Hydrodynamic and sedimentary responses to two contrasting winter storms on the inner shelf of the northern Gulf of Mexico

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Abstract

Results are presented from the deployment of three bottom-mounted instrumentation systems in water depths of 6–9 m on the sandy inner shelf of Louisiana, USA. The 61-day deployment included nine cold front passages that were associated with large increases in wind speed. Two of the most energetic cold front passages were characterized by distinct meteorological, hydrodynamic, bottom boundary layer, and sedimentary responses and may potentially be treated as end-member types on a continuum of regional cold front passages. Arctic surges (AC storms) have a very weak pre-frontal phase followed by a fairly powerful post-frontal phase, when northeasterly winds dominate. Migrating cyclones (MC storms) are dominated by a strong low-pressure cell and have fairly strong southerly winds prior to the frontal passage, followed by strong northwesterly winds.

On the basis of measurements taken during this study, AC storms are expected to have a lower average significant wave height than MC storms and are dominated by short-period southerly waves subsequent to the frontal passage. Currents are weak and northerly during the pre-frontal phase, but become very strong and southwesterly following the passage. Sediment transport rate during AS storms was not as high as during MC storms, and the mean and overall direction tended to be southwesterly to westerly, with low-frequency flows producing easterly transport, and wind-wave flows producing southeasterly transport.

MC storms had the most energetic waves of any storm type, with peaks in significant wave height occurring during both the pre- and post-frontal phases. The wave field during MC storms tended to be more complex than during AS storms, with an energetic, northerly swell band gradually giving way to a southerly sea band as the post-frontal phase progressed. Currents during MC storms were moderate and northerly during the pre-frontal phase, but became much stronger and southeasterly during the post-frontal phase. Shear velocity was high during both the pre- and post-frontal phases of the storm, although sediment transport was highest following the frontal passage. Mean and overall sediment transport was directed southeasterly during MC storms, with low-frequency and wind-wave flows producing northerly transport. In summary, the data sets presented here are unique and offer insight into the morphosedimentary dynamics of mid-latitude, micro-tidal coasts during extratropical storms.

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1. Introduction

The most common model for atmospheric forcing of inner shelf sediment transport in the mid-latitudes is one in which fair weather wave asymmetry gradually moves sediment onshore, while during storms,
fast wave orbital currents suspend sediment that is then transported offshore by downwelling flows (Niedoroda et al., 1985; Wright et al., 1991; Nittrouer and Wright, 1994). Furthermore, it is commonly assumed that along-shelf transport of suspended sediment during both fair weather and storm conditions is much higher than across-shelf transport, as a result of stronger along-shore flows. Considerable deviation from these general models results, however, from variability in meteorological conditions, local geology, bathymetry, and physical oceanography.

The characteristics of the northern Gulf of Mexico, and Louisiana, in particular, differ from those of most oceanic coasts. Specifically, the region has a much lower overall hydrodynamic energy, a higher frequency wave field, and an east–west coastal orientation, which has important implications with regard to prevailing weather systems. Average significant deep-water wave height and peak period off the Louisiana coast are approximately 1 m and 5–6 s, respectively (Penland et al., 1988; Jaffe et al., 1997), and considerable wave dissipation and refraction occur across the shallow shelf, causing a decrease in wave height closer to shore. Tides in the study area are diurnal, with a tropic range of approximately 0.4 m, resulting in only weak tidal currents (Wright, 1995; Wright et al., 1997). On the other hand, hydrodynamic characteristics during storms tend to be markedly different from those measured during fair weather. During cold front passages, for example, significant wave heights of 2–3 m may occur (Dingler et al., 1993), and water level setup along the coast may reach 0.9 m (Ritchie and Penland, 1988). Not surprisingly, therefore, cold fronts are important meteorological forcing mechanisms in the northern Gulf of Mexico (Crout and Hamiter, 1981; Roberts et al., 1987; Dingler and Reiss, 1990; Murray et al., 1993; Armbruster et al., 1995; Chaney and Stone, 1996; Stone and Wang, 1999). Two extratropical storm types, called arctic surges and migrating cyclones, have commonly been considered to be important in the region (Roberts et al., 1987, 1989). The influence of these two extratropical storm types on hydrodynamics, bottom boundary layer characteristics and sediment transport on the inner shelf of the northern Gulf of Mexico has not been documented in much detail.

Low-energy processes operate the majority of the time in the bottom boundary layer of the Louisiana continental shelf (Wright, 1995; Wright et al., 1997). Field studies conducted on the mid- and outer shelf have indicated that mean near-bottom flows and bed stresses are not strong enough to resuspend sediment during typical conditions (Adams et al., 1987; Halper and McGrail, 1988). Even in depths of 15–20 m, Wright et al. (1997) estimated a mean combined wave–current shear velocity of less than 0.7 cm s\(^{-1}\) and concluded that variations in suspended particulate concentration are generally the result of the advection of sediment plumes from nearby rivers. On the other hand, a few authors have evaluated field data with mathematical models that suggest that bottom stress may be large enough to suspend bottom sediment under certain conditions. For example, Crout and Hamiter (1981) analyzed pressure transducer data from a 10-m-deep location on the inner shelf of western Louisiana using the model of Komar and Miller (1975) and estimated that summer storms, winter cold front passages, and southeasterly wind events during the spring can generate sufficient stress to suspend bottom sediment. Jaffe et al. (1997) predicted sand resuspension on the shoreface adjacent to the Isles Dernieres during a variety of conditions using the Grant–Madsen–Glenn model (Grant and Madsen, 1979; Glenn and Grant, 1987). They concluded that bottom stress is incapable of suspending a significant amount of sediment except during storms and particularly, hurricanes, and suggested that extreme events are probably responsible for the vast majority of long-term sediment transport in the region, even considering their relative infrequency. Finally, Pepper et al. (1998, 1999) and Pepper and Stone (2002) reported transport of fine sand on the Louisiana inner shelf during winter storms.

Although there have been a few relevant studies, bottom boundary layer and sedimentary processes on the Louisiana inner shelf remain poorly quantified. These processes are not particularly well documented on low-energy inner shelves in general, since the majority of field research has taken place on the higher energy shelves of the Atlantic, Pacific, and North Sea. Nevertheless, it is clear that cold front passages may be an important forcing mechanism on low-energy shelves in the mid-latitudes, because they increase hydrodynamic energy and cause sediment transport. The present paper is an attempt to document, on the basis of direct field measurements, inner
shelf hydrodynamic and sedimentary processes during different types of frontal passages on the low-energy Louisiana inner shelf. In doing so, a classification system for cold front passages in the region will be outlined in terms of their sequence of processes and responses and their influence on the Louisiana inner shelf.

2. Methodology

The study area is located on the south-central Louisiana inner shelf, seaward of the Isles Dernieres, in water depths of 6–9 m (Fig. 1). Two deployment sites were chosen so as to occupy both the seaward and landward margins of Ship Shoal, the area’s most prominent bathymetric feature. The coordinates of the seaward location (Site 1) are 28°50.68’ N, 91°07.52’ W, and those of the landward site (Site 2) are 28°55.74’ N, 91°1.73’ W. Wind data were obtained from the Grand Isle C-Man station (GDIL1) located at 29°16.20’ N, 89°57.60’ W. All instrumentation was deployed on November 24, 1998. Three bottom-mounted instrumentation systems were used, two of which (Systems 1A and 1B) were deployed a few meters away from each other at Site 1, while the other (System 2A) was deployed at Site 2. System 2A was retrieved on January 12, 1999, and the others remained at Site 1 until February 2, 1999. Due to memory constraints, however, System 1A ceased logging on January 20, 1999. During each deployment and retrieval, divers collected sediment from the bed and observed and measured any visible bed forms.

The instrumentation consisted of two types of frame-mounted system, both of which included a self-contained data recorder module. The primary components of Systems 1A and 2A were SonTek™ downward-looking Acoustic Doppler Velocimeters (ADVs) that measured seabed elevation, relative particulate concentration and three-dimensional currents at an elevation of approximately 20 cm above the bed. Additionally, both included internal compasses and tilt and roll sensors to enable directional measurements to be rotated into a planetary frame of reference. System 1A was programmed to sample at 25 Hz for 81 s every 3 h. System 2A included a Paroscientific pressure sensor in addition to the ADV and was programmed to sample at 4 Hz for 8.5 min every 3 h.

System 1B was a multi-sensor package called WADMAS. It consisted of a Paroscientific pressure sensor, a sonar altimeter, and a vertical array of three co-located Marsh–McBirney electromagnetic current meters and Seapoint optical backscatter sensors (OBSs). This instrumentation enabled WADMAS to measure water level, directional wave parameters, and seabed elevation, as well as current velocity and suspended sediment concentration at heights of 20, 60, and 100 cm above the seabed. To conserve battery power and recorder memory, all of the sensors on WADMAS were programmed for burst-mode (i.e.,
discontinuous) sampling. Specifically, the sonar altimeter collected one measurement every 15 min, while all other sensors sampled for 8.5 min per hour at a frequency of 4 Hz (Table 1).

All instrumentation was calibrated, prior to deployment, by the Louisiana State University Coastal Studies Institute Field Support Group in their testing facilities. Dry sieving at 0.25 μ intervals was conducted to determine the grain-size composition of the samples of bottom sediment.

Hourly wind data for the deployment period were obtained from the National Oceanic and Atmospheric Administration (NOAA) station located on Grand Isle, Louisiana (GDIL1). These measurements were supplemented by daily national weather maps obtained from the National Weather Service, which were inspected visually to verify the occurrence of frontal passages. “Storm” winds associated with frontal passages were defined as those that exceeded one standard deviation above the mean wind speed for the study period, a value of 7.3 m s⁻¹. Pre-frontal winds were those that blew from a direction between 90° and 270° prior to the cold front passage, according to the weather maps. The post-frontal phase included the period subsequent to the frontal passage when wind direction was between 270° and 90°. It should be noted that the study period occurred during El Niño-Southern Oscillation (ENSO) conditions, which are known to influence mid-latitude Rossby waves (Aguado and Burt, 1999). Therefore, the frequency and characteristics of cold front passages during the study period may not have been typical of “normal” years.

Directional wave parameters were calculated from the pressure and current-meter data using the spectral approach of Earle et al. (1995). Two methods, depending on the instrumentation system, were used to calculate a value of shear velocity above the wave boundary layer (WBL). Values from Systems 1A and 2A, which recorded 3-D currents were computed using the Reynolds Stress technique (RS), while values from the stacked current meters on System 1B were calculated on the basis of the logarithmic profile (LOG) method. Sediment transport was calculated using three techniques, called for the purposes of this paper, (1) the GMR, or Grant–Madsen–Rouse method (Grant and Madsen, 1979, 1986; Rouse, 1937); (2) the MPM, or Meyer-Peter and Muller (1948) method; and (3) the SCP spectral cross-product method (Vincent et al., 1999). The first two of these (GMR and MPM) were based on the concept of shear velocity, while the SCP method was based on instantaneous field measurements of flow and particulate concentration. In this study, the MPM method was employed to calculate bed load transport, while the GMR and SCP methods were used to calculate suspended sediment transport.

The logarithmic profile method was used to calculate shear velocity and apparent bottom roughness from the System 1B data. Shear velocity and apparent bottom-roughness length were calculated for all profiles that had an \( r^2 \) exceeding 0.994 according to log-

<table>
<thead>
<tr>
<th>System</th>
<th>Sensor/measurement</th>
<th>Hours between bursts</th>
<th>Samples/burst</th>
<th>Burst duration (min)</th>
<th>Rate (Hz)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1A (ADV)</td>
<td>pressure</td>
<td>3</td>
<td>2048</td>
<td>8.5</td>
<td>4</td>
</tr>
<tr>
<td>1A (ADV)</td>
<td>3-D current</td>
<td>3</td>
<td>2048</td>
<td>8.5</td>
<td>4</td>
</tr>
<tr>
<td>1A (ADV)</td>
<td>suspended sediment concentration</td>
<td>3</td>
<td>2048</td>
<td>8.5</td>
<td>4</td>
</tr>
<tr>
<td>1A (ADV)</td>
<td>bed level</td>
<td>3</td>
<td>1</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>2A (ADV)</td>
<td>3-D current</td>
<td>3</td>
<td>2048</td>
<td>1.35</td>
<td>25</td>
</tr>
<tr>
<td>2A (ADV)</td>
<td>suspended sediment concentration</td>
<td>3</td>
<td>2048</td>
<td>1.35</td>
<td>25</td>
</tr>
<tr>
<td>2A (ADV)</td>
<td>bed level</td>
<td>3</td>
<td>1</td>
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<tr>
<td>1B (WADMAS)</td>
<td>pressure</td>
<td>1</td>
<td>2048</td>
<td>8.5</td>
<td>4</td>
</tr>
<tr>
<td>1B (WADMAS)</td>
<td>current</td>
<td>1</td>
<td>2048</td>
<td>8.5</td>
<td>4</td>
</tr>
<tr>
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<td>2048</td>
<td>8.5</td>
<td>4</td>
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<td>1B (WADMAS)</td>
<td>sonar altimeter</td>
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<td>1</td>
<td>–</td>
<td>–</td>
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<tr>
<td>GDIL1 (NOAA)</td>
<td>wind</td>
<td>1*</td>
<td>1</td>
<td>10</td>
<td>–</td>
</tr>
</tbody>
</table>

* Sampling schedule shown for the meteorological station refers to GDIL1 data selected for use in this study and not the entire data set collected by NOAA, which was more comprehensive.
linear regression. The von Karman–Prandtl equation was used:

\[
 u(z) = \frac{u_*}{\kappa} \ln \left( \frac{z}{z_0} \right) 
\]

where \( u(z) \) is the horizontal velocity at height \( z \) above the bed, \( \kappa \) is von Karman’s constant (0.4), and \( z_0 \) is the thickness of the roughness layer.

The Reynolds stress, or eddy correlation, technique was used to estimate bottom boundary layer parameters from the ADV data (Systems 1A and 2A). The total horizontal and vertical velocities (\( u \) and \( w \)) were represented as the sum of mean (\( \bar{u} \) or \( \bar{w} \)), periodic (\( u_p \) or \( w_p \)), and turbulent (\( u' \) or \( w' \)) components:

\[
 u = \bar{u} + u_p + u' 
\]

and

\[
 w = \bar{w} + w_p + w' 
\]

which is based on the assumption that turbulent and mean velocities were uncorrelated at all frequencies. The turbulent velocity was isolated by subtracting the periodic (wave-orbital) velocity component from the total-velocity-power spectrum (Green, 1992). To do so, wave orbital velocity was defined as the portion of the velocity spectrum (\( P_{UU} \)) that was coherent with pressure:

\[
 P_{u_w}(f) = \gamma^2 U_p(f) P_{UU}(f) 
\]

where \( P_{u_w} \) is the wave-driven component of the velocity spectrum and \( \gamma^2 U_p \) is the coherence between pressure and velocity (note that the same was done for the vertical, \( w \), component). Obviously, this also has the effect of removing any turbulence that is coherent with pressure, including wave-induced secondary flows. Although such flows were not directly observed during this study, they may have been present at certain times. However, it is assumed that their influence can be disregarded in calculating shear stress and bed roughness, since these parameters are based on diffusive rather than convective processes. Furthermore, there are difficulties inherent in using this method to remove all wave-induced turbulence in the presence of complex wave fields (Kitaigorodskii et al., 1983). Given these caveats, however, and also assuming that field measurements were taken in the constant stress layer, shear velocity is defined as:

\[
 u_* = -\sqrt{\frac{u' w'}{f}} 
\]

The value of shear velocity was then inserted into Eq. (1) to solve for bottom roughness.

The Grant–Madsen (prescribed eddy-viscosity distribution) model (1979, 1986) was used in this study to account for increased shear velocity caused by the combined effects of waves and currents, owing to its widespread familiarity and high level of empirical verification (Larsen et al., 1981; Cacchione et al., 1987; Huntley and Hazen, 1988; Lyne et al., 1990). According to the model, a wave boundary layer (WBL) of thickness (\( \delta_w \)) develops during wave activity within the lower extremity of the current boundary layer, as discussed in the previous section, and the velocity profile is defined separately within and above this layer as:

\[
 u_c = u_{ec} \left( \frac{u_{ec}}{u_{ecw}} \right) \ln \frac{z}{z_0}, \quad z \leq \delta_w 
\]

\[
 u_c = \frac{u_{ec}}{k} \ln \frac{z}{z_c}, \quad z \geq \delta_w 
\]

where \( u_{ec} \) and \( u_{ecw} \) are the current- and combined wave–current-induced shear velocities, \( z_0 \) is the roughness length produced by the sand grains, defined as \( D/30 \), where \( D \) is the mean grain diameter, and \( z_0c \) is the apparent bottom roughness length experienced by the current above the wave boundary layer. Apparent bottom roughness, \( z_{0c} \), is used because the current experiences drag due to the combined influences of physical elements (grain roughness and bed forms) as well as nonlinear interaction with the wave boundary and mobile bed load layers (Grant and Madsen, 1982; Gross et al., 1991). Eq. (1) was used to determine \( u_{ec} \) and \( u_{ecw} \) was calculated iteratively.

Current-induced shear velocity, \( u_{ec} \), was assumed to act in the same direction as the mean current, while the direction of \( u_{ecw} \) was expected to oscillate during the course of the wave cycle. When the wave orbital velocity was at a minimum (near zero), the direction of \( u_{ecw} \) was the same as that of the current; when it was at its maximum, its direction (\( \varphi_{max} \)) was between
the wave and current directions, specified by the equation (modified from Cacchione et al., 1994):

$$\varphi_{\text{max}} = \arctan\left(\frac{\sin \phi}{\cos \phi + \frac{V_0}{u}}\right)$$

(8)

where \(u_b\) is the maximum wave orbital velocity. Obviously, the direction of \(u_{cw}\) has implications for sediment transport within the wave boundary layer, which will be discussed in greater detail in a subsequent section.

Critical shear stress of the seabed sediment was calculated using a modified Yalin technique outlined by Li et al. (1996). The dimensionless Yalin parameter \((\Xi)\) is defined by:

$$\Xi = \left[\left(\rho_s - \rho\right)gD^3/\rho u^2\right]^{0.5}$$

(9)

where \(\rho_s\) and \(\rho\) are the densities of sediment (2.65 g cm\(^{-3}\)) and seawater (1.025 g cm\(^{-3}\)), respectively, \(D\) is grain diameter, and \(\varphi\) is kinematic fluid viscosity (0.013 cm\(^2\) s\(^{-1}\)). The Yalin parameter was used to calculate a critical Shield’s criterion \((\theta_{\text{crit}})\), and \(\tau_{\text{crit}}\) using:

$$\log \theta_{\text{crit}} = 0.041(\log \Xi)^2 - 0.356(\log \Xi) - 0.977$$

(10)

and

$$\tau_{\text{crit}} = \theta_{\text{crit}}(\rho_s - \rho)gD$$

(11)

Critical shear velocity was then simply calculated by \(u_{\text{crit}} = \left(\tau/\rho\right)^{0.5}\). Since sediment at the study site was very uniform fine sand \((D = 0.1 \text{ mm})\) with negligible amounts of silt or clay, it was assumed that a single value could be applied. The equations outlined above showed that the appropriate value for \(u_{\text{crit}}\) was 0.81 cm s\(^{-1}\). An additional parameter derived from critical shear velocity that was employed in this study was the normalized excess shear stress \((S')\):

$$S' = \left(\frac{\tau - \tau_{\text{crit}}}{\tau_{\text{crit}}}\right)$$

(12)

where \(\tau\) is the shear stress calculated from empirical measurements.

The sediment suspension profile over a sandy bottom was shown by Lynch et al. (1997) to be well represented by the standard Rouse (1937) equation, even under combined wave and current flows. This profile is the result of a balance between the upward-diffusive and downward-settling fluxes of sediment. It is represented by:

$$C(z) = C(z_0)\left(\frac{z}{z_0}\right)^{-s}, \text{ where } z = \frac{\gamma w_s}{k u_s}$$

(13)

where \(C(z_0)\) is the reference concentration at height \(z_0\), \(\gamma\) is the ratio of the eddy diffusivity of sediment to that of momentum (~1), and \(w_s\) is the sediment fall velocity. These equations are based on the concept of a reference concentration of sediment near the bed. The concentration \(C(z_0)\) is commonly defined by the equation from Glenn and Grant (1987):

$$C(z_0) = C_{\text{bed}} \frac{\gamma_0 S'}{1 + \gamma_0 S'}$$

(14)

where \(C_{\text{bed}}\) is the sediment concentration in the bed (~ 0.65) and \(\gamma_0\) is a dimensionless empirical constant with a value, according to Hill et al. (1988) and Gross et al. (1991), of approximately \(1.3 \times 10^{-4}\).

Suspended sediment transport is represented mathematically by time- and depth-integrating the product of the horizontal velocity of the fluid and the suspended sediment concentration. As simple as this may seem, it is a very complex problem in combined-flow regimes, owing to phase differences in velocity and concentration, and the possible occurrence of secondary flows including ejected vortices (Agrawal and Aubrey, 1992; Osborne and Greenwood, 1993; Davies and Li, 1997). As a result, the time scale chosen for this integration procedure is of great importance. In fact, Osborne and Vincent (1996) indicated that not only may the magnitude of transport vary on the basis of averaging period, but in some cases, the direction may be completely reversed. On the other hand, the use of instantaneous measurements is problematic, since the time scales of velocity- and suspended-sediment-profile development are different (Davidson et al., 1993). Lesht (1980) and Shauer (1987), for example, recommend scales of several minutes for the establishment of logarithmic velocity profiles. As such, two approaches were employed in this study, the first based on time-averaged values and the second on instantaneous field measurements.

The first technique, which was earlier labeled the GMR approach, was to multiply the burst-averaged velocity and concentration profiles as calculated on
the basis of the shear velocity. This approach has often been employed in wave-dominated environments (e.g., Vincent et al., 1981; Kim et al., 1997) despite the fact that it assumes temporally-uniform values, a condition that may not be satisfied during unsteady oscillatory flow. The profiles were integrated both within and above the WBL using:

\[
Q_{sn} = \frac{1}{t} \int_{z=\delta_w}^{z=z_0} \int_{0}^{t} uC_n dz dt \quad \text{for} \quad z > \delta_w \quad (15)
\]

\[
Q_{sn} = \frac{1}{t} \int_{z=\delta_w}^{z=z_0} \int_{0}^{t} uC_n dz dt \quad \text{for} \quad z < \delta_w \quad (16)
\]

where \( \eta \) is the sea surface elevation.

The cross-product of instantaneous values (i.e., every 0.04 or 0.25 s) of velocity and concentration from Systems 1 and 2A were also used to calculate suspended sediment transport. This had the advantage of accounting for time-varying effects of waves on the sediment suspension and velocity profiles, as well as allowing transport to be analyzed according to frequency components. In many instances during this paper, for example, sediment transport \((Q)\) calculated using this method will be presented on the basis of mean, low-frequency (LOW), wind-wave (wind), and turbulent (turb) components, which represent time periods of \(>81\), \(10.24-80.99\), \(2.34-10.23\), and \(<2.33\) s, respectively. It should be noted that the high-frequency cutoff for the low-frequency component used in the study (i.e., 0.09 Hz) was much lower than has often been used in marine environments, since the peak frequency of waves at the study site was always below this. Total transport \((Q_t)\) is the sum of the individual components:

\[
Q_t = Q_{mean} + Q_{LOW} + Q_{wind} + Q_{turb} \quad (17)
\]

Unfortunately, quantitative assessments of sediment transport made using this method may not have been particularly precise, since it was necessary to assume (very simplistically) that the mean sediment concentration and flow velocity throughout the water column were equal to the burst-averaged values measured at the sensor.

Bed load transport rate \((Q_{bl})\) was calculated by using the combined wave–current shear stress as an input to the empirical formula of Meyer-Peter and Muller (1948) as adapted by Wiberg et al. (1994):

\[
Q_{bl} = 8 \left( \frac{\tau - \tau_{crit}}{\rho_s - \rho} \right)^{3/2} g \quad (18)
\]

The direction of bed load transport under the combined flow of waves and currents is as yet an inadequately resolved issue. Cacchione et al. (1994) assumed that bed load transport would occur in the same direction as that of the maximum shear stress \((\varphi_{max})\) within the WBL. Although this seems to be a somewhat simplistic assumption since the direction of stress may vary up to \(180^\circ\) over the course of a wave cycle, these workers were able to reasonably represent observed trends of bed form migration.

3. Results

A total of nine cold front passages occurred during the deployment, six of which were responsible for large increases in wind speed, hydrodynamic parameters, and sediment transport, while the other three apparently had little influence on the marine environment. Clearly, all cold front passages are unique—even basic characteristics such as duration, synoptic-scale structure, and wind velocity, differ from storm to storm. However, certain regularities were apparently related to the arctic surges and migrating cyclones that occurred during the study period. These storm types will be discussed in the context of two particular storms that occurred during this deployment.

3.1. The AS storm (arctic surge)

The AS storm category is best exemplified by a frontal passage that occurred on January 9, 1999. It was preceded by a low-pressure trough, but was not associated with a nearby low pressure cell. Fairly weak winds (mean of 3.6 m s\(^{-1}\)) blew predominantly from the south within 24 h prior to the passage of the front, following which, the wind velocity became strong (up to 15.1 m s\(^{-1}\)) and north to northeasterly. These winds persisted for approximately 48 h, when the wind again became southerly. Fig. 2A is a vector plot of wind velocity during the frontal passage.
3.1.1. Currents

Mean current velocity was directly related to wind velocity at the study sites, with higher wind speeds causing stronger wind-driven currents. Currents tended to flow in the same direction as the wind, with important modulations, consisting predominantly of a clockwise rotation over time at a period of approximately 24 h, occurring as a result of tidal and inertial effects. It is not surprising, therefore, that aside from the rotational patterns, the trends in current velocity associated with the two storm types were similar to that of wind. Fig. 2B is a vector plot of current velocity for the AS storm. It indicates that currents were fairly weak and variable during the low-energy conditions prior to the frontal passage, while during the subsequent 48 h, currents became strong, steady, and southwesterly.

3.1.2. Waves

The AS storm was characterized by a dramatic increase in significant wave height and an accompanying decrease in peak wave period associated with the onset of northerly post-frontal winds (Fig. 2C). This is further elucidated by Fig. 3, which is a color-coded time series plot of hourly frequency spectra. A fairly narrow band of wave energy at
periods of 5–8 s occurred during the pre-frontal phase of the storm, following which, a conspicuous interval of reduced wave energy occurred nearly coincident with the frontal passage. Subsequent to the passage of the front, the AS storm had a unimodal spectrum, with wave energy concentrated around a period of 4 s.

The directional characteristics of waves associated with various phases of the storm provide additional information regarding their dynamics and generating mechanism. Fig. 4 is a vector plot of non-dimensional wave direction during the storm. Wave direction was from the southeast during the pre-frontal stage, but waves propagated predominantly from the north immediately following the passage of the front. This indicates a shift from swell originating offshore in the Gulf of Mexico during southerly winds, to locally generated storm waves that propagate from the North.

3.1.3. Bottom boundary layer parameters and sediment transport

Fig. 5 is a time series plot of current- and wave–current shear velocity during the AS storm. Peaks in shear velocity occurred during both the pre- and post-frontal phases of this event, separated by a period of distinctly lower values concurrent with the frontal passage. Despite the presence of two peaks, post-frontal shear velocities were considerably higher than they were prior to the frontal passage, with maximum values of nearly twice the magnitude.

Fig. 6 and 7 illustrate across- and along-shore suspended and bed load sediment transport patterns associated with an AS storm as calculated using the GMR and MPM methods for the Site 1 data. Prior to the cold front passage, sediment transport was near zero in both the across- and along-shelf directions, while 3–9 h subsequently, a pronounced peak in
south-southwesterly suspended sediment transport and southerly bed load transport occurred.

Fig. 8 and 9 illustrate predicted across- and along-shelf sediment transport over different frequencies at Site 2 during the AS storm. Mean transport was predominantly onshore prior to the frontal passage as well as during a transport peak immediately following it, while during the majority of the post-frontal phase, it was relatively high and offshore. Mean along-shelf transport was westerly throughout the storm and was particularly high subsequent to the frontal passage. Transport in the low-frequency and wind-wave bands varied considerably over the course of the storm, often reversing direction during the 3-h intervals separating bursts. However, low-frequency transport was most commonly northeasterly, particularly after the frontal passage. Across-shore transport in the wind-wave band, on the other hand, was most commonly onshore prior to the frontal passage, and offshore subsequently. A minor easterly component at wind-wave frequencies also occurred throughout both the pre- and post-frontal phases of the storm.

3.2. The MC storm (migrating cyclone)

The MC storm category was a migrating cyclone and is represented here by a frontal passage that occurred on January 2, 1999. The front was associated with a low-pressure cell that had a central pressure of 100.1 kPa, which tracked across northern Louisiana as the front obliquely crossed the coast. Fig. 10A shows wind velocity vectors for this MC storm. Moderately

![Fig. 5. Current and wave-current shear velocity (u_c and u_cw, respectively) at Site 1 (System 1B) during an AS storm. Trends were similar at Site 2.](image)

![Fig. 6. Across-shelf sediment transport during the AS storm at Site 1 (System 1A) as predicted using the Grant–Madsen–Rouse (GMR) and Meyer-Peter and Muller (MPM) methods.](image)
strong southeasterly winds blew during the pre-frontal phase of this event, followed by a shift to strong northwesterly winds, which became southerly 2–3 days subsequent to the frontal passage.

3.2.1. Currents

Fig. 10B shows current velocity during the selected MC storm. Unlike the previous example, a brief period of strong northerly currents, which reached a maximum speed of 41.3 cm s\(^{-1}\), occurred during the pre-frontal phase of the storm. These currents strengthened further and became southeasterly during the storm’s post-frontal phase, reaching a maximum speed of 53.2 cm s\(^{-1}\). The current direction then began to veer with a period of approximately 24 h. Nonetheless, southerly and southeasterly currents during this time were clearly the strongest, and as a result, southeasterly currents were dominant overall during the post-frontal phase. Interestingly, mean currents were generally stronger than wave orbital flows during this storm, a point that will be discussed further.

3.2.2. Waves

Wave response to the MC storm, as illustrated by Fig. 10C, was more complex than for the AS storm. The time series of significant wave height had two peaks, the lower of which occurred immediately prior to the frontal passage, while the higher occurred just

Fig. 7. Along-shelf sediment transport during a AS Storm, at Site 1 (System 1A) as predicted using the Grant–Madsen–Rouse (GMR) and Meyer-Peter and Muller (MPM) methods.

Fig. 8. Across-shelf suspended sediment transport during the AS storm at mean, low (LF), and wind-wave frequencies, as predicted for Site 2 on the basis of the cross-product of flow and concentration measured by System 2A.
subsequent to it. Maximum hourly significant wave height ($H_s$) during this event was 1.83 m, and $H_s$ exceeded 1.5 m for 10 consecutive hours around the peak of the storm. Peak wave period increased gradually to approximately 8 s prior to the frontal passage, following which, it decreased to 3.76 s. It then fluctuated between the high- and low-frequency values for 24 h, at which point it leveled off at approximately 4 s. As indicated in Fig. 11, the MC storm had a bimodal spectrum subsequent to the frontal passage, with peaks occurring at approximately 8 and 4 s periods. It appears, therefore, that longer period (swell) waves are more prevalent during both the pre- and post-frontal phases of MC storms than they are during AS storms, when post-frontal storm waves dominate and swell waves are less important.

Waves were predominantly from the southeast prior to the frontal passage accompanying the MC storm (Fig. 12). Following the frontal passage, wave direction vacillated between northeasterly and south-easterly before ultimately aligning with the (northerly) wind direction approximately 24 h later. It should be noted that this does not reflect sudden (i.e., hourly) shifts in wave direction, but instead, minor changes in the relative energy levels of the longer and shorter period wave bands. This is indicative of the continued importance of longer period waves throughout the duration of MC storms, accompanied by a significant contribution from locally generated storm waves.

3.2.3. Bottom boundary layer parameters and sediment transport

During the 24 h prior to the frontal passage, shear velocity was much higher for MC storms than for AS storms. The highest shear velocity during the MC storm actually preceded the frontal passage, although an extended period of elevated values persisted during the post-frontal phase. It has already been demonstrated that pre-frontal winds, waves, and currents tended to be more energetic during MC storms than AS storms, and it appears likely, therefore, that this resulted in higher relative pre-frontal shear velocities.

The temporal pattern of directional sediment transport during the MC storm was slightly more complex than during the AS storm, although trends in bed and suspended load transport were remarkably similar (Figs. 13 and 14). Prior to the frontal passage, and persisting until 6 h subsequent to it, sediment transport was offshore, at which time, a sharp onshore peak occurred. This was followed by a prolonged period of offshore transport. Aside from a short duration of westerly transport during the pre-frontal phase, along-shore transport was consistently easterly, with the highest values occurring immediately following the frontal passage.

During the MC storm (Figs. 15 and 16), maximum mean transport coincided with the frontal passage, when it was directed toward the southeast, as was the case for the majority of the storm. The only exceptions were two minor episodes of onshore flux that
occurred 3 h before, and after, the frontal passage. Low-frequency transport was uniformly small during the storm, with southwesterly transport occurring prior to the frontal passage, and south- to northeast- erly transport occurring subsequently. There were two peaks in transport at wind-wave frequencies, both directed toward the northeast, one prior to, and one following, the frontal passage.

It is not surprising that mean transport direction for the storm types predicted using the SCP method was similar to the results calculated using the GMR and MPM models. However, wind-wave and low frequency contributions to transport were highly variable and are difficult to explain, since similar wave characteristics were sometimes associated with widely differing trends in oscillatory transport during the storms.

Fig. 10. Wind velocity (A), current velocity at 100 cm above the bed (B), significant wave height (Hs), and peak wave period (Tp) (C) during the MC storm. The time of the frontal passage is indicated by the vertical line.

Fig. 17 and 18 show the across-shelf component of sediment transport at low and wind-wave frequencies relative to the mean wave direction during the two types of storms, as well as bed level change. The first
point to note is that the sediment transport patterns are surprisingly similar during the different storms. Second, as was the case with geographical transport direction, transport direction relative to waves fluctuated a great deal during the course of the storm. In other words, the highly variable direction of transport at wind-wave and low frequencies was not simply a function of a shift in the direction of the waves, but also of a shift in the transport direction relative to the waves. Generally speaking, low-frequency transport occurred in the same direction as wave propagation during two main peaks, one prior to the frontal passage, and one subsequent to it. During most of the storm’s remainder, low-frequency transport was directed against the waves. The largest peaks in wind-wave transport were aligned with the direction of wave propagation, except during two episodes of transport, one immediately prior to the frontal passage, and the second several hours subsequent to it.

4. Discussion

Wave orbital flows were dominant at both sites during the AS storm; however, comparatively stronger mean currents accompanied the MC storm, particularly at the near-shore site. The situation therefore contrasts both with surf zones, where orbital flows are nearly always dominant, and outer continental shelves, where mean currents are expected to be more important. The near parity between the magnitude of these hydrodynamic mechanisms has clear implications for sediment suspension, which is thought to be closely related to wave orbital flow, and suspended sediment transport, which is strongly influenced by the presence of a mean current (Green et al., 1995). This highlights the uncertainty inherent in the study of sediment transport on the inner continental shelf, which is a transition zone where either mean or fluctuating flow mechanisms may dominate.
The results regarding wind-driven flow are somewhat puzzling since most research, as discussed in the Introduction, indicates that onshore storm winds normally generate coastal setup, which causes downwelling (offshore) mean flows near the bed, while the reverse is true for offshore winds. Clearly, on the basis of mass conservation and an impenetrable coastal boundary, either return bottom flow or spatially variable along-shelf flow is necessary if across-shelf currents are to flow in the same direction for an extended period of time. Inertial currents, which can result when a wind blowing steadily in one direction ceases (Pond and Pickard, 1983), are a possible explanation for the observed behavior. Inertial currents are essentially “remnant” currents that continue to flow despite removal of the forcing mechanism, with their direction and intensity modified by Coriolis force and friction.

Daddio (1977) discussed the influence of inertial currents at a study site in south-central Louisiana. He stated that the location was sufficiently far from the coast (25 km) for the effect of sea surface slope (i.e., setup) to be negligible. Instead, Coriolis-driven inertial currents, which rotated clockwise with a period of approximately 24 h, accompanied frontal passages. This effect was enhanced when sudden removal of onshore wind forcing released sea surface setup. It is possible that the near-bottom currents measured during the present study were at least partially the result of this effect, and not exclusively a product of direct wind forcing. Unfortunately, the lack of on-site wind data precludes more detailed analysis of causal mech-
anisms. Despite this, the sequence of mean flow patterns that accompanied extratropical storm passages was distinctive and has clear implications for inner shelf sediment transport.

A second possibility for the lack of return flow at the bottom relates to the complex bathymetry in the region. Steering of currents into deeper water by Ship Shoal may have modified the expected pattern of downwelling return flow, although the exact nature of this phenomenon remains somewhat unclear. It appears likely that both inertial currents and complex bathymetry have a combined influence on flow velocity in the vicinity of the shoal during winter storms.

Wave energy, as indicated both by wave height and near-bed orbital velocity, was higher during MC storms (1.06 m and 0.19 m s$^{-1}$, respectively, at Site 1) than AS storms (0.8 m and 0.12 m s$^{-1}$). Mean peak wave period was considerably higher for MC storms (5.8 s) than for Storm AS (4.1 s) at both sites. It appears, therefore, that MC storms were more important in terms of wave generation.

Clearly, the storm types differed in terms of their associated wave characteristics. The MC storm had the most energetic wave field, particularly during the pre-frontal phase. Significant contributions to the energy spectrum resulted from both long-period northerly swell waves, and short-period, southerly storm waves, both of which were present during the majority of the post-frontal phase. The AS storm, on the other hand, was dominated by short-period southerly waves subsequent to the frontal passage. Unlike the MC storm, which was often characterized by

![Graph 1](image1.png)

**Fig. 15.** Across-shelf suspended sediment transport during the MC storm at mean, low (LF), and wind-wave frequencies, as predicted for Site 2 on the basis of the co-spectrum of flow and concentration measured by System 2A.

![Graph 2](image2.png)

**Fig. 16.** Along-shelf suspended sediment transport during the MC storm, at mean, low (LF), and wind-wave frequencies, as predicted for Site 2 on the basis of the co-spectrum of flow and concentration measured by System 2A.
complex, bimodal spectra, wave spectra during AS storms were essentially unimodal. These general tendencies appear to be the result of the relative influence of northerly and southerly winds during different storm types. As outlined previously, AS storms were characterized by weak southerly pre-frontal winds and energetic northerly pre-frontal winds, thus resulting in higher southerly post-frontal storm waves. In contrast, the strong southerly pre-frontal winds that accompanied MC storms had a longer fetch over which to act than did the northerly post-frontal winds, resulting in the generation of long-period southerly waves that were energetic enough to persist throughout much of the storm. This also explains the relative reduction in significant wave height and peak wave period at Site 2 during MC storms, since southerly, and not northerly, waves are primarily influenced by attenuation across Ship Shoal.

Sediment transport predictions, much like hydrodynamic and bottom boundary layer parameters, vary considerably, although somewhat regularly, based on the type of storm driving the response. As has been stressed repeatedly during this paper, as in most relevant literature, considerable uncertainty exists in estimating sediment transport rates, and as such, specific quantities are presented chiefly for the purposes of comparison. It is clear, first of all, that different methods of calculating sediment transport rate yielded different results in terms of magnitude, and at certain times, transport direction, although the overall trends were largely similar.

Overall mean sediment transport calculated using the GMR and MPM methods was southwesterly and southeasterly for the AS and MC storms, respectively, with a considerably higher overall magnitude being associated with the MC storm. It should be stressed
again, however, that this offshore transport following frontal passages does not appear to be the result of downwelling flows caused by coastal setup, but instead, of wind-driven mean currents modified by inertial effects. In the case of the MC storm, pulses of onshore transport occurred both before and after the offshore transport peak. This appears to have been caused by mean currents that rotated with a period of 24 h during the storm as a result of both tidal, and inertial, modulation. Further study is required to determine whether the difference in sediment transport magnitude was representative of the storm types, or whether it may have been a function of the individual storms themselves.

The causes of the drastic directional shifts at wind-wave and low frequencies cannot be positively identified from the data set since the possible factors responsible for generating the necessary phase differences are numerous and include wave asymmetry, vortex generation by bed forms, sensor location relative to the bed (and individual bed forms), and the interaction of various wave trains. Wave characteristics alone did not seem to account for these differences, and as such, the most likely explanation must incorporate changes in the bed beneath the instrumentation. As shown in Figs. 17 and 18, large changes in bed level accompanied directional shifts in sediment transport during storms. It is likely that as bed level varied, the lag time of sediment reaching the sensor following its suspension by oscillatory flow was altered, as was demonstrated by Vincent et al. (1991), Osborne and Greenwood (1993), and Osborne and Vincent (1996). In addition, the influence of wave vortices may have changed depending on the size, shape and spacing of bed forms; Davies (1985), for example, showed that vortex shedding occurs only when wave orbital excursion length exceeds the spacing of bed ripples. Finally, the position of the sensor in relation to bed forms is important (Osborne and Vincent, 1996; Vincent et al., 1999). These explanations remain a matter of speculation, since bed observations are not available; however, it should certainly be investigated in the future. Despite the intriguing nature of the problem of oscillatory flux, the fact remains that it accounted for a much smaller portion of overall suspended sediment transport than mean flux. Clearly, therefore, several issues with respect to directional sediment transport should be addressed during future studies, and in particular, the role of morphological changes of the bed in causing sediment suspension.

5. Conclusions

Extratropical storms are the most energetic events influencing the Louisiana inner shelf during the winter. Two storm types and their associated marine responses were identified during this study:

1. AS storms are arctic surges that are associated with anticyclonic activity, generating northeasterly winds, southerly storm waves, and southwesterly currents and sediment transport.
2. MC storms are migrating cyclones that generate energetic, rotational, pre- and post-frontal winds and currents and high northerly swell that transforms into southerly sea. Overall, currents and sediment transport are southeasterly.

Clearly, the response of the marine environment to such storms is crucial in determining the long-term sediment-transport pathways and morphological response of the shelf, and as such, a longer term climatological evaluation is currently under way, and additional instrumented deployments are recommended.

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