1983

Wind-Driven, Near-Bottom Currents Over the West Louisiana Inner Continental Shelf.

Richard Larry Crout
Louisiana State University and Agricultural & Mechanical College

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WIND-DRIVEN, NEAR-BOTTOM CURRENTS OVER THE WEST LOUISIANA INNER CONTINENTAL SHELF

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 Marine Sciences

by

Richard Larry Crout
B.S., University of South Carolina, 1976
M.S., Louisiana State University, 1978
May 1983
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ABSTRACT

Forcing mechanisms and water column response over the West Louisiana Inner Continental Shelf (WLICS) are investigated at various temporal and spatial scales. The major mechanisms that have an effect on shallow-water currents over the WLICS are winds, runoff from the Mississippi and Atchafalaya rivers, and circulation of the northwestern Gulf of Mexico.

Summer wind stress values are low and currents generally meander over the inner shelf during this period. A shift in the direction of the regional wind stress in late June causes a reversal in the current direction over the west Louisiana shelf. Autumn currents are primarily westward in response to predominantly westward wind stress. Winter currents are primarily westward, but during frontal passages the current swings rapidly to the east. As winds become easterly, the currents return to their westward set. During the spring the flood on the Mississippi and Atchafalaya rivers introduces a large amount of fresh water into the coastal waters. The fresh water causes density gradients, which decouple near-bottom flow from surface winds and increase the intensity of the westward currents. During the autumn and winter regional-scale cyclonic flow, generated by a succession of cold-front passages, helps to maintain flow over the WLICS in a westward direction.

Four periods of frontal passage in 1979 are studied in detail. Strong alongshelf wind stress is common during the January and March frontal periods. An investigation of water column dynamics reveals that the alongshelf wind stress accelerates the water column and generates a frictional boundary layer at the bottom.
A strong high pressure system pushes a front across the WLICS from the north during February. Dynamically, low alongshelf wind stress accelerates the water column, but it is not strong enough to establish a frictional balance. An alongshelf pressure gradient also influences the water column dynamics.

South-southeasterly winds and freshwater runoff dominate the April frontal period. The alongshelf wind stress accelerates the water column, but no frictional balance is established. A pressure gradient term due to the Atchafalaya flood contributes to the momentum balance.
CHAPTER 1

INTRODUCTION

The West Louisiana Inner Continental Shelf (WLICS), which includes the area where my investigation is focused, is located in the northwestern Gulf of Mexico (Fig. 1.1). Bottom contours parallel the relatively smooth coastline of west Louisiana, which trends east-west. The broad, shallow WLICS reaches depths of 50 meters at a distance of 150 kilometers from shore.

The gentle slope of the WLICS is interrupted only by the Sabine Bank, which rises within 8 meters of the sea surface approximately 25 kilometers from shore. Sabine Bank, a discontinuous ridge extending some 80 kilometers roughly parallel to shore, contains reworked sands, which are among the coarsest material on the Louisiana shelf west of the Mississippi Delta. A complex pattern of sand, mud, and shell material covers the WLICS (Curray, 1960). Adams et al. (1982) suggest that alternating winds accompanying cold-front passages induce currents that preferentially move sand and mud materials in opposite directions along the central Louisiana shelf, helping to maintain the integrity of sand banks. Much of the fine-grained materials constantly supplied by the Mississippi and Atchafalaya rivers remains in suspension and is carried westward by the prevailing nearshore and shelf currents. Some of the fine-grained sediment settles out into migrating mudbanks near shore (Wells and Kemp, 1981).
Figure 1.1. The Gulf of Mexico showing the location of the West Louisiana Inner Continental Shelf (WLICS).
The dynamics of the water column over the WLICS are poorly understood (Murray, 1976), and the void must be filled in order to predict transport through the region. Not only are sediments from the Mississippi and Atchafalaya rivers present over the Louisiana shelf, but nutrients and contaminants from a large area of the interior of the United States flow through this river system and subsequently along the Louisiana coast. The State of Louisiana is presently interested in the erosion and deposition of coastal lands. The materials involved in these processes are transported over the inner shelf, and thus this study acquires added significance.

The dominant mechanisms forcing circulation in coastal waters are wind and tide (Winant, 1980). In a microtidal environment (tidal range, 25-75 centimeters) such as the Louisiana shelf, the importance of tide in driving currents is limited. Strong local winds, though, transfer momentum rapidly into the shallow coastal waters over the WLICS. Thus, wind emerges as the major source of energy for driving inner shelf currents.

Another unique feature of this region is the massive freshwater input from the Mississippi and Atchafalaya rivers. The hydrography of the entire Louisiana shelf, and the Texas shelf at least as far west as Galveston, is influenced by their flood (Angelovic, 1977). They also help to maintain a narrow brackish zone along the west Louisiana coast during much of the year. Subtropical swamps and marshes of south Louisiana drain into numerous coastal rivers and bays, which also contribute a considerable amount of fresh water to the inner shelf during the fall and winter, when rainfall is high and Mississippi and Atchafalaya discharge is low. The large-scale circulation of the Gulf
of Mexico, which responds to large-scale wind stress variations, may also influence shelf currents in the northwestern Gulf.

While the principal concern of this dissertation is with circulation over the inner shelf, this shallow circulation is not independent of the large-scale motions and forcing of the northwestern Gulf. Therefore, in the next chapter, I describe our existing knowledge and ideas concerning the regional scale, low-frequency patterns of variability in wind velocity, freshwater runoff, and circulation of the western Gulf of Mexico and its northern shelf.

In Chapter 3, data from the WLICS is subjected to spectral analysis. The primary forcing mechanism for inner-shelf circulation is the alongshelf wind stress. Fluctuations of current speed and direction are the responses. The seasonal and synoptic variability of both of these parameters are described, as well as the variations of the statistical relationships between parameters on the same time scales.

An investigation of the inner shelf response to forcing by selected winter cold-front passages concludes the dissertation. These synoptic forcing events impart a large amount of energy to the shallow water. Response of the water column to the strong wind stress accompanying cold fronts, though, is highly variable.
CHAPTER 2

REGIONAL, SUBTIDAL VARIABILITY -
A HISTORICAL REVIEW

Meteorology

The yearly cycle of winds in the Gulf of Mexico is linked closely to the Bermuda High Pressure System and its seasonal migration in the North Atlantic Ocean (Leipper, 1954). During spring, isobars associated with the Bermuda High are aligned over the Gulf in a southeast-northwest orientation, resulting in predominantly southeasterly winds. Later in the summer, the high-pressure system reaches its southernmost point and southeasterlies prevail over most of the Gulf of Mexico; the exception is the northwestern Gulf, where winds are primarily from the southern quadrant. A low-pressure system that forms over Mexico may also contribute to the slow, southerly winds observed over the Louisiana shelf during the summer period.

Murray (1976) suggests that, since the southerly winds cause current reversals along the Texas coast, reversals over the central Louisiana shelf reported by Kimsey and Temple (1963) and Oetking et al. (1973) may be a response to the reversal of currents over the Texas shelf. Autumn wind conditions are due to the rapid northern migration of the Bermuda High, which leaves the Gulf under the influence of the northeasterly trade winds. During winter, winds associated with the Bermuda High would be primarily from the east. Instead, observed winds over the northwestern shelf are variable from the southeast, counterclockwise through north (Wiseman et al., 1976).
These variable winds are due to mid-latitude waves that form when cold, dry polar air masses move equatorward into regions of warm, moist tropical air. Fronts are formed at the intersection of the two air masses, and they translate eastward across North America. Lows are generated along the front, creating a region where wind direction shifts rapidly. These fronts move across the northern Gulf Coast at an angle of approximately 45° to the coastline (Fernandas-Partegas and Mooers, 1975). The WLICS is located within a trough where the maximum number of these winter fronts occurs (DiMego et al., 1976).

Winds prior to the passage of a frontal system are from the south and southeast; and, after a brief period of light and variable winds as the front passes, the winds shift to blow from the north at speeds that regularly exceed 10 m sec\(^{-1}\) in winter (Angelovic, 1977). The major component of winds accompanying a front is normal to the Louisiana coast. Alongshore winds usually drive coastal circulation, but what occurs when the primary winds are perpendicular to the coast?

Fronts begin to form in the autumn and continue to migrate through the region into the spring. During winter, as many as nine well-developed fronts per month cross the west Louisiana coast (DiMego et al., 1976). The fronts are therefore short in duration, and their recurrence interval may be as short as three days. Frontal passages with accompanying high winds and rapid wind shifts, coupled with the wide, shallow expanse of inner shelf water, exert a significant influence on coastal currents in winter (Daddio et al., 1978; Wiseman et al., 1978).

**Freshwater Runoff**

Runoff from the Mississippi and Atchafalaya rivers contributes almost 65 percent of all freshwater input into the Gulf of Mexico.
These rivers discharge into the Gulf on a yearly flood cycle (Fig. 2.1). Although inter-annual variability occurs, spring floods usually reach a maximum in April and are followed by a rapid decrease in river discharge. During the flood season, prevalent southeasterly winds transport the light surface water westward over the Louisiana shelf, lowering salinities in coastal waters as far west as Galveston (Angelovic, 1977).

The influence of the flood cycle on coastal salinities is illustrated in Fig. 2.2. In each of the three years presented, the spring flood is followed by a significant decrease in salinity at a station 8 kilometers offshore of Calcasieu Pass and a lesser, but appreciable, change in salinity 45 kilometers from shore. Changes in salinities on the shelf induced by the influx of fresh water remain over a period of months and will be discussed in a later section.

Gagliano et al. (1970) cite discharge from local rivers as an important factor for maintaining the narrow brackish band that is present along the WLICS during nonflood periods. It is evident that the local rainfall is highest during nonflood periods of the Mississippi and Atchafalaya rivers (Fig. 2.1), and therefore local rivers may significantly influence the salinity distribution along the west Louisiana coast during these periods of the year.

**Deepwater Circulation**

Deepwater circulation may influence nearshore circulation in one of two ways. Over narrow shelves deepwater currents move to within several kilometers of shore and overwhelm locally driven coastal circulation. On broad, shallow shelves like those found on the eastern and Gulf coasts of the United States, the deep currents do not move
Figure 2.1. Monthly mean discharge from the Mississippi and Atchafalaya rivers and monthly mean rainfall at Lake Charles, Louisiana (Angelovic, 1977).
Figure 2.2. Monthly discharge rates of all Louisiana rivers and surface salinity at two stations off Calcasieu Pass, 1963-1965 (from Temple et al., 1978).
near shore. In the bight between South Carolina and Florida (Blanton, 1981) and in the Middle Atlantic Bight (Beardsley and Winant, 1979), coastal currents are influenced by pressure gradients induced by the Gulf Stream.

The Gulf of Mexico may be divided into two regions with respect to large-scale circulation. An imaginary line connecting the Mississippi River delta to the northeastern tip of the Yucatan Peninsula serves as the boundary between the eastern and western Gulf of Mexico. The eastern Gulf is dominated by the Gulf Loop Current. Large detached rings separate from the Loop Current and drift into the western Gulf of Mexico, where they may influence circulation (Schroeder et al., 1974). Patterns in the western Gulf indicate a more complex circulation than in the eastern Gulf, although a large anticyclonic gyre is the dominant feature during most of the year.

Winter circulation in the western Gulf of Mexico is more predictable than the Loop Current, according to Nowlin (1972). The primary circulation consists of an anticyclonic gyre over the west-central Gulf oriented along a southwest-northeast axis (Fig. 2.3). This gyre is characterized by broad westward flow along its southern limb and narrow east-northeastward flow within the northern flank, which is supplemented by flow from off the Texas shelf (Molinari et al., 1978). Evidence indicates that the anticyclonic gyre is present throughout the year (Leipper, 1970; Merrell and Morrison, 1981; Schroeder et al., 1974), but that transports are highest during winter and summer. Sturges and Blaha (1976) and Blaha and Sturges (1981) attribute this seasonality to variations in the curl of the wind stress, which is also at a maximum in winter and summer.
Figure 2.3. Dynamic topography of the sea surface of the Gulf of Mexico relative to the 1,000-db surface from data averaged over all seasons (from Nowlin, 1972).
When the Loop Current is fully extended into the eastern Gulf, it is bounded on the west by a low-pressure trough. Elliott (1979) suggests that the trough moves into the western Gulf in conjunction with the westward migration of an anticyclonic ring. Both Elliott (1979) and Merrell and Morrison (1981) link the low-pressure trough with cyclonic circulation over the Texas-Louisiana shelf. The aperiodicity of the Loop Current seems to argue against this theory. Local hydrography over the shelf, however, may influence shelf circulation (see below).

**Shelf Hydrography**

Low-salinity water is present over the inner continental shelves of Louisiana and Texas year-round. Abundant rainfall over south Louisiana in early fall and winter and the spring flood waters of the Mississippi and Atchafalaya rivers supply a sufficient amount of runoff to maintain this brackish nearshore zone. Reduced-salinity water is present in large volumes as far west as Galveston in spring owing to the large influx of river water and its subsequent transport westward. Changes in shelf hydrography are greatest during the spring flood, but in late summer, following cessation of the flood, currents advect some of the water eastward and a second salinity decrease in east Texas coastal waters is reported (Smith, 1980; Lewis, 1980).

These changes in hydrography on the Texas-Louisiana shelf are readily apparent in Figure 2.4, which shows data collected by the National Marine Fisheries Service (Temple et al., 1978). Vertical profiles of temperature indicate that the water column at the innermost station, in 7 meters of water off Calcasieu Pass, is nearly isothermal throughout much of the year, although a range of from 13° to 35°C is encoun-
SURFACE SALINITY

Figure 2.4. Surface salinity in parts per thousand for six periods showing the influence of the Atchafalaya and Mississippi rivers flood over the Texas-Louisiana shelf during 1963 (data source: Temple et al., 1978).
tered. The greatest variance from isothermal conditions occurs during summer. As the depth and distance from shore increase, the surface-to-bottom temperature difference intensifies. At the deepest stations, the near-bottom maximum temperature occurs after surface temperatures begin to decline. Surface temperatures follow the atmospheric temperature trends, as expected, attaining a maximum in summer and a minimum in winter.

Salinities at the inner station also vary seasonally, ranging from 16.5 to 33.7 ppt, but the water column appears to be essentially vertically isohaline during autumn and winter. The seasonal salinity variability decreases in the offshore direction; and, over the outer shelf, salinity changes occur only in surface waters. Seasonal variations of salinity in inner and mid-shelf waters off Louisiana are rapidly influenced by discharge from the Mississippi and Atchafalaya rivers, whereas Texas shelf waters experience a lag of approximately one month in their response to the discharge from Louisiana sources.

An important conclusion may be inferred from the data presented by Temple et al. (1978). Plots of surface and bottom temperatures and salinities with time indicate that the water column is very nearly vertically isothermal and isohaline during late autumn and winter, particularly at the inner-shelf stations. The quasi-homogeneity of the water column allows surface values of salinity and temperature to be used to represent the entire water column during these seasons.

The normal winter shelf salinity distribution (Fig. 2.4a) shows isohalines paralleling the isobaths and salinity increasing by about 6 to 7 ppt from the coast out to the mid-shelf region. The salinity of the outer shelf varies little. Freshening of nearshore waters, increasing from west to east, is discernible as the flood begins in
April (Fig. 2.4b). The freshwater discharge is transported westward over the Texas shelf and south to Galveston as the season progresses (Fig. 2.4c). The mid-shelf water south of Timbalier Island remains at open Gulf salinities of greater than 36 ppt. Return of fresh water previously advected westward is suggested by the shape of the isohalines in Figure 2.4d. Conditions over the shelf seaward of Timbalier Island during August indicate that a large amount of fresh water has been dispersed into the offshore region (Fig. 2.4e). A return to winter conditions, characterized by isohalines parallel to shore and generally higher salinities, is observed by September (Fig. 2.4f). The salinity structure of the shelf in the northwest Gulf remains steady during the winter. Minor variations in this annual salinity cycle may occur, but the effect of fresh water from Louisiana sources on shelf waters and the significance of eastward summer currents are clearly important.

While salinity is the dominant parameter controlling density near the mouths of the Mississippi and Atchafalaya rivers, it interacts with temperature to an increasing degree as one moves westward over the shelf and offshore. Plots of surface sigma-t illustrate how the density of coastal waters is influenced by temperature. Cooling of the shallow water begins in late autumn as cold fronts cross the northwestern Gulf coast and intrude into the Gulf of Mexico. Nowlin and Parker (1974) chronicle the effect of a single strong frontal passage on Texas shelf waters. Temperatures decreased by as much as 4°C in the vertically isothermal waters. The persistent procession of winter cold fronts alters the hydrography of the shelf waters in a step-wise fashion, each cooling the waters further and eventually creating an alignment of isopycnals that suggests a strong westward flow over the inner shelf (Nowlin, 1972).
The density structure of the shelf undergoes a seasonal cycle as suggested by Cochrane (Hann and Randall, 1981), which is illustrated in Figure 2.5. October shows a fall pattern of isopycnals roughly paralleling the isobaths (Fig. 2.5a). The baroclinic pressure gradient implied by the alignment of the isopycnals suggests westward flow over the shelf. Data from December reveals the presence of a region of high-density water over the mid-shelf (Fig. 2.5b). The region of high-density water is attributed to the successive cooling of the shallow water by cold fronts traversing the shelf (Huh et al., 1978). During autumn, the cold fronts lose much of their cooling ability over the inner shelf because of the large temperature difference between air and water. As the frontal cold-air mass moves over warm inner-shelf waters, large amounts of sensible and latent heat are exchanged, resulting in a cooler water column and a much warmer air mass. Mid-shelf waters are more saline than nearshore waters, and they remain denser because of that fact. Although salinities over the mid-shelf region and the outer shelf are nearly the same, the mid-shelf waters are cooled more by frontal passages and, as a result, are more dense. An increase in density of all shelf waters accompanies the rapid passage of cold fronts that occur during January (Fig. 2.5c), as both heat and mass are stripped from the water column. Figures 2.5d and 2.5e show a fully developed cyclonic cell centered over the Texas-Louisiana middle continental shelf in February and March.

If the baroclinic density gradients over the shelf are in geostrophic balance, flow around the denser mid-shelf water implies westward baroclinic coastal currents and weaker eastward flow over the outer shelf. Csanady (1979) calculates baroclinic currents from density
Figure 2.5. Surface sigma-t values during 1963-1964 showing shelf water variability and the formation of a cyclonic gyre on the Texas-Louisiana shelf (data source: Temple et al., 1978).
sections over the New England continental shelf. A similar technique is applied to the January 1965 density transect seaward of Calcasieu Pass (Fig. 2.6). Computed baroclinic currents over the inner shelf are westward. The water column nearshore flows at a vertically averaged speed of 9.0 cm sec\(^{-1}\), as compared to 2.0 cm sec\(^{-1}\) somewhat farther offshore. Currents over the outer half of the continental shelf are eastward at 1.5, 6.0, and 3.5 cm sec\(^{-1}\). This information suggests a narrow ribbon of water moving rapidly westward over the inner shelf and a broad eastward flow over the outer region of the shelf. Such a baroclinic component of the current must be considered when winter flow regimes are discussed.

The weakening of the shelf cyclone and its westward translation occur in early spring as flood waters from the Mississippi and Atchafalaya move onto the shelf (Fig. 2.5f). The density structure is dominated by freshwater runoff until autumn, when the cycle begins again. The baroclinic current is expected to be important during the period when large amounts of freshwater are present on the shelf.

Two recent summer transects (Fig. 2.7a) from the WLICS show the hydrography of the coastal boundary layer during two isolated events. Cross sections of salinity along the two transects mimic the summer density structure (Figs. 2.7b and 2.9d). The early summer is characterized by lower salinity conditions that extend far offshore (Fig 2.7b), apparently because of the influence of the discharge of the Atchafalaya River and the nearby Calcasieu River. Salinities increase substantially by late summer (Fig. 2.7d). Comparison of the salinity and observed current cross sections (Fig. 2.7) emphasizes the importance of the salinity structure in the coastal boundary layer. Currents
Figure 2.6. Transect perpendicular to Calcasieu Pass showing vertical sigma-t structure of the water column and inferred vertically-averaged baroclinic currents with respect to 110 db. Eastward flow is positive (data source: Temple et al., 1978). Squares represent data points.
Figure 2.7. The coastal boundary layer along the WLICS is shown using: (a) a chart showing the location of two summer transects across the WLICS; (b) sigma-t structure during June 1980 transect; (c) velocity structure in centimeters per second during June 1980 transect showing alongshelf currents (eastward is positive); (d) sigma-t structure during September 1980 transect; (e) velocity structure in centimeters per second during September 1980 transect (eastward is positive).
during the early-summer study were westward above the halocline and eastward below it (Fig. 2.7c). These regions of opposing westward and eastward flow were separated by a halocline in late summer also (Fig. 2.7e). Scott and Csanady (1976) detected a similar flow regime south of Long Island. A CBL with two-layered alongshelf flow is common over the WLICS only when fresh water is present in large amounts and wind speeds are low. Stronger winds during the remainder of the year, in combination with a lesser volume of fresh water over the shelf, result in near-homogeneous conditions over the west Louisiana shelf.

The structure of the currents within the west Louisiana CBL resembles closely Csanady's model of a coastal jet (1977). Wind stress acts upon a stratified water column in shallow water to accelerate the upper layer. In adjusting to the presence of the shoreline, water is transported perpendicular to shore and a slope in the water surface is generated. The water slope and wind stress combine to drive strong alongshore currents in a coastal jet. In the Great Lakes and along the U.S. East Coast, Csanady finds the CBL to extend approximately 10 kilometers offshore, where the water depth is 30 meters.

Unlike Csanady's model of the coastal boundary layer, the CBL of west Louisiana contains two coastal jets. In the September transect, the outer jet registered higher velocities than those recorded near shore. The outer jet in September appeared to be associated with the topography of Sabine Bank and the strong density gradient 27 kilometers offshore (Fig. 2.7 d, e). There are insufficient data to determine the dynamical balances responsible for the occurrence of the outer jet, but it is obvious that there is some relationship between topography, the density structure, and the current.
As noted in previous paragraphs, the local wind also plays an important role in the distribution of fresh water from the Atchafalaya River and in helping to generate the coastal jet. Wind stress contributions to circulation in shallow water are far reaching, and for that reason the wind stress over the northwestern Gulf of Mexico shelf will be summarized in the following section.

**Regional Wind Stress**

The importance of wind-driven transport over the deep western Gulf is pointed out by Armstrong (1976), who attributes changes in direction of nearshore flow over the Texas shelf to seasonal fluctuations in the strength and direction of this transport. Sharp seasonal shifts in the magnitude and direction of Ekman transport are also indicated by Gunn (1978).

Autumn and winter transport in the deep Gulf is toward the northwest and constant, but is followed by a gradual increase in early spring. Transport during April is double that of the previous month, and the sudden increase in magnitude is accompanied by a shift in transport direction from northwest to northeast. A decrease in the magnitude of transport begins in May and continues until autumn, when the magnitude again levels off and the direction of transport shifts back to the northwest.

Although Gunn's analysis of Gulf surface water response to winds is informative, shelf currents are known to respond readily to local wind stress as well as regional patterns. Monthly mean wind vectors in 1-degree squares over the shelf region of the northwest Gulf of Mexico are presented by Angelovic (1977). Easterly winds, which drive westerly currents over the Texas-Louisiana shelf, are prevalent through-
out the winter. The winds shift to southeasterly in the spring and summer. The orientation of the south Texas coast and the spring wind shift combine to encourage the formation of a northerly current over the Texas shelf. Summer southerly winds intensify this current, and it moves onto the Louisiana shelf, causing the easterly summer currents reported by Kimsey and Temple (1963) and Oetking (1974) as far east as Timbalier Island. September winds are from the east; westward and southwestward currents are rapidly reestablished over the Texas-Louisiana shelf and prevail throughout the winter.

A similar situation occurs over the Oregon continental shelf. Winds are southwesterly throughout the winter, and the resulting currents are northward throughout the water column. A shift in the wind direction to northerly takes place in spring. The entire water column responds as currents reverse and flow southward (Huyer et al., 1975).

The dominant west-southwest flow over the shelf in the northwestern Gulf of Mexico and the swift wind-induced summer current reversal are evident in drifter studies conducted by Kimsey and Temple (1963, 1964), Hill et al. (1975), and Schideler (1979). In each case, the northward currents over the Texas shelf and the eastward current in Louisiana coastal waters are preceded by the shift to southerly winds.

**Tides**

Tidal currents do not appear to be sufficiently strong to drive a significant residual flow; so they will not be discussed here. Frey (1981) presents the results of standard tidal analyses for current time series from the WLICS.
SUMMARY

The current regime over the West Louisiana Inner Continental Shelf is not well understood. An attempt to coherently synthesize existing knowledge by describing the normal forcing mechanisms in shallow water, examining water motion in nearby regions, and adapting results from other regions to this unique environment results in a complex pattern of circulation. The two most important driving forces result from the uniqueness of the location of the WLICS near the Mississippi and Atchafalaya rivers and within the belt of maximum cold-front passages during winter.

A narrow band of brackish water exists over the WLICS during much of the year as a result of the influx of fresh water from the Atchafalaya River and the channeling of the high rainfall over south Louisiana through local drainage basins. The advent of spring flood on the Atchafalaya River alters the hydrography of the entire Louisiana shelf from April through August or September. Fresher water rafted westward onto the Texas shelf by spring winds is later transported eastward onto the Louisiana shelf in response to southerly winds, which cause generally northeastward currents.

Beginning in November, cold fronts move into the Gulf of Mexico from the northwest and cool the shelf waters. The process of cooling accelerates as the frequency of frontal passages and their strength increase into winter. Step-wise cooling of the water column results in the in situ formation of a local density maximum on the northwestern shelf of the Gulf of Mexico during January. This structure remains until destroyed by the spring flood and vernal heating. Geostrophic
flow associated with the density maximum would enhance westward flow over the inner shelf of west Louisiana and result in weak eastward currents over the outer shelf region. The western arm of the cyclonic gyre, which moves southward over the Texas shelf as winter progresses, could contribute approximately one-third of the winter transport of the northern limb of the western Gulf anticyclonic gyre (Molinari et al., 1978).

At the frontal time scale, west Louisiana shelf currents should vary markedly during winter owing to winds accompanying cold-front passages. An investigation of inertial currents in the Louisiana Bight generated by the alternating cross-shelf wind shows the response of the water column at the inertial frequency only (Daddio et al., 1978). The low-frequency response of the current field over the WLICS to frontal passages is unknown, although Smith (1978) presents data from the Texas shelf that indicate that the winds accompanying a cold front coherently drive currents. An in-depth investigation of several cold-front passages and their effects on the shallow inner-shelf currents of west Louisiana is the ultimate intent of this dissertation. The next step toward that aim is to understand the seasonal characteristics of wind stress and currents and the interaction between them. These topics are addressed in the following chapter.
CHAPTER 3

LOW-FREQUENCY SEASONAL VARIABILITY OVER THE WLICS

This chapter is devoted to an investigation of the low-frequency (<0.5 cycles per day (cpd)) variability of wind stress, water level, and near-bottom currents over the West Louisiana Inner Continental Shelf (WLICS). Seasonal fluctuations of these parameters are examined against the general background of shelf circulation presented in Chapter 2, utilizing information from a National Ocean Survey (NOS) data set.

Although the most common feature of continental shelf waters is the response of currents to local winds, the currents are modified by bottom topography, stratification, the offshore current regime, and, perhaps most importantly, the presence of a coastline (Allen, 1980; Winant, 1980). Investigations of shelf waters of North America reveal the importance of the seasonal variations of alongshelf wind stress to the low-frequency shelf circulation patterns. At somewhat higher frequencies, the dynamic response of shelf water to synoptic-scale storms has been shown to be even more impressive.

Allen (1980) summarizes much of the research that has been conducted over the continental shelf of the northwest United States. Seasonal fluctuations in the wind stress direction strongly affect water level and currents over the Oregon shelf (Smith, 1974; Huyer et al., 1975). During winter, southwesterly winds drive currents at all depths northward over the steep, narrow shelf. Winds shift to northerly in late spring, and the response of the water column is dramatic. Within a period of one week, currents throughout the water column are flowing southward.
Ekman transport caused by the southward current lowers sea level at the coast, and upwelling conditions prevail. Synoptic variations in sea level are highly coherent with current fluctuations throughout the year (Smith, 1974).

Investigations dealing with low-frequency fluctuations around the U.S. Gulf of Mexico are scarce. Niiler (1976) reports coherence between wind stress and current and a response to cold-front passages at the west Florida shelf break during winter. Inner shelf waters have been instrumented off Port Aransas (Smith, 1975), Bryan Mound (Hann and Randall, 1981), and northwest Florida (Marmorino, 1982), but the investigations emphasize diurnal and higher frequency fluctuations. Smith (1975) does point out the current reversal that results in predominantly north-eastward currents over the Texas shelf during mid- and late summer. In the previous chapter, evidence was presented that shows that this current reversal extends to the central Louisiana shelf during summer. Abrupt changes in current direction over the WLICS during winter frontal passages also occur. Using techniques similar to those employed for the Oregon shelf, I will further describe the seasonal and synoptic variability of the WLICS in response to wind stress forcing.

After the data set and analysis techniques are discussed, basic statistical techniques are used to determine the seasonal characteristics of the primary forcing mechanism, the wind, and the primary response, the current. Finally, the seasonal variability of interactions between wind stress and current is addressed.

DATA PREPARATION

The area 11 kilometers south of Holly Beach, Louisiana, is representative of the West Louisiana Inner Continental Shelf. This area was
instrumented during the period 1 June 1978 to 1 June 1979 by the National Ocean Survey (NOS) to provide background physical information on the lower layer of the water column in order to estimate possible effects of the brine discharge from the West Hackberry site of the Strategic Petroleum Reserve Project for the Department of Energy. Grundy 9021 current meters were fixed onto subsurface platforms at 100 and 300 centimeters above the bottom. Approximately 9.5 meters of water covers the relatively smooth bottom, and local isobaths near the moorings roughly parallel the east-west-trending coastline. Three subsurface platforms separated by 9.3 kilometers, were placed along one such isobath. Approximately 3.7 kilometers separated platforms placed normal to the isobath (Fig. 3.1). This configuration was chosen to maximize the amount of information about near-bottom currents around the proposed discharge site.

Currents were recorded at 5 minute intervals during the summer. Current data were not recorded during September because of Tropical Storm Debra's passage directly over the study site in late August and the consequent damage to several platforms and current meters (Frey, 1981). A new sampling rate of 10 minutes was used when the platforms were operational again in October. Furthermore, the meters were shifted from platform to platform, making it difficult to construct a long time series. It is logical to assume that the current at 100 centimeters is within the bottom boundary layer during much of the time and is certainly more affected by bottom friction than that at the 300 centimeter level. The current at 300 centimeters should therefore provide a better representation of the current throughout the water column. Hydrographic data from the area (Frey, 1981) suggest that the water column is often well mixed and a single meter can be assumed to be representative of the entire
Figure 3.1. The West Louisiana inner continental shelf showing the location and configuration of NOS subsurface platforms.
water column. The current data from 300 centimeters above the bottom at
the middle and western sites are, therefore, used in all further analyses
(Fig. 3.2). Because this study focuses on low-frequency currents, data
from these two sites were combined to produce the longest possible record.

In order to discuss most conveniently dynamics of these shallow
waters, current components parallel to shore (alongshelf) and normal to
shore (cross-shelf) were computed. As previously pointed out, the west
Louisiana coastline is aligned nearly east-west, and therefore I chose my
alongshelf axis to be aligned east-west. The normal right-hand co-
ordinate system is rotated 90° clockwise, resulting in the positive Y-
axis pointing to the east. Current components toward the east are
referred to as positive v, or alongshelf. Cross-shelf flow components, u,
are positive offshore, along the X-axis.

Water level was recorded by an Aanderaa Model WLR-5 water level
gauge, which was affixed to the center subsurface platform. This model
measures total pressure in millibars (including atmospheric pressure).
The water level record shows gaps during the year (Fig. 3.2), but it is
long enough to analyze the tide and estimate the statistics of long-term
(2-20 days) fluctuations in water level during summer, autumn, and
winter.

Owing to the absence of water level data during March and April and
the importance of this period for an in-depth investigation of how the
water column responds to individual cold fronts (Chapter 4), two months of
tide records from Calcasieu Pass were also obtained. A Stevens-type gauge
was maintained by the Corps of Engineers at that location. Those data
were found to be comparable to water level from offshore and thus were
used in later analyses.
Figure 3.2. Time lines showing the duration of the data sets used in the analyses of current, water level, and wind stress. L.C. wind indicates wind data from Lake Charles.
Atmospheric parameters were recorded at an instrumented rig near the center platform. An Aanderaa meteorological station provided wind speed and direction, barometric pressure, and air temperature for short periods during the investigation (Fig. 3.2). Large gaps in the wind record led to the search for a longer time series of wind speed and direction.

A complete year of wind speed and direction at 1-hour intervals was available for the period 1 June 1978 to 1 June 1979 from Lake Charles, Louisiana, approximately 55 kilometers northeast of the platforms. Wind speed from the rig and Lake Charles are compared (Fig. 3.3), and it is obvious that the offshore speeds are stronger. Hsu (1981) presents a method for converting land-based wind speed to offshore wind speed (the dashed line in Fig. 3.3). Following Hsu, the Lake Charles wind speeds are transformed using

$$U_o = 3.0 \times W^{0.6667}$$

where $W$ is the Lake Charles wind speed in meters per second. The computed wind speeds follow the general pattern of the winds recorded at the offshore rig site.

Total wind stress is then calculated,

$$TW = \rho_a C_D U_o^2$$

where $\rho_a$ is air density, $1.2 \times 10^{-3}$ g/cm$^3$, and $C_D$ is an atmospheric drag coefficient, $1.26 \times 10^{-3}$ (Hsu, 1978). Wind stress components are computed using the same coordinate system as for the current. These wind stress values for the WLICS are utilized in all further analyses.
Figure 3.3. Plot of Lake Charles (onshore) versus rig (offshore) wind speeds. The dashed line shows the fit of Hsu (1981) to the data set.
This methodology permits construction of a continuous wind stress record for use in an investigation of low-frequency fluctuations of inner-shelf dynamics. As noted previously, no similarly continuous current record exists. It is possible, however, to link overlapping records from neighboring platforms to create a longer time series. When dealing with low-frequency fluctuations, it is assumed that alongshore flow is coherent along the shelf (Boicourt and Hacker, 1976). Although alongshelf coherence (not shown) is not as high as expected during autumn and winter, it is statistically significant at the 95 percent significance level, and as a result the autumn record from the middle station is merged with the winter record from the west platform to form a continuous time series of current from October through April.

In order to investigate low-frequency (<0.5 cpd) fluctuations of wind stress, current, and water level, it is necessary to remove the high-frequency components. For this purpose a Lanczos filter with a half-power point of 40 hours is applied to all of the data and the output resampled at 3-hour intervals. (The mechanics and response of the Lanczos filter are presented in Appendix A.) Diurnal and higher frequency components are thus removed from the resulting time series. The sea breeze and storm events of less than one day's duration are no longer represented in the wind stress record, nor are their responses seen in the current records.

Diurnal and semidiurnal tidal components are no longer present, either. These high-frequency motions are described by Frey (1981); so the remainder of this chapter focuses on the low-frequency time series.

**Seasonal Variability**

A large amount of information about low-frequency variability in the water column over the WLICS is revealed through time series analysis
of the filtered data. Combined with the background information presented in the previous chapter and a knowledge of air-sea interactions, a better understanding of low-frequency motion over the WLICS is possible. The spectral analysis techniques to be utilized during this investigation are described in Bendat and Piersol (1971).

The autospectrum of each time series is examined to determine the temporal variability of that parameter's variance as a function of frequency. The spectra are computed using a 5 day lag window yielding a frequency resolution of 0.062 cpd and 15 degrees of freedom. Qualitative visual relationships between variables are inferred from the autospectra. Quantitative evidence of the interrelationship between any two time series is derived from the cross spectra and the background information provided in Chapter 2.

For uniformity and convenience, the components of low-passed wind stress and current are divided into the six 60-day periods presented in Table 3.1. The 30-day overlap between periods aids the understanding of how the interactions vary.

The autospectra of the cross-shelf and alongshelf components of wind stress and current during summer and from mid-autumn through winter are presented in Figure 3.4. Visual comparisons of the components during a particular period or over the entire study period are possible. In addition, the very low frequency variability is easily seen.

For the cross-shelf wind stress component (Figure 3.4a), a summer energy density maximum appears between frequencies of 0.05 and 0.25 cpd. The alongshelf wind stress maximum, which is approximately equal in amplitude to the cross-shelf maximum, is found below 0.15 cpd. Compared to the wind stress variance during the remainder of the year, these values
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<th>Division of 60-Day Periods</th>
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<td><strong>Summer</strong></td>
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<td><strong>Mid-Autumn</strong></td>
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<td><strong>Late Autumn</strong></td>
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<td><strong>Early Winter</strong></td>
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<td><strong>Mid-Winter</strong></td>
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<td><strong>Late Winter</strong></td>
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<td>2100 6 April 1979</td>
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Figure 3.4 The autospectra of (a) cross-shelf wind stress, (b) alongshelf wind stress, (c) cross-shelf current, and (d) alongshelf current. Inset: the 95 percent confidence intervals. Units for the wind stress autospectra are tenths of a dyne\,\text{cm} per cycles per day and tens of cm\,s\textsuperscript{2} per cycles per day for the current autospectra.
are extremely low. The low energy densities are indicative of the atmospheric conditions that prevail over the WLICS during summer. Other than the infrequent tropical storm or hurricane, the dominant meteorological event in this region during summer is the diurnal sea breeze system, which is not within the frequency range being investigated.

The low-amplitude, red spectrum of alongshelf wind stress points to a lack of organized weather systems in the alongshelf direction, at least at periods of 10 days of less. Concentration of cross-shelf energy density around 0.15 cpd points to a more frequent meteorological event as the source. DiMego et al. (1976) show that two to three fronts cross per month into the northwestern Gulf of Mexico during the summer. The fronts are weakly organized and would be more properly referred to as weak weather systems. The frequency of occurrence of the maximum cross-shelf wind stress variance coincides with the frequency found by DiMego et al. It is suggested that the increased variance found around 0.15 cpd is associated with these weak summer weather systems passing over the WLICS.

Autospectra for the cross-shelf and alongshelf currents are quite similar to each other (Fig. 3.4c and d). Both are red, with approximately the same variance at the lowest frequencies. Equality of the variances reflects the slow meandering nature of the summer current (Fig. 3.5). The coastal constraint (Csanady, 1981) appears to be only weakly operative at this location during summer. Since at this time of year we expect a stratified coastal boundary layer with a strong preference for alongshelf flow, the lack of a strong directional preference suggests that the current meter was beneath the pycnocline. Discussion of the interaction between the summer wind stress and current is postponed until coherence is examined.
Figure 3.5. The Summer 1978 progressive vector diagram showing the inferred Lagrangian movement over the WLICS.
Data are not available from August or September, hence the gap in the autospectrum plots (Fig. 3.4). At the beginning of the mid-autumn period, changes in the wind stress are evident. The cross-shelf wind stress variance is increased considerably over summer levels at all frequencies below 0.5 cpd. The largest increase coincides with a shift in the maximum variance to frequencies below 0.15 cpd. A different pattern emerges for the alongshelf wind stress autospectrum. Increased variance occurs only between 0.10 and 0.25 cpd. Variance at the remaining frequencies decreases or changes very little. It is apparent that the cross-shelf wind stress is higher than the alongshelf component at the frequencies being investigated during this particular mid-autumn period.

Figure 3.4c shows that the cross-shelf current variance decreases from summer values at frequencies below 0.15 cpd. The alongshelf current variances have increased at all frequencies below 0.45 cpd. The greatest increase in variance results in a mid-autumn maximum near 0.20 cpd, where it is over 500 percent greater than the summer value. The low-frequency alongshelf current is beginning to dominate the cross-shelf current at this time. The coincident appearance of peaks in the autospectra of the alongshelf components of wind stress and current suggests a relationship near 0.20 cpd.

The late-autumn period is similar to mid-autumn in many respects (Fig. 3.4). The variance of both wind stress components increases in magnitude, and peaks shift to higher frequencies. However, the largest increase occurs in the cross-shelf component. A significant maximum appears near 0.20 cpd. The highest values of alongshelf variance are also found in the 0.20 cpd frequency band, but they are approximately one-half the magnitude of the cross-shelf wind stress variance, suggesting that the dominant frontal winds parallel the north-south-trending fronts.
Alongshelf current continues to dominate flow over the WLICS. Although the alongshelf current energy density increases in magnitude over the entire low-frequency range, the cross-shelf current remains very small. Apparently the high wind stress mixes the water column and drives the current alongshelf. The largest energy density values occur within the 0.10 cpd frequency band, although they are also high near 0.20 cpd.

During the early winter the total variance for cross-shelf wind stress at low frequencies appears to be approximately the same as that observed during late autumn. The primary difference between the two periods is the shift in maximum variance from near 0.20 cpd in late autumn to near 0.30 cpd in early winter. Although the total variance of the alongshelf wind stress remains lower than that for the cross-shelf component, it is noticeably larger than that found during the previous late autumn. Figure 3.4 shows two distinct peaks of approximately the same magnitude for alongshelf wind stress variance near 0.15 and 0.30 cpd. It is probable that the two wind stress components are related near 0.30 cpd. This frequency corresponds to the frequency of winter cold-front passage over the northwest Gulf of Mexico (DiMego et al., 1976). Further analysis is necessary to determine whether the two components are related in a statistically significant fashion.

Alongshelf current variance undergoes a further increase in the early winter period. Most of the increase is between 0.20 and 0.35 cpd, but the variance also increases between 0.05 and 0.20 cpd. The dominant alongshelf maximum remains near 0.30 cpd. Cross-shelf energy density also increases around 0.25 to 0.30 cpd, but the values are much lower than those for the alongshelf current. The alongshelf flow remains dominant.
Energy density levels for the cross-shelf wind stress at frequencies below 0.25 cpd experience an abrupt change in their magnitude between early and mid-winter. The variance peak near 0.30 cpd remains relatively unchanged, but a measurable decrease in variance occurs near 0.20 cpd. A 200 to 300 percent increase in variance near 0.10 cpd forms the new dominant cross-shelf wind stress variance maximum. Sizable increases are also noted around 0.05 and 0.15 cpd. Unlike the cross-shelf component, little change is noted for the alongshelf wind stress. Variance near 0.30 cpd rises to a moderately greater level, but most other variances remained relatively constant, including the secondary maximum around 0.15 cpd. Maintenance of the previous variance levels indicates that the cross-shelf and alongshelf wind stress components are approximately equal to each other in total variance.

Cross-shelf current variance continues to increase into the mid-winter period at most frequencies below 0.40 cpd. The highest cross-shelf current variance occurs around 0.30 cpd. A minimum near 0.20 cpd is also evident and will be addressed later. The alongshelf current variance continues a pattern begun during the early winter. A maximum spanning the 0.25 to 0.30 cpd bands dominates the mid-winter autospectrum, although relatively high variance values are also present below 0.15 cpd. A minimum near 0.20 cpd is also shown in this autospectrum.

Late winter (February and March) is the final period investigated in this dissertation. An abrupt change in the autospectra of all components occurs during the transition from mid- to late winter of this year (Fig. 3.4). The increase in cross-shelf wind stress variance at the lowest frequencies continues at a rapid rate and is coupled with the decrease of variance near 0.30 cpd. A large part of the cross-shelf wind
stress energy is confined to the lowest frequencies. In fact, the variance near 0.10 cpd is the highest of the record.

The alongshelf wind stress change is not as abrupt as that which occurs in the cross-shelf component, although the variance decreases at most frequencies. Total cross-shelf wind stress once again exceeds alongshelf wind stress at low frequencies. Maximum alongshelf wind stress variance has shifted to frequencies below 0.15 cpd.

The current autospectra also show changing energy levels and a shift in the frequency of maximum energy density. Cross-shelf current variance is maximum near 0.10 cpd and is much lower elsewhere. It is also important to note that the maximum is relatively large, suggesting significant cross-shelf flow during this period. Alongshelf current variance also increases to the highest values of the record with a shift in the variance peak toward lower frequencies. The shift is apparently due to changing weather patterns. The winter variance peak near 0.30 cpd decreases by more than 70 percent from its mid-winter high.

The autospectra presented in Figure 3.4 show the variability of the wind stress and current fields over the WLICS. Significant temporal variability occurs in both the shape and the magnitude of the autospectra. A synopsis of this section serves to introduce the next section, which will contain an explanation of some of the probable causes of the variability.

During the summer period both wind stress and current are low and variance is concentrated at the lowest frequencies. Both cross-shelf and alongshelf levels of the alongshelf flow are significant. Energy density levels indicate that the alongshelf current and both wind stress components increase during the autumn period. Little cross-shelf current
energy is observed. Instead of occurring at the lowest frequencies, the energy density maxima of wind stress and alongshelf current shift to the 0.15 to 0.20 cpd band.

The winter must be considered in two sections. During early and mid-winter both components of wind stress and current have relatively high energy levels in the 0.30 cpd frequency band. The wind stress components show approximately equal energy levels during this period. Alongshelf wind stress and both current components have a maximum near 0.15 cpd. Late winter is characterized by a shift of the variance maxima to the lowest frequencies and large increases for all components at those frequencies. All components are at their most energetic levels of the record.

Seasonal Interactions

Figure 3.6 is a plot of the low-frequency coherence between paired components for the six 60-day time periods. Figure 3.6a, which shows the coherence between the cross-shelf and alongshelf wind stress components, sheds some light on the causes for the changing frequency of maximum variance throughout the period of interest. Coherence exceeds the 95 percent significance level between 0.20 and 0.25 cpd during the summer. If the summer wind stress coherence near 0.20 cpd is interpreted as wind stress associated with a coherent, migrating weather system, then the relatively high variance of the cross-shelf wind stress suggests that the weather system contained predominantly cross-shelf winds. An indication of the weakness of the weather system and its associated winds is the low autospectra values near 0.20 cpd. The patchy summer coherence between cross-shelf wind stress and alongshelf current is very difficult to interpret physically.
Figure 3.6. Coherence squared values for 60-day periods between (a) cross-shelf wind stress and alongshelf wind stress, (b) cross-shelf wind stress and alongshelf current, (c) alongshelf wind stress and cross-shelf current, and (d) alongshelf wind stress and alongshelf current. SL represents 95% significance level.
There does appear to be a relationship between the alongshelf wind stress and both current components at the lowest frequencies. However, the negative phase (not shown) between the alongshelf components of wind stress and current indicates that the wind stress does not accelerate the near-bottom current. The wind stress may generate a pressure gradient against the Texas Coast or the Mississippi delta, which would then cause near-bottom currents to flow against the wind. Weak weather systems that move across the WLICS near a frequency of 0.20 cpd do not appear to have an effect on the water column.

Coherence between wind stress components during mid-autumn is significant only at the highest frequencies considered. Figure 3.6b-d, which shows the coherence for wind stress/current combinations, indicates that the highest coherence is around 0.20 cpd. The mechanism for the interaction is unclear. Although the highest coherence is between the alongshelf components of wind stress and current, the negative phase (not shown) shows that the wind stress does not force the current directly. Chuang and Wiseman (1982) suggest that cross-shelf wind stress may pile water up against a local nearshore shoal and initiate a pressure gradient which drives the current against the prevailing alongshelf wind stress.

A slightly clearer picture of low-frequency wind stress/current interactions is found in the plots of late-autumn coherence. The highest significant coherence between cross-shelf and alongshelf wind stress appears near 0.20 cpd, suggesting the formation of organized weather systems. Although the cross-shelf wind stress variance exceeds the alongshelf variance in this band, both have a greater magnitude than during summer, inferring that the weather system is not only organized during late autumn but stronger than in summer.
The effects of these weather systems are evident in the coherence plots for wind stress/current combinations. The highest coherence values for cross-shelf wind stress/alongshelf current and alongshelf wind stress/cross-shelf current occur near 0.20 cpd. The weather systems (fronts) are apparently involved in forcing the water column in this frequency band, but the mechanics of the interaction are not immediately clear. Although coherence between alongshelf components of wind stress and current is also high in this band, higher coherence is found at lower frequencies. Both wind stress components appear to influence circulation near 0.20 cpd; and, at lower frequencies, the alongshelf wind stress forces the alongshelf current.

During the early-winter period the energy density appears to be partitioned into two distinct frequency bands. Figure 3.6a shows maximum coherences between the cross-shelf and alongshelf wind stress components occurring around 0.15 and 0.30 cpd during this period. A frequency of 0.30 cpd corresponds to a 3.3-day period, which is in agreement with the timing of winter frontal passages (DiMego et al., 1976). The maximum at 0.15 cpd may indicate that every other cold front is stronger and more organized.

For the early-winter period the highest coherences are centered near 0.15 and 0.30 cpd. Cross-shelf wind stress is coherent with the alongshelf component at both frequencies; therefore it is important to examine the coherence of both wind stress components with the current components before suggesting possible forcing mechanisms. The alongshelf wind stress/cross-shelf current coherence is maximum at 0.15 and 0.30 cpd, whereas the alongshelf component coherence is highest at 0.30 cpd. The alongshelf components of wind stress and current are in phase. Energy
density levels were also highest for these components within these frequency bands. Although the cross-shelf component of the wind stress is greater than the alongshelf component, higher coherence exists between the alongshelf wind stress and both current components. High coherence between the cross-shelf wind stress and alongshelf current is in large part a result of the high coherence between the wind stress components in these frequency bands.

The suggested scenario during early winter is of frontal systems, particularly the alongshelf component, forcing both the alongshelf and cross-shelf currents at frequencies near 0.30 cpd. The energy density peak at 0.15 cpd again suggests that every other front is stronger. Even though cross-shelf wind stress, associated primarily with "northerns" following a cold-front passage, is greater than the alongshelf component, the latter is more effective in forcing the water column.

A similar pattern is present for the mid-winter period (January and February). Although coherence between the wind stress components near 0.30 cpd is lower than during the early-winter maximum, it remains significant (Fig. 3.6a). Fronts are evidently less well organized during this 60-day period. The alongshelf wind stress variance exceeds the cross-shelf component around 0.30 cpd, indicating a change in the dominant winds associated with frontal passages to the alongshelf axis. High variance values associated with the cross-shelf wind stress near 0.10 cpd do not appear to be related to the alongshelf wind stress (Fig. 3.6a).

Although cross-shelf wind stress and alongshelf current are coherent at frequencies of 0.30 cpd and higher, the highest coherence levels occur around 0.30 cpd and are associated with the alongshelf wind stress and current interactions. Alongshelf wind stress and current are in phase
near 0.30 cpd. The dominant pattern suggested for the mid-winter period is wind stress forcing of the water column, predominantly the alongshelf current, by alongshelf wind stress accompanying frontal passages. Cross-shelf wind stress energy density near 0.10 cpd does not have a significant effect on the currents over the WLICS.

An abrupt change in the weather patterns and consequent forcing of the water column is shown in the late-winter period. Significant coherence between the wind stress components is found only at frequencies above 0.40 cpd (Fig. 3.6a). The organized winter weather systems have apparently dissipated and been replaced by a predominantly cross-shelf wind system, such as the south-southeasterly winds that are common over the WLICS during spring. It is obvious that spring weather patterns are becoming dominant at this time.

Significant coherence levels between wind stress and current components that are in phase infer a continued dominance of alongshelf wind stress/current interactions at the low frequencies investigated (Fig. 3.6b, c, d). High coherence near 0.30 cpd is not associated with frontal passages during this period, although the maximum coherence between the alongshelf wind stress and both current components is located near that frequency. Even though the cross-shelf wind stress variance is at its highest level of the record near 0.10 cpd, there appears to be no significant interaction with shallow-water currents. The low-frequency (less than 0.15 cpd) alongshelf current is associated with the alongshelf wind stress, although the transfer function is higher than during early and mid-winter.
CONCLUSION

Weather systems that influence the shallow-water currents over the WLICS are shown to vary greatly in their spectral composition during the course of the year. Spectral analysis of the low-frequency wind stress and current components allows interpretation of the intensity of interactions between components and insight into how these occur. The findings are very similar to those from other inner-shelf regions in that alongshelf wind stress provides the primary forcing for shallow-water currents.

Summer is characterized by slow-moving weather patterns. Weak, poorly organized systems with a dominant cross-shelf wind stress component cross the WLICS at a frequency of about 0.20 cpd and have no significant effect on the water column. Alongshelf wind stress energy density is highest below 0.15 cpd, where it is coherent with alongshelf and crossshelf currents that are of approximately equal energy. Low-frequency alongshelf wind stress forces currents in the water column during the summer period.

During the autumn, weather systems become better organized. These weather systems, which have their strongest signal in the frequency band near 0.20 cpd, have a higher coherence between wind stress components than during summer, suggesting a coalescing of the weather systems into more coherent fronts. Although the cross-shelf component of the wind stress exceeds the alongshelf component, it is the latter that has the strongest interaction with the water column. Because cross-shelf flow is very weak, it appears the alongshelf wind stress drives the alongshelf current during autumn.

Fronts become more organized in winter and the alongshelf wind stress component strengthens. Alongshelf wind stress and current are
highly coherent (>0.90) near 0.30 cpd, suggesting forcing of the water column by the alongshelf wind stress. The cross-shelf variance also increases during the winter and appears to be related to the alongshelf wind stress. Figure 3.7 shows the response of the water column to several frontal passages during January 1979. A large amount of cross-shelf flow is recorded during the frontal events.

Late winter appears to be a representation of spring conditions. The return to predominantly cross-shelf wind stress is in agreement with the south-southeasterlies of spring. Currents are strongly alongshelf, apparently in response to the alongshelf wind stress, which is lower than cross-shelf stress. It is unlikely that the alongshelf wind stress is responsible for all of the extremely high variance of alongshelf current present at the lowest frequencies. The high discharge during the flood of the nearby Atchafalaya River is probably responsible for part of that variance, although quantitative data are lacking.
Figure 3.7. Winter 1979 progressive vector diagram showing inferred Langrangian movement over the WLICS.
CHAPTER 4

LOW-FREQUENCY RESPONSE OF THE WLICS TO COLD-FRONT PASSAGES

Introduction

Generation and translation of cold fronts dominate atmospheric conditions in mid-latitudes during winter. Evidence from the New England shelf (Beardsley and Butman, 1974), and now from the West Louisiana Inner Continental Shelf, indicates that these transient wind events are strongly coupled with circulation in coastal waters. The response of the water column depends greatly on frontal characteristics and the location of the front with respect to the coastal waters under investigation.

Fronts are discontinuities between air masses of different densities. The density differences are usually attributable to the temperature of the air masses. In North America during winter, waves on the polar front, between the polar northeasterlies and temperate westerlies, move into the continental United States from the northwest. As a wave moves into the lee of the Rockies, an extratropical cyclone with a southwesterly trailing front often forms (DiMego et al., 1976). The frontal system then moves rapidly (20-30 km hr\(^{-1}\)) toward the southeast.

Polar and equatorial air masses move away from their source regions and along the front; the warm sector ahead of the front is nourished by moist winds from the southern quadrant while northwesterly to north-easterly winds following the front transport dry polar air southward. The cold, dense air behind the front wedges beneath the warmer air ahead. Because of frictional effects at the surface, a relatively steep frontal zone of intense activity forms.
The front is a region of relatively low pressure. As it approaches a fixed location, barometric pressure falls rapidly. Frontal passage is followed by a steady increase in barometric pressure. Also associated with the frontal passage is a veering of surface winds from a south-easterly or southwesterly to northerly direction (Fig. 4.1). These low-level winds differ markedly from the jet stream and greatly affect the inner shelf waters as the front crosses the Gulf Coast into the northwestern Gulf of Mexico.

Abrupt changes in the water column occur as a result of cold-front passage. Heat is released from the water column into the cold, dry air mass behind a front by evaporation and conduction. The shallow inner-shelf water may cool by 10°C in response to a single cold frontal passage (Nowlin and Parker, 1974). Momentum is transferred into the water column by high wind stress in the vicinity of the front, mixing the water column to great depths. Successive cold fronts mix and cool the water column over much of the shelf, beginning in the bays and estuaries and progressing offshore (Huh et al., 1978).

Attendant wind stress also generates currents and water level changes. Over the New England continental shelf, the shelf waters' response depends on the location of the storm with respect to the shelf region (Beardsley and Butman, 1974). Easterly winds (storm center south of shelf) produce westward currents and a sea level rise along the coast. Little alongshelf flow results from westerly winds, although significant alongshelf and cross-shelf pressure gradients are observed.

In the previous chapter, I show that winter alongshelf winds at the frequency of frontal passages are closely associated with the water column response over the WLICS. It is my purpose in the remainder of this
Figure 4.1. Wind stress (direction from) and barometric pressure accompanying a cold-front passage.
dissertation to describe the response of the water column to individual cold-front passages and the variability of that response. In order to accomplish this goal, four periods during the winter of 1979 are chosen for analysis. In contrast to the New England shelf, the relative strengths of the alongshelf and cross-shelf stress will be the factor found to most control shelf current response to frontal forcing. The selected periods (Table 4.1) contain fronts with differing characteristics. The general characteristics of each of these periods and the inner-shelf response are revealed by examination of the relationships between wind stress and current. Finally, the dynamics of the water column are investigated by estimating and comparing the terms in the alongshelf momentum balance.

### Frontal Periods

Daily weather maps (NOAA, 1979) for the selected periods show the general meteorological conditions for the northwestern Gulf of Mexico. The formation of extratropical cyclones and fronts and their translation across the United States can be followed using these maps. These general weather patterns provide background information for a better understanding of meteorological forcing of the WLICS waters. Adjusted wind stress and barometric pressure from northeast of the WLICS at Lake Charles supplement the regional meteorological picture. The response of the water column over the WLICS is revealed through the current and water level records for each of the selected periods.

#### January

During the January study period, three similar frontal systems passed through the region. During the initial three days (16-18 January)
Table 4.1
Frontal Periods

<table>
<thead>
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<th>Month</th>
<th>Day</th>
<th>Date</th>
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<th>Date</th>
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<tbody>
<tr>
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<td>0000</td>
<td>16 January</td>
<td>0000</td>
<td>31 January</td>
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<td>February</td>
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<tr>
<td>March</td>
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<td>16 March</td>
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<td>31 March</td>
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<td>April</td>
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<td>6 April</td>
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<td>19 April</td>
</tr>
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an extratropical cyclone crossed into the central United States, and was followed by a strong Canadian high-pressure system that pushed cold polar air down the Mississippi Valley. Circulation around another high-pressure system centered over the northeastern Gulf of Mexico kept the cold front associated with the extratropical cyclone well north of the Gulf Coast. Barometric pressure at Lake Charles fell steadily during this period, and winds were primarily southeasterly (Fig. 4.2a, b). Currents over the WLICS were predominantly westward at approximately 10 cm sec\(^{-1}\) (Fig. 4.2c, d), and the dominant water level signal was tidal (not shown). A second extratropical cyclone formed in the lee of the U.S. Rockies and trailed a cold front across the southwest on 19 January. The movement of this and two succeeding frontal systems is shown in Figure 4.3.

During the following day (20 January) the cyclone moved to the northeast and its associated cold front overtook the front trailed by the preceding cyclone. At 1200 on 20 January (CMT) the combined cold fronts approached Lake Charles (Fig. 4.3a). Local conditions indicate a shift in the wind to a southwesterly direction (Fig. 4.2a, b). The current over the WLICS remained predominantly westward at approximately 10 cm sec\(^{-1}\) (Fig. 4.2c, d). A strong high centered over Idaho pushed polar air far into the Texas area behind the front.

The daily weather map for 21 January (Fig. 4.3b) shows the extratropical cyclone located over the mid-Atlantic states. The actual passage of the front is clearly indicated in the wind stress record (Fig. 4.2a, b). The time series shows the characteristic rapid shift in wind direction from southeasterly to northwesterly as the front passed over Lake Charles. Wind-stress values during the frontal passage were high. The wind stress toward the east reached 3.0 dynes cm\(^{-2}\), and the offshore
Figure 4.2. The sequence of daily weather maps (NOAA, 1979) for selected times of the January frontal period showing the movement of frontal systems across the United States. The black square represents the location of the WLICS.
Figure 4.3 January time series of (a) cross-shelf wind stress, (b) alongshelf wind stress, (c) cross-shelf current, and (d) alongshelf current. Positive is offshore or eastward showing inferred Lagrangian movement over the WLICS. Start time is 0000 16 January 1979. Distance between tick marks on horizontal axis is two days.
component approached 2.0 dynes cm\(^{-2}\). Currents also shifted rapidly (6-12 hours) after the wind reversal (Fig. 4.2c, d). Onshore flow occurred for a short period during the current reversal, then the current flowed predominantly eastward at speeds reaching 40 cm sec\(^{-1}\). Water level (not shown) decreased quickly soon after the wind shift and fell by more than a meter in response to strong wind-stress values.

A cold front trailing from a newly developed cyclone in the western United States moved eastward on 21 January. Strengthening of the cyclone and rapid southward movement characterized the mesoscale meteorology during the following day (22 January). No longer was the northwestern Gulf of Mexico under the influence of the previous extratropical cyclone, which was then centered over Maine. Wind data from Lake Charles indicates that southeasterly winds similar to those preceding the first frontal passage existed during the next 24 hours (Fig. 4.2a, b). Currents shifted back to westward soon after the wind stress attained an easterly component.

On 23 January the cyclone moved eastward and at 1200 hrs the associated cold front was located south of the cyclone, through Oklahoma and Texas (Fig. 4.3c). A high pressure system again followed the front bringing polar air southward, but the central pressure was not as high as during the previous front. Wind stress over the WLICS was southeasterly and relatively low (<1.0 dyne cm\(^{-2}\)) during this period (Fig. 4.2a, b). The current was predominantly westward at 25 cm sec\(^{-1}\) and had a slight offshore component (Fig. 4.2d). Water level returned to its previous level and the tidal character of the water column returned.

Later on the afternoon of 23 January the front pushed through the WLICS. Just prior to the frontal passage the alongshelf wind stress
shifted rapidly while the cross-shelf component remained constant, suggesting southwesterly winds immediately ahead of the front (Fig. 4.2a, b). After the frontal passage the alongshelf wind stress increased to 3.0 dynes cm\(^{-2}\) and strong west-northwesterlies dominated. Response of the water column, as revealed in the current records (Fig. 4.3c, d), was rapid and very similar to the response described for the first frontal passage.

By 1200 on 24 January the front stretched across Florida and the northwestern Gulf was under the influence of flow behind the front (Fig. 4.3d). Winds over the WLICS were predominantly easterly during this period (Fig. 4.2a, b) and the water column moved westward at 25 to 30 cm sec\(^{-1}\). Once again water level returned to previous values and the tidal signal dominated. Another cyclone over the northwestern United States began to push a cold front southeastward across the mountain states. During the next 24 hours frontolysis occurred. A new cold front formed as the low moved into the lee of the Rockies and stretched across the southwestern United States. During this interfrontal period (25 January) alongshelf winds over the WLICS were dominantly easterly (Fig. 4.2c, d), due to the presence of a high pressure center over the central United States.

The cyclone and front translated eastward during the following 24 hours, and by 1200 on 26 January the cold front was located across central Texas (Fig. 4.3e). A low previously located over northern Mexico joined with the moving system. Rapid eastward movement continued for the next 24 hours, during which the front passed through the WLICS (Fig. 4.3f). Barometric pressure decreased slowly in advance of this particular cold front. Both components of the wind stress shifted at approximately the
same time, indicating a rapid directional change as the front passed (Fig. 4.2a, b). The cross-shelf wind stress exceeded the alongshelf component suggesting a wind stress aligned almost perpendicular to the coastline. A high pressure system that moved into the central United States was responsible for the predominantly cross-shelf winds, which were prevalent for the remainder of the period.

The response of the water column during this final frontal passage differed from those of the two previous fronts. Currents shifted toward the east, but they were not as strong as during the previous frontal passages (Fig. 4.2 c, d). The current shifted back toward the west soon after the wind shifted back to easterly. The water level response to the predominantly cross-shelf wind was much less than the response to the primarily alongshelf winds of the first two fronts.

The varied response of the water column to the three January frontal passages is also revealed in a progressive vector diagram (Fig. 4.4). The pvd shows that during inter-frontal periods currents were westward and slightly offshore. Following the shift in wind direction into the westerly quadrant with the frontal passage, the currents moved eastward and slightly onshore. At least that was the pattern for the first two frontal passages. The predominantly offshore wind stress was related to a current in the onshore direction after 1200 on 27 January. After midday on 29 January the current resumed its westward course.

February.

The frontal passage selected from the month of February differed from the three passages during January. In this instance a weak, low pressure system was located over Ohio at 1200 on 12 February, trailing a cold front behind it. High level winds transported the cyclone rapidly
Figure 4.4. Progressive vector diagram during January frontal period showing inferred Lagrangian movement over the WLICS.
eastward and it was not present on the daily weather map two days later (Fig. 4.5a). Circulation around a high pressure system over the northeastern Gulf kept the cold front north of the WLICS and in doing so, a weak stationary front formed. Weak south-southeasterly winds associated with another high pressure system over the Atlantic, and lows along the stationary front dominated WLICS weather until 15 February. Wind stress over the WLICS during these three days was predominantly cross-shelf. The onshore wind stress varied between 1.0 and 2.0 dyne cm$^{-2}$ and the alongshelf wind stress was very low (Fig. 4.6a, b). The water column response was rather inconsistent. Currents were low (<15 cm sec$^{-1}$) and appeared to be predominantly offshore (Fig. 4.6c, d). The water level times series (not shown) was tidally dominated.

On 15 February the onshore wind stress increased to greater than 1.0 dyne cm$^{-2}$ (Fig. 4.6a). Strong northerly winds accompanying a high pressure system (1048 mb) located over Canada pushed polar air down the Mississippi Valley (Fig. 4.5a). The currents responded to the onshore wind ahead of the front by moving slowly onshore and toward the east (Fig. 4.6c, d).

Late on 15 and 16 February, the Canadian high forced the stationary-turned-cold front through the central Gulf Coast region. The WLICS experienced the cold northerlies prior to 1200 on 16 February and remained under their influence for at least the next 48 hours (Fig. 4.5b, c, d). Cross-shelf wind stress became strongly offshore (1.5 to 2.0 dyne cm$^{-2}$) and continued for approximately three days (Fig. 4.6a). During the initial 24 hours after the frontal passage, the alongshelf wind stress was very low. It then became westward for approximately 18 hours, but was not as great as the cross-shelf component (Fig. 4.6b).
Figure 4.5. Sequence of daily weather maps (NOAA, 1979) for selected parts of the February frontal period showing the movement of frontal systems across the United States. The black square represents the location of the WLICS.
Figure 4.6. February time series of (a) cross-shelf wind stress, (b) alongshelf wind stress, (c) cross-shelf current, and (d) alongshelf current. Positive is offshore or eastward. Start time is 0000 12 February 1979. Distance between tick marks on horizontal axis is two days.
The current response after the frontal passage was in both the cross-shelf and alongshelf directions. Initially the current increased in the onshore and eastward directions (Fig. 4.6c, d). As the dynamics of the water column began to come into balance after a period of time the current became primarily westward and slightly onshore. It was apparent that the current meter at 3 meters was responding to the return flow in the near-bottom layer and not directly to wind-driven flow. Particulars of the dynamics are to be investigated later. The time series of water level indicates little or no response to the cross-shelf dominance of this February frontal passage (not shown).

By 20 February (Fig. 4.5f) the high pressure system was located over the east coast and southeasterly winds were again prevalent. The remainder of the period was characterized by a series of stationary fronts crossing the WLICS in the Gulf of Mexico. These frontal passages did little to affect the prevailing weak south-southeasterly winds and the current was generally westward with no major shifts in direction (Fig. 4.6c, d). The progressive vector diagram shows the offshore movement prior to the frontal passage and the predominantly onshore and eastward motion following that passage (Fig. 4.7). The current remained westward during the passage of the stationary fronts as the February period ended.

March.

The early portion of the March frontal period was dominated by the movement of an extratropical cyclone across the northern United States. From 16 March to 18 March the cyclone moved from its area of cyclogenesis in the lee of the Rockies to a position over the Oklahoma panhandle. Winds over the WLICS during this period were strong and southeasterly. Wind stress values of 2.0 to 4.0 dynes cm$^{-2}$ were common in advance of the
Figure 4.7. Progressive vector diagram during February frontal period showing inferred Lagrangian movement over the WLICS.
cold front (Fig. 4.8a, b). Currents were also strong (25.0 cm sec\(^{-1}\)) and predominantly westward (Fig. 4.8c, d). The time series of water level during the two day period revealed no obvious wind-driven response.

During the following 24 hours (up to 1200 on 19 March) the cyclone moved to the north and its cold front migrated only slightly eastward to the position shown in Figure 4.9a. A low pressure center, present over the west coast the previous day, was now over the southwest United States, pushing a cold front ahead of it. Meteorological and oceanographic conditions over the WLICS remained the same as during the two previous days.

As the extratropical cyclone over the central United States moved farther northward during the early hours of 20 March, the central portion of the attached cold front moved eastward. Circulation around an area of high pressure located over the Florida panhandle appeared to be responsible for maintaining the southern extreme of the cold front in its position over central Texas (Fig. 4.9b). The low pressure system over the southwest translated little. Winds over the WLICS remained predominantly southeasterly, but the speeds diminished and fluctuations were common (Fig. 4.8a, b). Tides remained the primary influence on water level and although the current speeds diminished, their set remained westward (Fig. 4.8c, d).

During 21 March the extratropical cyclone moved northward again, causing a weakening of the southern extremes of the cold front. Enough weakening occurred that in response to southeasterlies generated by the high pressure system over the northeastern Gulf, a stationary front formed inland of the Texas coast (Fig. 4.9c). The time series of wind stress and current from the WLICS (Fig. 4.8a, b, c, d) showed low fluctuating values for each of the components.
Figure 4.8. March time series of (a) cross-shelf wind stress, (b) alongshelf wind stress, (c) cross-shelf current, and (d) alongshelf current. Positive is offshore or eastward. Start time is 0000 16 March 1979. Distance between tick marks on horizontal axis is two days.
Figure 4.9. Sequence of daily weather maps (NOAA, 1979) for selected parts of the March frontal period showing the movement of frontal systems across the United States. The black square represents the location of the WLICS.
Over the next 24 hours the low pressure system located in the southwest moved northeastward and a cold front formed. At 1200 on 22 March (Fig. 4.9d) the cold front was located near central Texas. The stationary, now turned warm, front located inland of the northern Gulf Coast remained near its position of the previous day. A steadying of the wind stress during the early part of the day and a drop later characterized the wind stress values in advance of the cold front (over the WLICS). The current also became steady toward the west (Fig. 4.8a, b, c, d).

A rapid wind shift to northwesterlies occurred late 22 or early 23 March indicating the passage of the cold front over the WLICS. Wind stress values approached 2.0 dynes cm\(^{-2}\) during this period (Fig. 4.8a, b). The daily weather map of 23 March indicates frontal positions at 1200 hours (Fig. 4.9e). In effect the low pressure system caused two fronts to form, one associated with the cyclone and a second which formed as the warm front moved south. The current shifted to onshore for a brief period before becoming primarily a northeastward current (Fig. 4.8c, d). Water level also responded to the frontal passage. Although the tidal signal remained apparent, a noticeable drop in the water level occurred.

The second frontal passage occurred shortly after noon on 23 March and a rapid eastward movement was indicated. Cold air pushed into the region behind the front by the low pressure system located over the midwest, continued to move over the WLICS from the northwest for the next 48 hours (Figs. 4.8a, b, and 4.9f). This circulation maintained the northeastward currents and the lowering of water level for an even longer period (Fig. 4.8c, d). After 1200 on 26 March another cold front (not shown) approached the Gulf coast initiating southeasterly winds which lasted for the remainder of the period. The atmospheric circulation eventually caused the current to reverse and become westerly.
Conditions during the March period were similar to those found during the January cold front passages. The sharp reversals in wind direction accompanying the frontal passages and predominantly alongshelf wind stress values were comparable to January conditions. Although the water level response during March was not as strong as that experienced during January, it was significant. Current response to the wind reversals accompanying frontal passages with predominantly alongshelf wind stress was also similar during the two monthly time series examined (Fig. 4.10).

April.

The final period began on 6 April. An extratropical cyclone which was located east of the Great Lakes trailed a cold front down the east coast of the United States and then westward across Mississippi, Arkansas, and Oklahoma. A high pressure system behind the cold front brought polar air into the central United States. Winds along the northwestern Gulf coast were low during these first two days of the period. Wind stress values from the WLICS indicated that the cross-shelf component increased up to 2.0 dynes cm$^{-2}$ in the onshore direction during these two days (Fig. 4.11a). The alongshelf wind stress varied from easterly to westerly, but the values were low (Fig. 4.11b). Flow during the period was predominantly westward. Current speeds to the west exceeded 25 cm sec$^{-1}$ on several occasions. After the first few hours of the period cross-shelf flow became very low and offshore (Fig. 4.11c, d).

An extratropical cyclone that formed in the lee of the Canadian Rockies crossed the western United States quickly during this period and at 1200 on 8 April, was located over Iowa (Fig. 4.12a). A cold front trailed to the southwest. Intensification of the cyclone occurred during
Figure 4.10. Progressive vector diagram for March frontal period showing inferred Lagrangian movement over the WLICS.
Figure 4.11. April time series of (a) cross-shelf wind stress, (b) alongshelf wind stress, (c) cross-shelf current, and (d) alongshelf current. Positive is offshore or eastward. Start time is 0000 6 April 1979. Distance between tick marks on horizontal axis is two days.
Figure 4.12. Sequence of daily weather maps (NOAA, 1979) for selected parts of the April frontal period showing the movement of frontal systems across the United States. The black square represents the location of the WLICS.
the next 24 hours and by noon of 9 April, it was located over West Virginia. The daily weather map indicates that the front stretched to the southwest through New Orleans (Fig. 4.12b).

During 8 April southwesterly winds, which were relatively strong, were present for a period of approximately 12 hours (Fig. 4.11a, b). The frontal passage was characterized by a very rapid shift to northeasterly winds. Values of wind stress were not very high (less than 1.0 dyne cm$^{-2}$) and combined with the rapid directional shift, suggest a weak frontal passage. The time series of current components and water level shows that there is little or no response by the water column to the weak frontal passage (Fig. 4.11c, d).

During 10 April the cold front stalled and became stationary south of the WLICS (Fig. 4.12c). A low pressure system moved past the Rockies during this period, but had no effect on the weather over the WLICS as yet. The wind stress over the the WLICS underwent a shift during this day, becoming strongly south-southeasterly at 2 dynes cm$^{-2}$ (Fig. 4.11a, b). Cross-shelf flow did not show a strong response to the high wind stress values, but the alongshelf current did. In fact there was a strong visual correlation between onshore wind stress and a westward current (Fig. 4.11a, d). The alongshelf current increased to 30 to 40 cm sec$^{-1}$ during this period. The alongshelf wind stress was not negligible (1.0 to 2.0 dynes cm$^{-2}$) and may have been involved in the dynamics of the water column (Fig. 4.11b), but it is unlikely that the alongshelf wind stress was responsible for the entire alongshelf current.

The stationary front remained over the northern Gulf Coast during the following day (11 April). Cross-shelf wind stress (Fig. 4.11a) increased further to 3.5 to 4.0 dynes cm$^{-2}$ in the onshelf direction. The
high cross-shelf wind stress was maintained for about four hours, then fell steadily to near zero early on 12 April. As the cross-shelf wind stress reached its peak, the alongshelf wind stress decreased to near zero values and remained there for much of the day (Fig. 4.11b). The cross-shelf component of the current began to fluctuate over a range of 20 cm sec\(^{-1}\), from 10 cm sec\(^{-1}\) onshore to 10 cm sec\(^{-1}\) offshore (Fig. 4.11c).

Alongshelf currents remained westward at greater than 40 cm sec\(^{-1}\) during much of 11 April (Fig. 4.11d), before decreasing to about half that value late in the day. Fluctuations in the alongshelf current did occur during the increase in speed of 11 April, but the rise was steady. The time series of the water level (not shown) shows that although the tidal characteristics of water level were clearly observed, a rise in the mean water level of approximately 30 centimeters had occurred over a period of about one day. Visual correlation between the water level record and the cross-shelf wind stress suggests that the two were strongly related from the time of the onshore shift in the wind stress.

The daily weather map from 12 April shows that the extratropical cyclone, which was located over the Rockies on 10 April, was now positioned over the Midwest (Fig. 4.12d). The cold front trailing from it became a stationary front over the eastern portion of Texas, evidently in response to the warm moist air being pushed northward by the strong southerly winds. Both components of the wind stress reacted as though a weak front was passing over the WLICS (Fig. 4.11a, b). Cross-shelf wind stress increased in the onshore direction for a brief period, then fell and eventually became directed offshore. Although the alongshelf wind stress was low, the pattern of the wind shift from easterly to westerly appears to suggest a weak frontal passage. Wind stress values for both components remained at or below 1.0 dyne cm\(^{-2}\).
Cross-shelf current continued to oscillate between onshore and offshore flow with the ranges in speed becoming larger (Fig. 4.11c). One strong burst in both the onshore and offshore components occurred during this 24 hour period. The alongshelf current increased again during 12 April and reached its highest values recorded from the winter and spring seasons, approximately 65 cm sec\(^{-1}\) westward, (Fig. 4.11d). Within 12 hours after the maximum was reached the current slowed to 10 cm sec\(^{-1}\) and another 4 hours later, it was less than 5 cm sec\(^{-1}\). Neither component of wind stress was very strong during this period and it appears that the water column was reacting to a non-local phenomenon. Some possibilities will be discussed shortly.

The remainder of the April period was uninteresting with respect to frontal passages, but the record appears to suggest a change from winter conditions. Figure 4.12e, f, show that the stationary front broke up before or while crossing the WLICS. This is a pattern that occurs during spring; the cold fronts weaken and the seasonal wind patterns change. Combining the two components of the wind stress from 13 April to the end of the record appears to suggest the formation of a relatively strong sea breeze system over the WLICS. Westward wind stress values preceded onshore wind stress once each day indicating the clockwise rotation common to sea breeze systems. Wind stress values approaching 1 dyne cm\(^{-2}\) were common during the course of the record.

Oscillations in the cross-shelf component of the current are still the most common characteristic (Fig. 4.11c). After 15 April the oscillations decreased in magnitude and oscillated in a pattern more common of the semi-diurnal tide. The alongshelf component of the current built during the final four days of the period, approaching speeds of 65 cm sec\(^{-1}\) on 17 April, before slowing during the next 36 hours (Fig. 4.11d).
Comparison of the wind stress and alongshelf current indicate that the wind stress alone may not have been responsible for high westward currents over the WLICS. Very similar alongshelf current records resulted from two totally different wind events. Figure 4.13 indicates that freshwater runoff from the Atchafalaya and Mississippi rivers was high during the April period. Dominant southeasterly winds during this period held the freshwater against the Louisiana coastline and the easterly wind component tended to push the current westward. The large volume of freshwater held against the coast would also set up a pressure gradient which would accelerate the water column.

**Alongshelf Momentum Balance**

The descriptions of the cold front passages and the response of the waters over the WLICS just presented reveal, qualitatively, the different properties of each cold front and the varying responses of the water column. In order to quantitatively assess each period with reference to water column dynamics and the part that the cold front winds play, the momentum balance is examined. The four frontal periods presented in Table 4.1 are used. In this case, however, the data has been filtered to remove diurnal and higher frequencies and resampled at one-hour intervals.

The alongshelf momentum balance may be estimated by determining the terms in the vertically-averaged alongshelf momentum equation

\[
\frac{dv}{dt} = -fu - g\frac{\partial h}{\partial y} + \frac{\tau_{wy}}{h} - \frac{\tau_{by}}{h}
\]

where \(u\) and \(v\) are vertically averaged cross-shelf and alongshelf current, respectively, \(f\) is the Coriolis parameter, \(g\) is the acceleration due to gravity, \(\frac{\partial h}{\partial y}\) is the alongshelf surface slope, \(h\) is the water depth, \(\tau_{wy}\) is
Figure 4.13. Runoff from the Mississippi and Atchafalaya rivers during 1978-1979 (Hann and Randall, 1981).
the alongshelf surface wind stress, and \( \tau_{by} \) is the alongshelf bottom stress. Due to high wind values accompanying cold front passages, a homogeneous water column is assumed. All available data support the validity of this assumption. Because the purpose of the NOS investigation was to gather information about near-bottom conditions, the current from the 300 centimeter level must be used to represent the vertically-averaged velocity. This approximation appears reasonable for the alongshelf component of flow due to the tendency for the shallow water currents to follow the local isobaths and for the water column to be well mixed. Because flow is directed primarily alongshelf and the current generally deviates only a few degrees from being alongshelf, cross-shelf currents are small. Bottom boundary layer effects may cause the cross-shelf current at 300 centimeters to misrepresent the vertically averaged value when the boundary layer thickness is large. This fact must be considered when interpreting the results of this analysis.

Because of the configuration of the subsurface platforms and problems with obtaining concurrent data (see Chapter 3), it is not possible to calculate horizontal and vertical acceleration terms. Estimation of the Rossby Number, \( \text{Ro} = \frac{v}{fL} \), assuming \( v = 20 \text{ cm sec}^{-1} \), \( f = 7.29 \times 10^{-5} \text{ s}^{-1} \), and \( L = 20 \text{ km} \), yields a value of 0.14 which indicates that the convective accelerations over the WLICS are small and may be ignored. The alongshelf pressure gradient term is also omitted because of the lack of alongshelf data acquisition sites.

The vertically-integrated alongshelf momentum balance thus becomes

\[
\frac{\partial v}{\partial t} = -fu + \frac{\tau_{wy}}{h} - \frac{\tau_{by}}{h} \quad (4.2)
\]
where $\tau_{by}$ is calculated from $\tau_{by} = \rho_w C_D v_b |v_b|$, where $\rho_w$ is the density of seawater, $C_D$ is the drag coefficient ($1.2 \times 10^{-3}$), and $v_b$ is the velocity at 300 centimeters. If the balance suggested by Equation 4.2 is not complete, a residual term of the form

$$RES = -\frac{\partial v}{\partial t} - fu + \frac{\tau_{wy}}{h} - \frac{\tau_{by}}{h}$$ (4.3)

may be calculated. The residual term represents the combined contribution of pressure gradient and acceleration terms to the total momentum balance. The cross-shelf momentum balance is assumed to be quasi-geostrophic (Pettigrew 1981).

The momentum terms in Equation 4.3 are correlated by pairs. The linear correlation coefficient, $R$, and phase information which result provide the data from which a more complete characterization of the force balance can be made. The results for the four winter periods described earlier in this chapter are presented in Table 4.2.

A strong correlation exists between the wind and bottom stress terms during the January frontal period. The positive correlation means that the alongshelf wind stress is opposed by bottom friction. Figure 4.14 shows that the magnitudes of these two components are similar and that the bottom stress lags the surface wind stress. Significant positive correlation between the alongshelf components of wind stress and the temporal acceleration suggests a significant transfer of momentum into the water column, accelerating it. Comparison of the magnitudes of the alongshelf wind stress and temporal acceleration terms shows that the wind stress term is consistently greater (Figure 4.14). This balance is not expected to be a major contributor to the momentum balance.

The correlation between the Coriolis term and alongshelf wind stress is significant. The negative relationship is out of phase with
Figure 4.14. Magnitudes of the alongshelf momentum balance terms during the January frontal period. Figure a) shows the magnitude of the temporal acceleration and Coriolis force, and b) shows the magnitudes of the wind stress, bottom stress, and residual terms. The start time is 0000 16 January 1979 (GMT) and each tick mark along the horizontal axis represents 2 days.
Table 4.2
Correlation Values Between Alongshelf Momentum Terms

<table>
<thead>
<tr>
<th></th>
<th>fu</th>
<th>( \tau_w/h )</th>
<th>( \tau_b/h )</th>
<th>RES</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>January</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>( \partial v/\partial t )</td>
<td>-0.229</td>
<td>0.504*</td>
<td>-0.002</td>
<td>-0.626*</td>
</tr>
<tr>
<td>( fu )</td>
<td></td>
<td>0.649*</td>
<td>-0.708*</td>
<td>-0.210</td>
</tr>
<tr>
<td>( \tau_w/h )</td>
<td></td>
<td>0.738*</td>
<td>0.738*</td>
<td>0.154</td>
</tr>
<tr>
<td>( \tau_b/h )</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>February</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>( \partial v/\partial t )</td>
<td>-0.452</td>
<td>0.729*</td>
<td>0.008</td>
<td>-0.001</td>
</tr>
<tr>
<td>( fu )</td>
<td></td>
<td>-0.363</td>
<td>0.054</td>
<td>-0.774*</td>
</tr>
<tr>
<td>( \tau_w/h )</td>
<td></td>
<td>0.100</td>
<td></td>
<td>-0.237</td>
</tr>
<tr>
<td>( \tau_b/h )</td>
<td></td>
<td></td>
<td></td>
<td>0.189</td>
</tr>
<tr>
<td><strong>March</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>( \partial v/\partial t )</td>
<td>-0.523*</td>
<td>0.583*</td>
<td>0.087</td>
<td>-0.226</td>
</tr>
<tr>
<td>( fu )</td>
<td></td>
<td>-0.851*</td>
<td>-0.674*</td>
<td>-0.412</td>
</tr>
<tr>
<td>( \tau_w/h )</td>
<td></td>
<td>0.670*</td>
<td></td>
<td>-0.014</td>
</tr>
<tr>
<td>( \tau_b/h )</td>
<td></td>
<td></td>
<td></td>
<td>0.573*</td>
</tr>
<tr>
<td><strong>April</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>( \partial v/\partial t )</td>
<td>0.384</td>
<td>0.723*</td>
<td>-0.025</td>
<td>-0.694*</td>
</tr>
<tr>
<td>( fu )</td>
<td></td>
<td>0.442</td>
<td>0.324</td>
<td>-0.567*</td>
</tr>
<tr>
<td>( \tau_w/h )</td>
<td></td>
<td>0.161</td>
<td></td>
<td>-0.810*</td>
</tr>
<tr>
<td>( \tau_b/h )</td>
<td></td>
<td></td>
<td></td>
<td>0.313</td>
</tr>
</tbody>
</table>

*Significant at 95 percent level.
normal Ekman transport to the right of the wind. The solution to this apparent contradiction lies in the correlation between the Coriolis force and bottom stress. Near-bottom Ekman dynamics are apparently active up to the 300 centimeter level during the January period, resulting in near-bottom cross-shelf flow to the left of the wind stress. As was the case with the acceleration term, the Coriolis force is consistently smaller than the wind and bottom stress terms (Fig. 4.14), suggesting that the balances between the Coriolis and stress terms are not as important as the frictional balance.

Inclusion of the January residual term somewhat complicates the momentum balance. A significant correlation between the temporal acceleration term and the residual term is of the proper phase to suggest that the temporal acceleration of the water column either sets up a pressure gradient to oppose it and/or is balanced by the horizontal convective acceleration terms. However, the significant negative correlation between alongshelf wind stress and the alongshelf residual term is unusual, especially if the residual term represents the alongshelf pressure gradient. In that case the relationship should be positive if the wind stress accelerates the water column and builds the pressure gradient. Although the residual term is generally smaller than the alongshelf wind stress and bottom stress terms, Figure 4.14 shows that the three terms together constitute a near-frictional balance.

The dynamical balance in the water column during the January period may be summarized as follows: Wind stress accelerates the water column and a frictional balance results within the water column. Momentum from the wind stress is balanced by friction at the bottom. The local acceleration and residual terms appear to be secondarily important to the
momentum balance. Cross-shelf flow (tied to the alongshelf momentum balance through the Coriolis term) appears to be associated with near-bottom Ekman dynamics. Within a bottom Ekman layer the cross-shelf flow is to the left of the prevailing current.

In my analysis, the cross-shelf component of flow at 300 centimeters is used as an estimate of the vertically-averaged cross-shelf speed. This is probably a valid assumption if the water column moves alongshelf as a slab with a thin bottom boundary layer (Frey 1981). Murray et al. (1982) also show that nearshore currents often meander over areas like the WLICS. Finally, in the presence of even modest stratification, vertically-sheared cross-shelf flow can occur. Thus, cross-shelf currents are difficult to interpret and conclusions about the alongshelf momentum balance must be tempered with this knowledge.

The March frontal period is quite similar to January. As was pointed out in the earlier portion of this chapter, winds associated with frontal passages during these two periods are predominantly cross-shelf, and the water column response is expected to be similar. A strong correlation between the alongshelf wind stress and bottom stress terms is present (Fig. 4.15). Correlation between wind stress and the temporal acceleration is also significant. The wind stress accelerates the water column to the bottom and the bottom stress is generated in a bottom boundary layer. The Coriolis and stress terms are correlated, indicating that cross-shelf flow within the bottom boundary layer (which extends to include the 300 centimeter level) is to the left of the prevailing wind stress and accelerating water column and that the Ekman dynamics are important.
Figure 4.15. Magnitudes of the momentum balance terms during the March frontal period. The start time is 0000 16 March 1979 (GMT) and each tick mark along the horizontal axis represents 2 days.
During the frontal passage the stress terms, the Coriolis force, and the residual term constitute the major portion of the momentum balance (Fig. 4.15). The importance of the residual term during the March period is greater than during the January period. A balance between the residual term and the bottom stress seems to exist. If the residual term is interpreted as the pressure gradient, this balance would imply a pressure gradient which does not depend on the local wind, but it is balanced by the bottom stress.

Like the January period estimation of the alongshelf momentum balance indicates that a partial frictional balance is present within the water column. Acceleration of the water column helps to generate a bottom boundary layer where Ekman dynamics are important to cross-shelf flow. The residual term begins to acquire added significance during this period. The Atchafalaya flood of 1979 began during March and the importance of the residual term (pressure gradient) may be related to the freshwater discharge.

Frontal characteristics during the February period were shown to differ substantially from the January and March periods, with cross-shelf wind stress being dominant. Correlations between the momentum terms show that the water column dynamics differ also. Significant positive correlation between the temporal acceleration term and the alongshelf wind stress suggests acceleration of the water column by the alongshelf wind stress. The time series of the two components (Fig. 4.16) reveals that the components are not only of the same magnitude, but they vary together throughout the period. Compared to the January and March periods, the magnitudes are quite small. Alongshelf wind stress values during the period are not high enough to accelerate the water column to the point
Figure 4.16. Magnitudes of the momentum balance terms during the February frontal period. The start time is 0000 12 February 1979 (GMT) and each tick mark along the horizontal axis represents 2 days.
where a frictional balance between the wind stress and bottom stress terms exists.

Inclusion of the residual term results in one significant correlation and that is with the Coriolis term. The negative phase relationship is such that cross-shelf flow would be to the right of forcing by a pressure gradient, if the residual term is interpreted as such. As shown in Figure 4.16, both terms are of about the same magnitude. The importance of the residual term suggests a geostrophically-balanced cross-shelf flow. Cross-shelf wind stress may set up alongshelf pressure gradients which influence cross-shelf flow.

During April significant positive correlation exists between the alongshelf wind stress and the temporal acceleration term. As was the case during each of the other periods, acceleration of the water column by the alongshelf wind stress is inferred. During the first half of the April period, the magnitude of the temporal acceleration term is generally less than one-half the value of the alongshelf wind stress term (Fig. 4.17). These two terms do not represent the total momentum balance within the water column, especially during the latter half of the period, when both components are very small. From the time series of momentum terms it is apparent that the alongshelf wind stress is not as strong as it was during the January and March periods. Although greater than February values, the alongshelf wind stress is not strong enough to generate a frictional balance.

In order to complete the April momentum balance, the residual term must be included. Figure 4.17b shows that the residual term is of the same magnitude or larger than the alongshelf wind stress term during much of the period. Linear correlations with the residual term are significant
Figure 4.17. Magnitudes of the momentum balance terms during the April frontal period. The start time is 0000 6 April 1979 (GMT) and each tick mark along the horizontal axis represents 2 days.
for the temporal acceleration, Coriolis force, and alongshelf wind stress. Because this period corresponds to the time of maximum flooding on the nearby Atchafalaya River, it is likely that alongshelf pressure gradients, both baroclinic and barotropic, contribute greatly to the residual term. A barotropic pressure gradient caused by southeasterly winds which holds the large freshwater runoff against the coast near the Atchafalaya River (a positive pressure gradient), would induce a temporal acceleration toward the west (a negative temporal acceleration). The temporal acceleration would increase the westward currents over the WLICS without a strong alongshelf wind stress. Likewise, a salinity gradient from fresher water in the east (Atchafalaya Bay) to more saline water in the west, would also induce westward currents.

The significant correlation between the alongshelf wind stress and the residual term appears to be the result of the correlation between the wind stress and temporal acceleration and the temporal acceleration and the residual terms. The significant negative correlation between the Coriolis force and residual terms is similar to the relationship found during February, when the cross-shelf wind stress is dominant. Cross-shelf wind stress may also contribute to setting up alongshelf pressure gradients (Chuang and Wiseman 1982) and the cross-shelf currents may result partially from the associated Coriolis accelerations.

Summary

Winter cold front passages cause many changes in the waters over the west Louisiana inner continental shelf. Chief among these is the response of the water column to the accompanying wind stress. Frontal characteristics differ somewhat during this period and so do the dynamics of the water column.
Four periods of cold front passage during the winter of 1979 have been examined. From these periods three different water column responses are described. During the January and March periods the fronts are closely associated with extratropical cyclones. Winds ahead of the front are southeasterly until the front is very near, when the winds shift to the southwest and parallel the front. Northwesterly winds follow the frontal passage and continue until another system approaches which is strong enough to push the old system out. Winds then shift through north and northeast rapidly and finally become southeasterly.

The currents over the WLICS follow the alongshelf wind stress. Prior to frontal passage currents are driven westward and slightly offshore by the southeasterly wind. As the front moves through the region, currents reverse abruptly and flow eastward until southeasterlies drive them westward again. There appears to be little net transport associated with this type of frontal passage.

Dynamically, during the frontal period the water column is accelerated by the alongshelf wind stress and a bottom frictional layer is generated. Within the bottom layer Ekman dynamics cause the cross-shelf flow to be to the left of the prevailing alongshelf current. The residual term which contains contributions from both the spatial acceleration terms and the pressure gradient term does not appear to be of primary importance to water column dynamics. A frictional balance (wind stress balanced by bottom stress) governs the dynamics of the water column over the WLICS during the January and March periods.

Frontal characteristics during the February period are unlike those of the preceding and following periods. In this instance a high pressure system located over central Canada pushes cold, arctic air down the
Mississippi Valley and into the Gulf of Mexico. The front parallels the
coast as it passes over it and associated winds remain northerly for
several days. There is little alongshelf wind stress associated with the
frontal passage.

The response to the frontal passage is largely in the cross-shelf
direction. Analysis of the alongshelf momentum balance indicates that
the alongshelf wind stress, though small, acts to accelerate the water
column. The acceleration of the water column is not large enough to
generate enough bottom friction to cause a bottom boundary layer 300
centimeters thick. Significant correlation between the Coriolis force
and the residual term probably is a manifestation of a longshore pressure
gradient generated by the cross-shelf wind stress; the sign of the cor­
relation infers flow to the right of the pressure gradient force.

The April 1979 period signals an end to the winter period. Two
stationary fronts pass rapidly over the WLICS and the associated wind
shifts are swift and weak. The remainder of the period is dominated by
southeasterly winds. April also marks the peak of the flood season on the
Atchafalaya River.

Dynamically this period is related to the February frontal period.
Though not as strong as during January and March, the alongshelf wind
stress accelerates the water column. No significant bottom boundary
layer is generated. Inclusion of the residual term adds to the under­
standing of the April dynamics. During this period of maximum flood on
the Atchafalaya River, the residual term probably represents the contri­
bution of barotropic and baroclinic pressure gradients generated by the
large freshwater output and the persistent southeasterly winds. Thus,
both alongshelf wind stress and pressure gradients force April circu­
lation over the WLICS.
CHAPTER 5

CONCLUSIONS

Forcing mechanisms and water column response over the west Louisiana inner continental shelf (WLICS) have been investigated at varying temporal and spatial scales. The major forcing mechanisms over the WLICS are 1) circulation over the northwestern Gulf of Mexico, 2) runoff from the Mississippi and Atchafalaya rivers, and 3) winds.

During the winter, large-scale cyclonic flow is observed over the Texas-Louisiana shelf. Generated by cooling due to a succession of cold front passages, the cyclonic flow contributes to the circulation of the anticyclonic gyre which dominates the western Gulf of Mexico. The southern limb of the cyclonic gyre, over the outer Louisiana shelf, merges with the northeastward flowing, northern limb of the anticyclonic gyre of the deep Gulf. On the inner shelf, the westward component of the gyre adds to the dominant wind-driven winter circulation.

Due to the proximity of the Atchafalaya and Mississippi rivers to the study site, seasonal fluctuations in freshwater discharge have a tremendous affect on the WLICS. Spring flooding introduces an enormous amount of freshwater into the inner shelf region. Density gradients cause some decoupling of the lower layer of the water column from the wind. During the spring the freshwater is held near the coast by southeasterly winds; barotropic and baroclinic pressure gradients and consequent strong westward current components result.
Winds also exhibit strong seasonal variability. Summer wind stress values are low, but they undergo a directional shift from southeasterly to predominantly southerly to southwesterly which causes a current reversal over the Texas and west Louisiana shelves. The reversal is a regional scale phenomenon. Slow meanders of the summer current field are superimposed on this pattern.

Autumn winds are stronger than those during summer. Weak frontal passages with a stronger cross-shelf component begin to pass through the area at a period of approximately 5 days. Fluctuations within the water column appear to respond to the alongshelf component of wind stress.

Cold front passages dominate the winter wind patterns. Winds ahead of a front are primarily southeasterly. Following the frontal passage, winds shift to northwesterly and increase in their intensity. As the front moves away the winds rotate clockwise back to southeasterly. Currents are shown to follow these wind patterns closely. The frontal systems occur at periods of just over three days and contain approximately equal components of alongshelf and cross-shelf wind stress. The alongshelf wind stress is found to be strongly coherent with the alongshelf current.

A more in-depth investigation of several periods of cold front passage in 1979 shows that the orientation of the winds accompanying the cold front plays a large role in determining the resultant water column dynamics. Although frontal characteristics during the four selected periods differ, strong wind stress values are one common trait.

Strong alongshelf wind stress accompanies each frontal passage during the January and March periods. Extratropical cyclones trailing cold fronts translate across the United States during these two periods.
Southeasterly winds dominate the interfrontal periods. An investigation of the water column dynamics during January and March reveals that the alongshelf wind stress accelerates the water column and generates a frictional boundary layer at the bottom. A frictional balance dominates the WLICS during the January and March periods.

During the February frontal period a strong Canadian high pressure system pushes a front across the WLICS from the north. Wind stress accompanying the frontal passage is primarily cross-shelf for several days. Dynamically, the low alongshelf wind stress accelerates the water column, but it is not strong enough to establish a frictional balance. There appears to be some influence on water column dynamics by the strong cross-shelf wind stress.

South-southeasterly winds and fresh water runoff from the Mississippi-Atchafalaya flood dominate the April period. Two weak frontal passages occur during this period and cause minimal water column response. The alongshelf wind stress accelerates the water column, but no frictional balance is established. However, the residual term acquires added importance during this period. Apparently the onshore winds hold the floodwaters from the Atchafalaya River against the coastline, setting up strong pressure gradients which accelerate the water column westward.

Although this dissertation adds considerably to the understanding of circulation over the west Louisiana inner continental shelf and in particular to the dynamics of cold front-water column interactions during winter, one negative outcome is the realization of the necessity for a regional-scale, long-term investigation in order to further interpret water column dynamics.
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APPENDIX A

Lanczos Filter Characteristics

The Lanczos filter is gaining increased acceptance in the physical oceanographic community. A useful filter is one which allows only the data of interest within a given frequency band to pass through the filter. In this dissertation the Lanczos filter is used to remove high frequency components from both the wind stress and currents in order to investigate slowly changing processes. Filter weights are calculated from the Lanczos kernel,

\[ R(n) = \cos \left( \frac{n\pi}{M} \right) \]

where \( n \) is the filter weight number and \( M \) is one-half the filter width. The curve calculated from the kernel is folded around the center point. A diagram of the Lanczos taper filter is presented in Figure A.1a. A filter width of five days (120 hours) and a half-power point of 40 hours were selected. The half-power point is the period where one-half of the amplitude of the time series passes through the filter.

The frequency response for the Lanczos filter is calculated from

\[ \Phi(\omega) = \sum_{n=-M}^{M} 2R(n) \cos(n\omega) \]

where \( \omega \) is radian frequency. Figure A.2b shows the frequency response of the Lanczos filter. The half-power point (amplitude response = 0.50) corresponds to a frequency of 0.025 cycles per hour or a 40 hour period. At the diurnal frequency, 0.042 cycles per hour, the amplitude response is
near zero, indicating that the filter effectively suppressed diurnal and higher frequencies. The filtered time series in this study are resampled at one-hour intervals and that data is used for all of the analyses in this dissertation.
Figure A.1. Characteristics of the Lanczos filter: a) the filter window in the time domain, and b) the frequency response of the filter.
Appendix B
Estimates of the Friction Coefficient and Alongshelf Pressure Gradient
over the West Louisiana Inner Continental Shelf,
Winter 1979

With the means available at the present time it is not possible to
measure accurately the alongshelf pressure gradient. Scott and Csanady
(1976) present a simple method by which the pressure gradient term and the
friction coefficient may be estimated from measures of the alongshelf
components of wind stress and near-bottom current. Their estimate of the
alongshelf pressure gradient in the water column off Long Island agrees
very well with geodetic leveling (Sturges 1974).

Scott and Csanady (1976) assume a frictional balance in the shallow
water (30 m) south of Long Island, New York. Bottom stress is charac­
terized by a linear friction coefficient (r) with the resulting along­
shelf balance in the form, appropriate to my data,

\[ \tau_w = \rho r v_b \]  

(B1)

where \( \tau_w \) is alongshelf wind stress, \( \rho \) is density, \( r \) is a friction co­
efficient with units of velocity, and \( v_b \) is the alongshelf velocity at 300
centimeters. A discussion of the linear friction parameterization is
presented by Winant and Beardsley (1980). Regression analysis is then
performed on the wind stress and near-bottom current pairs.

Linear regression results in an equation of the form
\[ \mathbf{T}_w = A v_b + \kappa \]  

where \( \mathbf{A} \) represents the slope of the regression line and \( \kappa \) is the intercept. \( \mathbf{A} \) may be related to the friction coefficient by \( \tau = \mathbf{A}/\rho \). The intercept is interpreted as an alongshelf pressure gradient and is given by \( \kappa = gh \frac{\partial \eta}{\partial y} \).

Tides have been found to alias the estimated value of the friction coefficient (Marmorino, 1982). Therefore, the low-frequency data from Chapter 4 (Table 4.1) is used in this analysis. The results of the linear regression on the data from each period are presented in Table B-1.

Linear friction coefficients during the January and March frontal periods are almost identical. These values are approximately twice as large as those presented by Marmorino (1982) for the west Florida shelf and an order of magnitude smaller than reported by Scott and Csanady (1976). The variability of the linear friction coefficients for these three locations may be due to tidal activity. Tides are greatest off Long Island and smallest on the west Florida shelf, with the west Louisiana shelf falling in between.

The February linear friction coefficient is approximately one-half the value of the January and March periods, and the April coefficient is one-half the February value. Lower friction coefficients occur during February and April because of the incompleteness of the frictional balance found to be occurring during these two periods (Chapter 4). Because the bottom stress is not an important part of the momentum balance, solving for the linear regression between the wind stress and near-bottom velocity results in erroneous information.
Table B.1

Linear Regression of the Filtered Alongshore Wind Stress, $\tau_w$, Current, $v_b$, and Pressure Gradient, $\partial \eta / \partial y$

<table>
<thead>
<tr>
<th>Month</th>
<th>$\tau_w = a v_b + b \partial \eta / \partial y$</th>
<th>$(\partial \eta / \partial y) \times 10^{-7}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>$0.052 v_b + 0.263$</td>
<td>2.7</td>
</tr>
<tr>
<td>March</td>
<td>$0.044 v_b + 0.065$</td>
<td>0.7</td>
</tr>
<tr>
<td>February</td>
<td>$0.021 v_b - 0.189$</td>
<td>-2.0</td>
</tr>
<tr>
<td>April</td>
<td>$0.009 v_b - 0.266$</td>
<td>-2.8</td>
</tr>
</tbody>
</table>

NOTE: The assumed regression equation is

$$ \tau_w = r v_b + \kappa $$

where $\tau_w$ is alongshore wind stress, $v_b$ is alongshore near-bottom current, and $\partial \eta / \partial y$ is the alongshore pressure gradient, which is linearly related to $\kappa$. 
Due to the incomplete frictional balance during February and April, the alongshelf pressure gradient values have little meaning. On the other hand, the alongshelf pressure gradients that result from the more complete frictional balances during January and March should be more accurate. Although both of the alongshelf pressure gradients indicate a slope down to the west, the January pressure gradient exceeds the March value by a factor of four. January values are comparable to those off Long Island, New York (Scott and Csanady, 1976).

The positive pressure gradient slope may be due to regional or local effects, and it may involve the barotropic or baroclinic component of the pressure gradient. Local changes in the pressure gradient occur on too small a scale to be measured using the platform configuration employed in this investigation, although the site's proximity to the mouths of the Calcasieu and Sabine rivers ensures that salinity gradients induced by freshwater discharge are important. On the regional scale, outflow from the Atchafalaya is expected to be important in setting up pressure gradients.

Estimates of the alongshelf baroclinic pressure gradient from the 1963-1965 Gus experiment (Temple et al., 1978) provide information about this regional gradient. During January 1965 the regional alongshelf baroclinic pressure gradient in the 7-meter water depths between Calcasieu Pass and Atchafalaya Bay was $2.07 \times 10^{-7}$. This is of the same order as the pressure gradient resulting from the linear regression analysis. In order to have a positive baroclinic pressure gradient, salinities must be higher at the Atchafalaya station. One must also assume that the barotropic component is zero, which may arise because of the mass distribution further offshore. The January rainfall peak and
relatively low Atchafalaya discharge shown in Chapter 2 are responsible for the higher salinities off Atchafalaya Bay.

The extremely high wind stress accompanying January and March frontal passages may also be active in setting up both baroclinic and barotropic pressure gradients on a local scale. Freshwater plumes may be rafted over large regions of the inner shelf and cause rapidly changing density gradients. Setup and setdown also occur, and alongshelf variations in the amount of setup or setdown may contribute to alongshelf pressure gradients.
VITA

Richard L. Crout was born on 24 November 1953 in Columbia, South Carolina. In 1972 he received an Academic Diploma upon graduation from Brookland-Cayce High School. He then entered the Marine Science program at the University of South Carolina and graduated cum laude with a B.S. degree in 1976. In the autumn of 1976 he began studies at Louisiana State University which led to a M.S. degree in Marine Sciences in 1978. His thesis was titled "Momentum Balance on a Shallow Shelf: Moskito Bank, Nicaragua." In 1979 he entered the Ph.D. program in Marine Sciences at Louisiana State University.
EXAMINATION AND THESIS REPORT

Candidate: Richard Larry Crout

Major Field: Marine Sciences

Title of Thesis: WIND-DRIVEN, NEAR-BOTTOM CURRENTS OVER THE WEST LOUISIANA INNER CONTINENTAL SHELF

Approved:

[Signature]
Major Professor and Chairman

[Signature]
Dean of the Graduate School

EXAMINING COMMITTEE:

Robert P. Randall

R. A. Kinney

S. O. Hurl

Stephen P. Murray

Date of Examination:

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