, during this time the water was above the marsh and gradually rising through the vegetation. Comparatively, the normal tidal peak (as measured after November 15) took approximately 8 hours to peak at gage G. The relative lags in time to peak can be seen in Fig. 5.7.

![Fig. 5.7. Comparison of surge during Tropical Storm Ida (left) and normal tide (right) peaks in open water (J) at the southern end and marsh (G) at the northern end of the basin (November 2009).](image)

During the normal tidal cycle, the water propagates northwards only through bayous (primarily Bayou Terre aux Boeufs) and small rivulets and connecting ponds. This is an efficient route for water to propagate, since it is moving through open water bodies devoid of any vegetation. In contrast, during the storm surge event, once the water rises above canal banks and starts propagating as an increasingly deepening sheet flow, it encounters more resistance due to the marsh vegetation, which slows its northward movements. This slower movement, however, cannot be entirely attributed to the vegetation resistance. During this period, winds out of the north must have also caused some resistance. Additionally, decreasing average head differential between the northern and southern ends of the basin must have played some role in slowing down the northward propagation of surge.
Water level data from three USGS monitoring gages (Fig. 5.2) were also available in this period. The stations are USGS No. 07374527 (Northeast Bay Gardene near Point-a-La-Hache, LA), USGS No. 073745257 (Crooked Bayou Northwest of Little Cuatro Caballo near Delacroix) and USGS No. 073745253 (Reggio Canal near Wills Point, LA). The datum for the Crooked Bayou gage could not be confirmed, so that station was not used for comparison. Water level records from the Bay Gardene and Reggio Canal gages are plotted in Fig. 5.8. The plot shows that the surge heights and peak times compare well with records at gages J and G. The USGS gage records also show that a surge of 1.7 m in the bay decreased to 0.7 m as it propagated approximately 40 km northwards through the wetlands.

Fig. 5.8. Surge recorded at the USGS gages during Tropical Storm Ida (November, 2009).

5.4 Analysis of Tropical Storm Lee Surge

Winds during Tropical Storm Lee turned northwards in the study area in the morning hours of September 3, 2011 (Fig. 5.4). The pressure sensors deployed on September 2, 2011 recorded a major portion of the storm surge over the first two days, followed by three more days of tidally forced inundation of the marsh surface (Fig. 5.9). The time series show 15-min averaged data. For further analysis, only the first two days, when the surge was the highest, are considered. The storm surge progressively propagated northwards from Gage S1 to S6, a straight line distance of 6.1 km. On the first day, the recorded peak surge was 1.4 m, NAVD88 at gage
S1 and 1.5 m, NAVD88 at S6. On the second day the peaks were 1.0 m, NAVD88 and 1.1 m, NAVD88 at S1 and S6 respectively.

To examine the effect of marsh and vegetation, the rate of surge rise (RSR) was analyzed. The instantaneous RSR was calculated by dividing consecutive instantaneous water elevation measurements by the time interval between them. The southernmost gage, S1, was the closest to the open bay water and is expected to have the least influence from the marsh and vegetation, so the RSR at S1 is treated as the one without the influence of the marsh and vegetation. The values of RSR at all other gages are then compared with those at S1 in a scatterplot (Fig. 5.10). A linear regression line on each plot indicates deviation of the data from the 1:1 dashed line. If the surge rose at the similar rate at S1 and another gage, then the data points would be closer to the 1:1 dashed line with unit slope. As seen in Fig. 5.10, the slopes of the regression lines decrease from approximately 1 to 0.4 from gage S1 to S6.

![Wind Speed vs Time](image_a)

![Water Level vs Time](image_b)

Fig. 5.9. Wind recorded at WAVCIS CSI-06 and water levels recorded at the surge gages during Tropical Storm Lee (September, 2011).
This indicates that, compared to the rate of rise at S1, the surge rose at a progressively slower rate going northwards to S6. This can be attributed to the increasing resistance offered by the marsh and the vegetation. Thus Fig. 5.10 indicates the spatial differences in the RSR, as the surge encountered greater extent of marsh and vegetation.

As the RSR affects time to peak in the direction of propagation, it is instructive to study the travel times of surge peaks between successive gages. Fig. 5.11 shows a close-up view of the 5-min averaged time series in the vicinity of the highest peak on each day. The highest peaks were selected to estimate the travel times. Table 5.3 lists the time-lag between the peaks at the six gages, on the two days.

Table 5.3: Travel times of peaks between consecutive gages

<table>
<thead>
<tr>
<th>Gage Pairs</th>
<th>Distance (km)</th>
<th>Time of travel of peak on day 1 (min)</th>
<th>Time of travel of peak on day 2 (min)</th>
</tr>
</thead>
<tbody>
<tr>
<td>S1-S2</td>
<td>1.2</td>
<td>10</td>
<td>10</td>
</tr>
<tr>
<td>S2-S3</td>
<td>1.2</td>
<td>0</td>
<td>5</td>
</tr>
<tr>
<td>S3-S4</td>
<td>1.2</td>
<td>10</td>
<td>10</td>
</tr>
<tr>
<td>S4-S5</td>
<td>1.8</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>S5-S6</td>
<td>1.5</td>
<td>5</td>
<td>10</td>
</tr>
</tbody>
</table>
Fig. 5.11. A close-up view of major peak on the (a) first, and (b) the second day. Vertical minor grid is 5 min apart.

The indication from Table 5.3 and Fig 5.11 is that, the travel times between S1 through S4 and from S5-S6 are approximately similar, but the travel time between S4-S5 is almost zero. Moreover, the S4 and S6 time series are almost parallel. This suggests that the surge crest travelled perpendicular along S1-S4 transect, but parallel to the S4-S5 transect. Thereafter, it travelled again perpendicular to the S5-S6 transect. This behavior is a result of refraction of the surge as it moved northeast along the higher topographic features on the east bank of Bayou Little Caillou. Fig. 5.12 shows a schematic of the surge crest inferred from the travel times and the topography.

Fig. 5.12. Refraction of Tropical Storm Lee surge
5.5 Discussion

Wetlands and vegetation affect the storm surge in an estuary by primarily slowing surge propagation, and reducing the surge height through the vegetation-induced drag force. The surge water levels in estuaries may not be considerably reduced by the marsh and vegetation if the wetlands are fragmented and small in size, as seen in the differences in the measured surge height changes in Breton Sound and Terrebonne Bay. Notice that the storm parameters (maximum wind speed, storm size, track and forward speeds) were also different between Tropical Storms Ida and Lee.

The collected dataset of storm surge attenuation caused by wetland vegetation is unique in several aspects. First, although there have been rapid surge measurements during Hurricane Rita (2005) and Hurricane Gustav (2008) by USGS and other institutions, few data have been collected in the longitudinal direction across a marshland. Second, there are significant differences between the vegetation-induced drags under hurricane and tropical storm conditions, because the drag strongly depends on the degree of submergence. Few surge data collected on wetlands under tropical storm conditions exist in the literature. Thus, this field dataset will fill the gap and be used to test storm surge models that incorporate the vegetation effects under moderate surge conditions.

5.6 References


CPRA (2012), Louisiana’s Comprehensive Master Plan for a Sustainable Coast. Coastal Protection and Restoration Authority of Louisiana. Baton Rouge, LA.


CHAPTER 6: WAVE CLIMATE IN A SHALLOW ESTUARY OF A RAPIDLY ERODING COAST

6.1 Introduction

In the world’s major deltaic plains the land-loss has been estimated to be 95 km$^2$/yr over the past 14 years (Coleman et al., 2008). The Mississippi River delta in Louisiana has particularly experienced dramatic wetland loss. Between 1956 and 2006, annual land loss rates ranged from 34 to 104 km$^2$/yr with an average annual land loss rate over that time period was approximately 70 km$^2$/yr (Barras et al., 2003). This loss represents 80% of the coastal wetland loss in the entire continental United States. The public use value of this loss is estimated to be in excess of $37 billion by 2050 (LCWCRTF, 1998).

On the Louisiana coast, the reasons for wetland loss are complex and both natural and anthropogenic (Day et al., 2000; Gagliano, 2003; Morton et al., 2006). One of the important causes of erosion is the constant wave action along the marsh edges. Analysis by Penland et al. (2000) showed that 26% of the wetland loss in the Mississippi river delta from 1932 to 1990 can be attributed to erosion due to wind waves.

Wind waves also influence sediment re-suspension in the nearshore area (Sanford, 1994; Sheremet et al., 2005; Kineke et al., 2006; Jaramillo et al., 2009). Wind waves have been shown to play an important role in the morphological evolution of intertidal regions (Defina et al., 2007; Fagherazzi et al., 2007; Fagherazzi and Wiberg, 2009). Kirby (2000) noted that the shape of the mudshore profile is controlled by tidal currents and particularly by wave climate. Importantly, wind waves degrade salt marsh through scarp erosion (Tonelli et al., 2010). The role of wave attack on coastal marshes is compounded by the conversion of marsh platforms to open-water, thereby increasing the fetch and wave forces on exposed marsh edges.

On the Louisiana-Mississippi coasts, the marshes are typically protected by barrier islands. When the barrier islands disappear, so do the marshes, mainly because of the wave-induced damage and erosion. Studies have found a strong correlation between the level of wave energy and the survival of wetland marshes (e.g., Roland and Douglass, 2005).

The Northern coast of Gulf of Mexico annually experiences tropical storms and hurricanes, and the coastal wetlands provide a natural first line of defense against approaching storm surge and waves (e.g., Lopez, 2009). By one estimate, in the US, the coastal wetlands were estimated to provide $23.2 billion in storm protection services annually (Costanza, 2008).

In this study the characteristics of the wave environment in Terrebonne Bay, a rapidly eroding shallow estuary on the fragile Gulf coast of Louisiana is investigated. Analyzing directional wave gage data collected over a period of 7 months, the magnitude of wave energy and bed shear stresses affecting the bay and the fringing eroding salt marshes is examined. Most Louisiana estuaries are partially sheltered from offshore wave energy by bordering barrier islands. This sheltering effect is one of the main reasons for a barrier island restoration program in the region (CPRA, 2012). The important benefit of barrier islands in mitigating waves in the
back bays has been demonstrated using numerical models (e.g., Stone et al., 2005). However, no
long term field measurements exist to quantify this benefit. With field measurements, the
reduction in swell height is quantified by comparing offshore and bayside measurements. Based
on wave power calculations, marsh retreat rates are estimated and compared with the recent
monitoring data in the area.

6.2 Study Area

Terrebonne Bay is a shallow estuary on the Louisiana coast of Northern Gulf of Mexico
on the west side of the mouth of the Mississippi River (Fig. 6.1). Although part of the abandoned
deltaic lobes, currently the basin receives no major fluvial discharge. The bay is bounded by the
natural levees of Bayou Terrebonne on the east and the Houma Navigation Canal on the west.
Salt marshes line the upper portions of the bay with vegetation communities of smooth cordgrass
(Spartina alterniflora) and saltmarsh meadow (Spartina patens). On the south, the bay is
bordered a series of narrow, low-lying (Elevation 1-2 m MSL, Rosati and Stone, 2009) barrier
islands of the Isles Dernieres and the Timbalier Islands. The wave environment in the bay
comprised of generally locally generated seas but offshore swell do propagate inwards through
the gaps in the barrier island chain. The region has a microtidal environment (tide < 0.5 m) and
depths in the bay vary from 1 to 3 m. Fetch mainly exists in the southeast quadrant and varies
between 10 to 24 km at the measurement site. Every year from October to April about 30 to 40
cold weather fronts pass through the region (Moeller et al., 1993). A typical front lasts from 3-7
days when winds build up from the southerly quadrants and then turn clockwise to strong
northerly winds. The dominant wind directions are southeast and northwest. The region also
experiences tropical storms and hurricanes annually, but none occurred during the data collection
period of the present study.

6.3 Instrumentation, Data and Analysis

Directional data in the bay was collected using an acoustic doppler velocimeter, Sontek
Triton-ADV Wave/ Tide/ Current Gage (ADV). The ADV was deployed at 29°11'13.20"N
90°36'33.59"W, approximately 10 km north of the Timbalier Islands (ADV in Fig. 6.1). Outside
the barrier island chain, at approximately 15 km to the south, wave gage CSI-05 collects hourly
non-directional wave parameters. The system consists of Paroscientific digiquartz pressure
transducer and Campbell CR23X data-logger. The ADV location has a very limited fetch from
northwest to southwest. However, it is directly to the north of Cat Island Pass which provides a
break in the barrier island chain allowing low energy swell to propagate northwards into the bay.
Over the periods from February 23, 2010 through April 29, 2010 and from July 24, 2010 through
February 14, 2011, 17 min bursts were sampled at 4-Hz frequency every 30 minutes to record
puv (pressure, x-component of velocity and y-component of velocity) time series. The gap in the
record from April 30 to July 23 was a result of damage suffered by the ADV from the boat
traffic. The wave records were analyzed using standard spectral methods to produce integral
parameters of zero-moment wave height, $H_{mo}$, and peak period, $T_p$. For the analysis presented in
this paper, only sea and swell records exceeding 0.05 m in wave height were considered as the
bimodal spectral peaks were well defined above these levels. This subset represents about 40%
of the 7 month dataset.
The nearest wind records were available from the meteorological station (Wind monitor model No. 05103, R.M. Young Company) located at the LUMCON Marine Center about 8 km NNW of the ADV site. The wind data was available at 1-min frequency measured at 10 m height.

Fig. 6.1 Study area, bathymetry and locations of monitoring gages.
Majority of the wave spectra measured in the bay showed presence of low frequency swell (Fig. 6.2 bottom panel, reddish brown low frequency bands). To examine the wind wave and swell characteristics, all the bimodal spectra were further partitioned into sea and swell. Starting with the conceptual algorithm of Gerling (1992) several partitioning schemes (e.g., Voorrips et al., 1997, Hanson and Phillips, 2001) have been developed. Various schemes in the literature differ primarily in the strategies to combine peaks in a multimodal spectrum and use arbitrary criteria (Portilla et al., 2009). In the present study, majority of the measured bimodal wave spectra exhibited relatively distinct low and high frequency energy peaks. These were partitioned using the following procedure. First, spurious peaks in the high frequency region were replaced by applying a tail with exponent -4 starting from 1.2*fp (peak frequency). Second, spurious peaks in the low frequency region were ignored by truncating spectrum below frequency 0.05 Hz. Third, the highest two peaks in the spectrum were identified provided that they were separated by at least a frequency difference of 1.2*fp from each other. Finally, the spectrum was split at the lowest point between the two peaks, provided the lowest point was 85% of the smaller peak.

6.4 Wind Wave and Swell Climate

An example of wind and wave field produced by a typical winter front during our study is shown in Fig. 6.2. Spectral wave heights ($H_{mo}$) of smaller than 0.1 m are not plotted, however, corresponding peak periods ($T_p$) are shown to identify and emphasize the presence of swell.

![Fig. 6.2 Measured wave heights and periods at ADV (bay) and CSI-05 (offshore) during the last week of October, 2010. Wave heights ($H_{mo}$) less than 0.1 m not shown but corresponding peak periods ($T_p$) are shown to reveal the low frequency nature. Bottom panel shows energy spectra highlighting bimodal nature of the wave field.](image-url)
At the beginning of the front, when winds are calm (25-Oct), low energy swell enter the bay and sea is negligible. As winds start building up from the south (on 26-Oct), wind waves slowly increase in wave height to around 0.4 m and peak periods between 2.7-2.9 sec. In the subsequent days, although the winds continue to blow from south, the speeds are low (around 5 m/s), resulting in no significant wind waves. The swell however continues to be present throughout. As the winds turn clockwise and start blowing from the north (28-Oct, noon), swell energy subsides. As there is no fetch to the north of the ADV station, no significant waves are produced. The intensity and nature (long or short) of the wave field in the bay can be seen in the statistical distributions of the entire wave height and period data set (Fig. 6.3a). In the case of swell, the average spectral wave height was 0.10 m while average peak period was 6.9 sec. Over the entire data set, the sea wave height average was 0.29 m and the average peak period was 2.7 sec. The wind wave field was primarily generated from the southeast quadrant with northeast being the secondary dominant direction. The swell largely approached from the directions between 150° to 170° (meteorological convention, measured clockwise from North).

Fig. 6.3 (a) Discrete and cumulative probability of observed sea and swell wave heights, (b) probability of observed peak wave periods and (c) probability of observed mean wave directions.

To investigate the protection provided by the barrier islands, we compared swell height measured offshore to that measured at our site, ADV, in the bay. Fig. 6.4 shows the fraction of swell height propagated into the bay for a given incident swell over the entire data collection periods. To represent offshore incident swell, spectral significant wave heights \(H_o\) for which the peak time period was larger than 5 sec \((f_p < 0.2 \text{ Hz})\) were selected from the observations reported at station CSI-05. Fig. 6.4 shows that swell heights reduce to at least 25% at the ADV station. This reduction is the result of processes of diffraction, refraction and dissipation through bottom friction. Barrier islands play an important role of sheltering the inner bay. Note that the data presented is from fixed location, actual swell energy will have spatial variation within the Bay due to wave diffraction and refraction resulting from the varying bathymetry (Fig. 6.1).
The measured wave climate data was used to evaluate the potential for the landward retreat of the marsh edge caused by attacking waves. For this purpose, an empirical expression proposed by Schwimmer (2001) is used. The expression is as follows.

\[
R = 0.35 P_w^{1.1}
\]

(6.1)

where, \( R \) is the shoreline retreat rate (m/yr) and \( P_w \) is the annual cumulative wave power (\( kW/m \)). This expression is based on the field measurements on the northwestern margins of Rehoboth Bay, Delaware. This study area shares several common features with our site such as vegetation type, vertical scarp shorelines and exposed rootmats with underlying mud. Similar expressions have been proposed for some other shorelines (e.g., north shore of Lake Erie, Kamphuis, 1987).

In our study, for each wave record, the wave power was calculated as, \( P_w = E \cdot C_g \), where \( E \) is the wave energy and \( C_g \) is the group velocity based on the peak frequency. Wave power in a given direction was then summed over the entire data set. Note that this cumulative wave power does not cover the entire year but a subset of data (7 months) as explained before. Nevertheless, Eq. (6.1) was used to estimate the erosion rates. Fig. 6.5 shows the estimated potential erosion rates.

Although the wave power estimates are based on measurements about 4 km away from the marsh edge and the possibility of differing retreat rate relation from that proposed by Schwimmer (2001), the estimates indicate grave retreat potential. At three sites in the northern marsh edges of the bay, during the period from 1998 to 2005, the retreat rate averaged 3-6 m/yr (CPRA, 2010) which is on the same order of magnitude of our estimates. In addition to the relentless wind wave action, the marsh edge is also subject to persistent low-energy swell as captured by our data. Studies have shown that the long waves produce strong swash currents resulting in marsh substrate detachment (Priestas and Fagherazzi, 2011).
Fig. 6.5 Cumulative wave power and estimated erosion rate for waves coming from southeast quadrant (meteorological directions).

### 6.5 Wind Sea Growth

Following Young and Verhagen (1996), fetch limited wind wave growth in Terrebonne Bay is examined. Similar to Young and Verhagen (1996), the 2-min wind records were first averaged to produce 10-min averaged records. The raw dataset is first narrowed down to consider only the winds coming from southeast quadrant as it has appreciable fetch compared to the other quadrants. Further, to eliminate potential duration-limited conditions, data points exhibiting directional change of 10° or more or wind speed change of 10% or more were eliminated. For the same reasons, data with significant wave height, $H_{mo}$, less than 0.2 m were ignored. This lower wave height limit was also necessary because for smaller, high frequency waves, the peak frequency tended to be close to the high-frequency noise level cut-off of the spectra.

The data are organized in terms of non-dimensional variables, namely, non-dimensional energy, $\varepsilon = g^2 E / U_{10}^4$, non-dimensional frequency, $\nu = f_p U_{10} / g$, non-dimensional fetch, $\chi = g x / U_{10}^2$ and non-dimensional depth, $\delta = g d / U_{10}^2$ (Bretscherneider, 1958). Fig. 6.6 shows non-dimensional energy, $\varepsilon$, and non-dimensional frequency, $\nu$, in terms of non-dimensional depth, $\delta$, for the entire data set. Solid lines show limits given by the following equations (Young and Verhagen, 1996).

\[
\varepsilon = 1.06 \times 10^{-3} \delta^{1.3} \tag{6.2}
\]

and

\[
\nu = 0.20 \delta^{-0.375} \tag{6.3}
\]

Fig. 6.7 compares observed wave heights and peak periods to those predicted by empirical relations provided by Young and Verhagen (1996) shown below.

\[
\varepsilon = 3.64 \times 10^{-3} \left\{\tanh A_1 \tanh \left[ \frac{B_1}{\tanh A_1} \right] \right\}^{1.74} \tag{6.4}
\]
where

\[ A_1 = 0.493 \delta^{0.75} \]  \hspace{1cm} (6.5)

\[ B_1 = 3.13 \times 10^{-3} \chi^{0.57} \]  \hspace{1cm} (6.6)

and

\[ v = 0.133 \left\{ \tanh A_2 \tanh \left( \frac{B_2}{\tanh A_2} \right) \right\}^{-0.37} \]  \hspace{1cm} (6.7)

where

\[ A_2 = 0.331 \delta^{1.01} \]  \hspace{1cm} (6.8)

\[ B_2 = 5.215 \times 10^{-4} \chi^{0.73} \]  \hspace{1cm} (6.9)

### 6.6 Discussion and Conclusions

Directional wave measurements were carried out inside a rapidly eroding shallow bay partially protected by barrier islands to quantify the intensity and nature of the wave field. In addition to dominant seas, frequently occurring swell energy was observed. For swell, the average wave height was 0.10 m and average peak period was 6.9 sec. As observed from the regional meteorology, the dominant wind direction was 120°-130°. The dominant swell direction was 160°-170° where a gap in the barrier island is present. About 10% of the swell entered from 130°-140° direction where another gap is located. Wind seas during the frontal passages provide the dominant wave energy component in the bay.

These were the first long-term measurements (7 months) inside an estuary of this fragile coast where erosion and land loss has reached catastrophic proportion and threaten commercial, recreational and community well being. Reliable quantification of wave environment is an important piece in understanding physical processes and developing erosion mitigation. For example, knowledge of quantified wave environment is important in sediment deposition on the salt-marshes. The deposition depends on both the availability (created by waves) of suspended sediment and the opportunity (created by wind-induced high water levels and current) for that sediment to be transported over the marsh (Reed, 1989). An example of importance of quantified wave environment for coastal protection projects is found in the northern marshes of our study area. In this area, the Louisiana Coastal Protection and Restoration Authority (CPRA) has invested over one and half million US dollars to evaluate shoreline protection treatment (e.g., gabion mats) and to enhance oyster habitat (CPRA, 2010). Our estimates of marsh retreat rate based on wave power show grave potential. Reliable wave data are critical to the design of such systems in estuaries. To this end, simultaneous long-term measurements (several years) of waves
and shoreline retreat rates are needed to develop reliable empirical expressions such as Eq. (6.1) for Terrebonne Bay.

The presented data can be used to test numerical models of waves in shallow estuaries; a validated numerical model is an important tool to predict waves near wetlands. For coastal engineers and coastal scientists involved in developing wetland protection measures, these results underscore the severity of marsh retreat potential and importance of considering oceanic swell in shallow bays. For coastal ecologists involved in the salt-marsh deterioration and sediment delivery; for estuarine geomorphologists studying intertidal mudflat evolution; for biologists concerned about shellfish colonization and habitats, our results provide the magnitudes of wave energy as an important driving force.

Fig. 6.6 A scatter plot of non-dimensional energy, $\varepsilon$, and non-dimensional depth, $\delta$. Solid line shows Eq. (6.2) and (6.3). Color bar indicates wind direction in degrees.
Fig. 6.7 A scatter plot of observed and predicted wave heights and peak periods. Color bar indicates non-dimensional depth, $\delta$. 
6.7 References


Fagherazzi, S., and P. L. Wiberg (2009), Importance of wind conditions, fetch, and water levels on wave-generated shear stresses in shallow intertidal basins, J. Geophys. Res., 114, F03022.


Hanson, J. L., and O. M. Phillips (2001), Automated analysis of ocean surface directional wave spectra, J. Atm. and Oceanic Tec., 18, 277–293.


Coastal ecosystems are some of the most productive and threatened ecosystems in the world. They include salt and brackish marshes, coral reefs, mangroves, and seagrasses and provide important ecological and economic services. When considering mitigating hurricane impacts, it is generally acknowledged that coastal wetlands provide a natural first line of defense against approaching storm surge and waves. The goal of this research was to examine and quantify the effectiveness of coastal wetland vegetation in reducing storm-induced surge and waves, and the physical sustainability of the wetlands in the presence of waves. The problems have been studied with supporting evidence from field investigations carried out in the unique environment of coastal Louisiana. The research broadly covers two areas. First, the impact of salt marsh vegetation on wave attenuation, wave energy dissipation, probability distribution of wave heights, and storm surge is investigated. Second, the general wave climate in a typically shallow estuary in relation to the erosion potential is studied.

To quantify wave attenuation and wave energy dissipation by vegetation, wave data were measured along a 45 m transect using 4 pressure transducers. The tropical storm force winds produced waves up to 0.4 m (zero-moment) that propagated over vegetation of Spartina alterniflora submerged under a surge of over 1 m above the marsh floor. Largely bimodal spectra consisted of low-frequency swell (7-10 s) and high-frequency (2-4.5 s) wind seas. Measured wave heights, energy losses between gages, and spectral energy dissipation models of rigid vegetation were utilized to estimate wave height decay rates and bulk drag coefficients induced by the vegetation. Measurements showed that incident waves attenuated exponentially over the vegetation. The exponential wave height decay rate decreased as Reynolds number ($R_e$) increased. Larger waves decayed at a slower rate than smaller waves with similar frequencies. The linear spatial wave height reduction rate increased from 1.5% to 4% /m as incident wave height decreased.

The swell was observed to decay at a slower rate than the wind sea, regardless of the wave height. The wind sea energy dissipated largely in the leading section of the transect, but the low-frequency swell propagated along the entire transect, with limited energy loss. The bulk drag coefficient estimated from the field measurement decreased with increasing Reynolds ($R_e$) and Keulegan-Carpenter ($K_C$) numbers. The fitted empirical expression of the form $C_D = 2(a/R_e + b)$ produced coefficients ($a,b$) in the range reported in the literature. Further, the bulk drag coefficients for the longer-period waves were found to be smaller than those for the shorter-period waves, suggesting frequency dependence of the bulk drag coefficient.

The vegetation-induced wave energy dissipation varied across the frequency scales with the largest magnitude observed near the spectral peaks, above which the dissipation gradually decreased. The wind sea energy dissipated largely in the leading section of the instrument array, but the low-frequency swell propagated to the subsequent section with limited energy loss. Across a spectrum, dissipation did not linearly follow incident energy, and the degree of non-linearity varied with the dominant wave frequency.

A rigid-type vegetation model was used to estimate the frequency-dependent bulk drag coefficient. For a given spectrum, this drag coefficient increased gradually up to the peak frequency and remained generally at a stable value at the higher frequencies. This spectral
variation was parameterized by employing a frequency-dependent velocity attenuation parameter inside the vegetation canopy. This parameter had much less variability among incident wave conditions, compared to the variability of the bulk drag coefficient, allowing its standardization into a single, frequency-dependent curve for velocity attenuation inside a canopy. It is demonstrated that the spectral drag coefficient predicts the frequency-dependent energy dissipation with better accuracy than the integral coefficient.

The probability distribution of zero-crossing wave heights was investigated. Wave height distribution was observed to deviate from the Rayleigh distribution. Assuming Rayleigh distributed incident wave heights to the vegetation patch, existing vegetation-induced wave attenuation formulations were employed to derive a special form of two-parameter Weibull distribution. The scale parameter of the distribution is theoretically shown to be a function of the shape parameter, which agrees with the measurements. This effectively makes the distribution a one-parameter Weibull distribution. The derived distribution depends on the local parameters only, and is shown to fit well to the observed distribution of heights of waves dissipating over vegetation. Empirical relationships are developed to estimate the shape parameter from the local wave parameters.

Field measurements showed that the marsh and vegetation affected storm surge in an estuary by slowing the propagation speed, and reducing the surge height attributable to the vegetation-induced drag. Surge water levels in an estuary may not be considerably reduced by the vegetation if the wetlands are fragmented and small in size. Wave energy has been noted as an important factor in salt marsh erosion, but, unlike ocean environments, long-term wave monitoring data typically do not exist for estuarine systems. Using seven months of directional wave measurements spanning all seasons, this study examines the extent of wave energy present in rapidly eroding Terrebonne Bay. Wind seas are the dominant wave energy in the bay. In the northern marshes of the study area, the estimated retreat rates based on wave power calculations are up to 10 m/yr, consistent with the recent land loss monitoring data.

Swell frequently enter Terrebonne Bay through gaps in the natural barrier islands. It was observed that up to 25% of large offshore swell at the measurement site. It is critical to restore and maintain the coastal barrier islands to limit swell-caused erosion in the bays. The presented wind sea and swell data will help in engineering restoration and protection strategies for the vanishing Louisiana coastal salt marshes.

The field data collected during two tropical storms and winter cold front passages in this study is unique because it represents high wave energy conditions, and includes measurements of vegetation properties. More such field investigations, especially under hurricane conditions, are needed to improve the robustness of the proposed relationships and conclusions drawn from this study. Future studies will benefit if the orbital velocity is measured within and above the vegetation canopy concurrently. The measured wave energy spectra in such experiments are typically broad-banded. The impacts of vegetation on the joint distribution of wave heights and periods need to be studied.

The field data collected for this research quantifies wave attenuation by salt marsh during tropical storms for the first time in the scientific and engineering literature, and characterizes the range of attenuation that can be expected in such conditions. The empirical relationships between the estimated vegetation drag coefficient and Keulegan-Carpenter number and the Reynolds
number can be applied in wave modeling of similar salt marsh systems. The theoretical wave height probability distribution function presented in this dissertation can be used to determine characteristic wave heights for the design of coastal defense structures (e.g., levees) fronted by large swaths of salt marsh vegetation. Measurements of storm surge with an array of surge sensors in two estuaries of different size and topography, provide a realistic assessment of surge reduction potential of salt marsh for use by engineers and policy makers with case studies from two storms. More data from storms with different parameters and wetlands with different configurations are needed to capture a larger range of benefits. The seven months of measurements of wave climate in Terrebonne Bay provide evidence on the intensity of normal wave erosion forces on salt marshes. This is valuable information for marsh protection projects in south Louisiana. Simultaneous long-term measurements (several years) of waves and shoreline retreat rates are needed to develop reliable empirical expression relating these two parameters for Terrebonne Bay.
APPENDIX: VEGETATION-INDUCED WAVE ENERGY DISSIPATION MODEL WITH VELOCITY ATTENUATION FACTOR

The derivation presented in this appendix closely follows the procedure in Chen and Zhao (2012), up to the introduction of the velocity attenuation factors.

Wave energy dissipation due to bottom friction is expressed as (Hasselmann and Collins, 1968),

\[
\langle \tau u_k^2 \rangle = -\Theta(k, z) \Delta k
\]  

(A.1)

where, \( \tau \) is the shear stress, \( u_k^2 \) is the velocity of the frequency component with the wave number \( k \) at elevation \( z \), and \( \Theta(k, z) \) is the dissipation function. This equation is used to develop an expression of vegetation-induced wave energy dissipation.

According to the quadratic friction law, the shear stress, \( \tau \), on a vegetation stem of length \( \Delta z \), at elevation \( z \) is expressed as,

\[
\tau = -\frac{1}{2} \rho C_d b_v N_v \Delta z u_{v,x} |u_{v,x}|
\]  

(A.2)

where, \( C_d \) is the drag coefficient, \( b_v \) is the stem diameter, \( N_v \) is the number of vegetation stems per unit square, \( \rho \) is density of water, and \( u_{v,x} \) is the vegetation-affected velocity at elevation \( z \).

It is assumed that the magnitude of vegetation-affected velocity at elevation \( z \) inside the vegetation canopy exhibits a profile similar to that of velocity in the absence of vegetation. Therefore, \( U_{v,v}(z) \), the RMS vegetation-affected velocity inside the canopy at elevation \( z \) can be written as,

\[
U_{v,v}(z) = \sqrt{\frac{2 \sigma^2 \cosh^2 k(h + z)}{\sinh^2 kh} \sigma^2 \Theta_{v,v}(\sigma) d\sigma}
\]  

(A.3)

where, \( \sigma \) is the angular frequency, \( k \) is the wave number, \( h \) is the still water depth and \( \Theta_{v,v} \) is the vegetation-affected energy density spectrum.

Based on Hasselmann and Collins (1968) we can express the dissipation function as,

\[
\Theta(k) = -\frac{1}{2} \rho C_d b_v N_v \Delta z \frac{\rho g^2 k^2 \cosh^2 [k(h + z)]}{\sigma^2 \cosh^2 kh} F(k) U_{v,v}(z)
\]  

(A.4)
In addition, we can define the RMS velocity attenuation coefficient as,

\[ \alpha_{z,r}(z) = \frac{U_{v,r}(z)}{U_r(z)} \]  \hspace{1cm} (A.5)

where \( U_r(z) \) is the RMS velocity in the absence of canopy. Then Eq. (A.4) becomes,

\[ \Theta(k) = -\frac{1}{2} \rho C_d b_v N_v \Delta z \frac{\rho g^2 k^2 \cosh^2[k(h + z)]}{\sigma^2 \cosh^2 kh} F_v(k) \alpha_{z,r}(z) U_r(z) \]  \hspace{1cm} (A.6)

From Eq. (A.1) and Eq. (A.6), the total energy dissipation inside a vegetation canopy at frequency \( \sigma \) and elevation \( z \) can be defined in the following terms,

\[ \langle \tau u_k^z \rangle = -\Theta(k) \Delta k = -\frac{1}{2} \rho C_d b_v N_v \rho g^2 k^2 \cosh^2[k(h + z)] \frac{\alpha_{z,r}(z) U_r(z) F_v(k) \Delta k}{\sigma^2 \cosh^2 kh} \]  \hspace{1cm} (A.7)

Expressed in terms of energy spectrum, \( S_{e,v}(\sigma) \), the total wave energy dissipation rate inside a canopy, at an elevation \( z \), can be written as,

\[ \langle \tau_z u_z \rangle = -\frac{1}{2} \int \frac{C_d b_v N_v \rho g^2 k^2 \cosh^2[k(h + z)]}{\sigma^2 \cosh^2 kh} \alpha_{z,r}(z) U_r(z) S_{e,v}(\sigma) d\sigma \]  \hspace{1cm} (A.8)

Rearranging above equation using the dispersion relation gives,

\[ \langle \tau_z u_z \rangle = -\frac{1}{2} \int \frac{C_d b_v N_v \rho g^2 k^2 \cosh^2[k(h + z)]}{\sinh^2 kh} \frac{\alpha_{z,r}(z) U_r(z) S_{e,v}(\sigma) d\sigma}{\sinh kh} \]  \hspace{1cm} (A.9)

The vegetation-affected velocity spectrum inside a vegetation canopy at elevation \( z \) is given by,

\[ S_{u,v}(\sigma, z) = \left( \frac{\sigma \cosh[k(h + z)]}{\sinh kh} \right)^2 S_{e,v}(\sigma) \]  \hspace{1cm} (A.10)
We can substitute this relationship into Eq. (A.9) to obtain,

\[
\langle \tau_z u_z \rangle = - \int \frac{1}{2} \rho C_d b_v N_v \alpha_{z, \sigma}(z) U_r(z) S_{u, \nu}(\sigma, z) d\sigma
\]  

(A.11)

Furthermore, the frequency-dependent velocity attenuation coefficient can be defined as,

\[
\alpha_{z, \sigma}(\sigma, z) = \left( \frac{S_{u, \nu}(\sigma, z)}{S_u(\sigma, z)} \right)^{1/2}
\]  

(A.12)

where \( S_{u, \nu}(\sigma, z) \) is the velocity spectrum with vegetation and \( S_u(\sigma, z) \) is the velocity spectrum without vegetation at elevation \( z \). Using this definition, Eq. (A.11) becomes,

\[
\langle \tau_z u_z \rangle = - \int \frac{1}{2} \rho C_d b_v N_v \alpha_{z, \sigma}(z) U_r(z) \alpha_z^2(\sigma, z) S_u(\sigma, z) d\sigma
\]  

(A.13)

Assuming \( \alpha_{z, \sigma}(z) \sim \alpha_r \) and \( \alpha_{z, \sigma}(\sigma, z) \sim \alpha_{\sigma}(\sigma) \), i.e., are depth-independent, Eq. (A.13) becomes,

\[
\langle \tau_z u_z \rangle = - \int \frac{1}{2} \rho C_d b_v N_v \alpha_r U_r(z) \alpha_z^2(\sigma) S_u(\sigma, z) d\sigma
\]  

(A.14)

To obtain the total energy loss, integrate along \( z \) from \( -h \) to \( -h + sh \), where \( s \) is the submergence ratio, \( s = h/h_v \),

\[
\sum \langle \tau_z u_z \rangle = - \int \frac{1}{2} \rho C_d b_v N_v \alpha_r \alpha_z^2(\sigma) \left( \int_{-h}^{h+sh} U_r(z) S_u(\sigma, z) dz \right) d\sigma
\]  

(A.15)

Next, we define the frequency-dependent drag coefficient as,

\[
C_D = C_d \alpha_z^2 \alpha_r
\]  

(A.16)
Now, we can derive the expression for the integral drag coefficient.

The velocity spectrum without vegetation is expressed as,

$$S_u(\sigma, z) = \left( \frac{\sigma \cosh[k(h + z)]}{\sinh kh} \right)^2 S_e(\sigma) \quad (A.17)$$

Substituting Eq. (A.17) into Eq. (A.15), and rearranging, the energy dissipation rate of random waves due to vegetation can be expressed as,

$$\langle S_v \rangle = \frac{\Sigma(\tau_z u_z)}{\rho g} = -\int \frac{1}{2g} \rho C_d b_v N_v \alpha_r \frac{\sigma^2}{\sinh^2 kh} S_e \left( \int_{-h}^{h+sh} U_r(z) \cosh^2[k(h + z)] \, dz \right) \, d\sigma \quad (A.18)$$

Note that, Eq. (A.5) can also be written as,

$$\alpha_r^2 = \frac{\int S_{u,v}(\sigma, z) \, d\sigma}{\int S_u(\sigma, z) \, d\sigma} \quad (A.19)$$

We can substitute a term based on Eq. (A.12) for $S_{u,v}(\sigma, z)$ to obtain,

$$\alpha_r^2 = \frac{\int \alpha_r^2 S_u(\sigma, z) \, d\sigma}{\int S_u(\sigma, z) \, d\sigma} \quad (A.20)$$

This expression can be used to eliminate $\alpha_\sigma$ from Eq. (A.15), and we get,

$$\sum \langle \tau_z u_z \rangle = -\int \frac{1}{2} \rho C_d b_v N_v \alpha_r^3 \left( \int_{-h}^{h+sh} U_r(z) S_u(\sigma, z) \, dz \right) \, d\sigma \quad (A.21)$$

In the form of Eq. (A.18), this becomes,

$$\langle S_v \rangle = \frac{\Sigma(\tau_z u_z)}{\rho g} = -\int \frac{1}{2g} \rho C_d b_v N_v \alpha_r^3 \frac{\sigma^2}{\sinh^2 kh} S_e \left( \int_{-h}^{h+sh} U_r(z) \cosh^2[k(h + z)] \, dz \right) \, d\sigma \quad (A.22)$$

Chen and Zhao (2012) use an integral formulation similar to Eq. (A.22), where

$$C_d = C_d \alpha_r^3 \quad (A.23)$$

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Substituting for $C_d$ in Eq. (A.18) based on the Chen and Zhao (2012) formulations (Eq. (A.23)), we get,

$$
\langle S_v \rangle = \frac{\sum (\tau_x u_x)}{\rho g} = - \int \frac{1}{2g} \sigma^2 \frac{\alpha^2_n}{\alpha^2_N} \frac{\sigma^2}{\sinh^2 k h} S_e \left( \int_{-h}^{h+sh} U_r(z) \cosh^2[k(h + z)] dz \right) d\sigma \tag{A.24}
$$

We can define the normalized velocity-attenuation parameter as,

$$
\alpha_n = \frac{\alpha_\sigma}{\alpha_r} \tag{A.25}
$$

Using this parameter, Eq. (A.24) becomes,

$$
\langle S_v \rangle = - \int \frac{1}{2g} \sigma^2 \frac{\alpha^2_n}{\alpha^2_N} \frac{\sigma^2}{\sinh^2 k h} S_e \left( \int_{-h}^{h+sh} U_r(z) \cosh^2[k(h + z)] dz \right) d\sigma \tag{A.26}
$$

The frequency-dependent drag can be expressed as,

$$
C_D = \overline{C_D} \alpha^2_n \tag{A.27}
$$
VITA

Ranjit S. Jadhav grew up in India, where he attended the Indian Institute of Technology (Mumbai), graduating with a bachelor’s degree in civil engineering (1990) and a master’s degree in environmental science and engineering (1992). He, then, moved to the United States in 1992 to pursue a master’s degree in civil engineering at the University of Cincinnati, Cincinnati, Ohio. His thesis research presentation on numerical modeling of flow through wetlands received the Best Masters Student Paper award at the 14th AGU Hydrology Days Conference in Fort Collins, Colorado in 1994. Upon graduation, he joined FTN Associates in Little Rock, Arkansas as a water resources engineer. While in Little Rock, he worked on projects involving numerical modeling of constituent transport and flood hazard. He moved to Baton Rouge, Louisiana in 2001 to lead a branch office of FTN. In Baton Rouge, he has conducted several estuarine modeling studies related to the coastal restoration projects in Louisiana. He is a member of the American Society of Civil Engineers, and Founding Diplomate of the American Association of Water Resources Engineers.