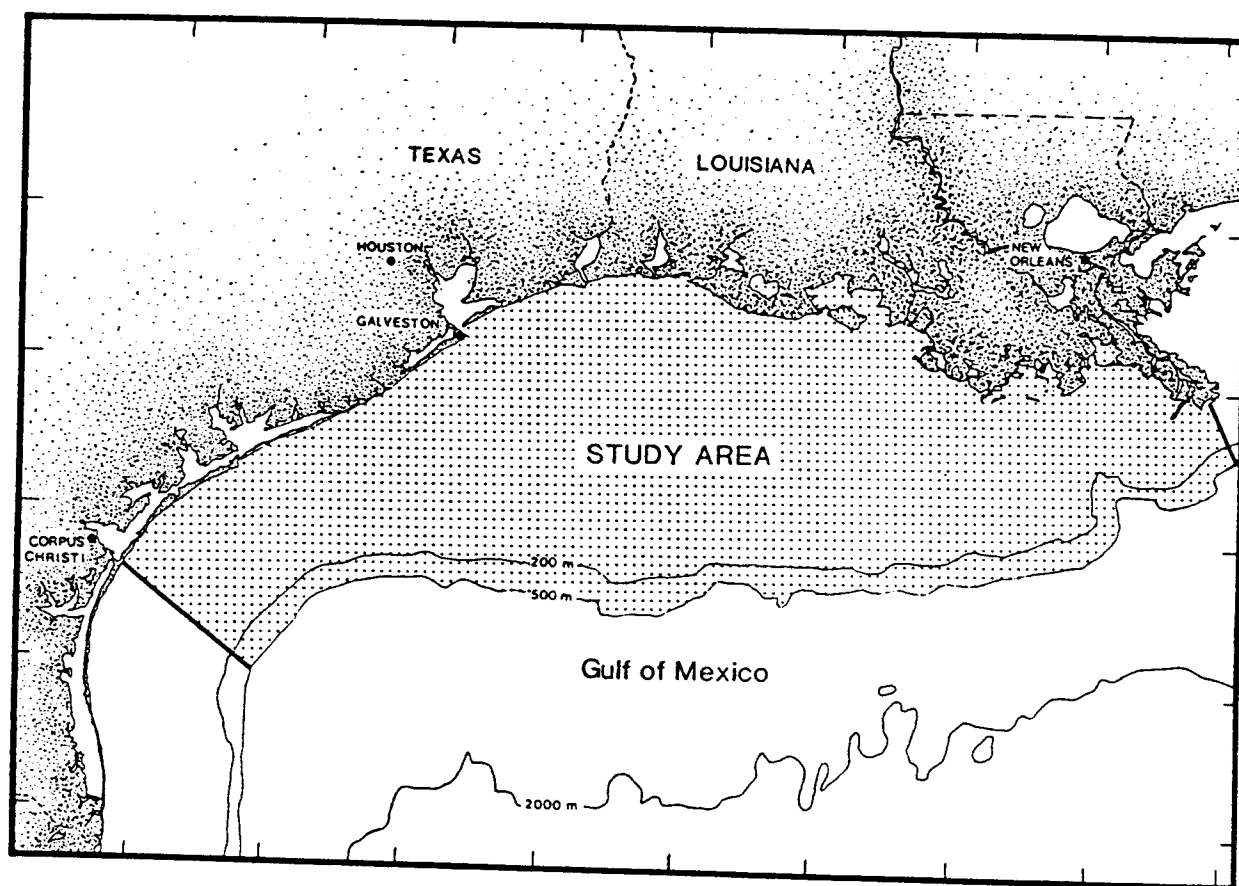


Offshore Texas and Louisiana Marine Ecosystems Data Synthesis

Volume II: Synthesis Report

Chapter 4 Only



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CHAPTER 4 PHYSICAL OCEANOGRAPHY AND METEOROLOGY

F.J. Kelly

4.1 INTRODUCTION

Each continental shelf region in the world has a different combination of meteorological, hydrological, astronomical, and deep ocean forces that act in concert with the local topography and geometry to produce a unique set of physical processes and water masses. The characteristics of the Texas-Louisiana shelf waters are determined, principally, by the large input of fresh water from one of the world's great rivers and an annual cycle of prevailing winds acting on a broad shelf with a coast that is arcuate, concave to the southeast (Figure 4.1). Astronomic tides, interaction with eddies and currents in the deep Gulf, and the momentum input and heat and mass exchange across the sea surface caused by intrusions of polar fronts and tropical cyclones, significantly modulate the effects of the principal factors, but do not obscure them.

Although various topics are discussed individually in this chapter, all of the topics are interrelated. Currents, the unifying theme, are presented later (Section 4.8). Some of the many biological, chemical, and sedimentary processes directly related to physical processes are also noted in this chapter, but the details of these interactions are discussed in other chapters.

Figure 4.1 shows the locations of geographic references and study sites from which data are used to illustrate points in the discussion. Bryan Mound, Big Hill, and West Hackberry are sites at which the Department of Energy's Strategic Petroleum Reserve (DOE/SPR) Office conducted environmental studies from 1977 through 1985 for its brine disposal operations. Section 1.1.2 summarizes the DOE/SPR study as well as other major environmental studies conducted off the Louisiana and Texas coasts. As Figure 1.1 indicates, the study area for this synthesis effort formally extends from the Mississippi Delta to Corpus Christi Bay. However, in this chapter some of the discussion includes results of studies conducted in the far southwest region of the shelf because its hydrography and circulation are integral parts of the physical processes on the shelf.

The Minerals Management Service (MMS) Gulf of Mexico Physical Oceanography Program: Year 5 Study, which includes hydrographic transects and an array of moorings along approximately 92°W from near the coast to mid-Gulf, is in progress at the time of this report. It is likely that the results of the Year 5 Study, will address some of the data gaps and questions about dynamic processes noted in Section 4.9, particularly those pertaining to interaction between the shelf and oceanic Gulf waters.

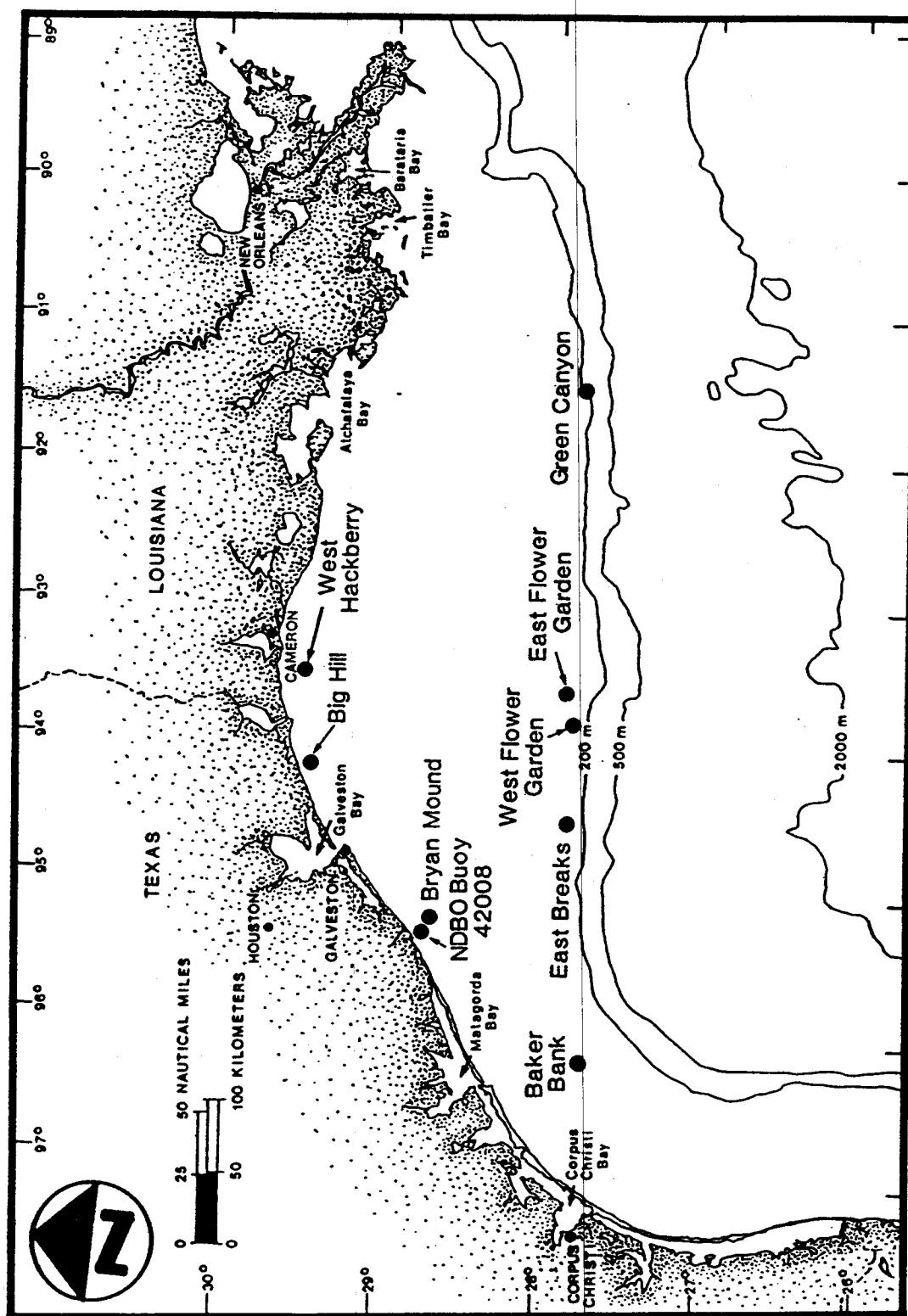


Figure 4.1. Chart of observational and geographic locations referenced in the text. Bryan Mound, Big Hill, and West Hackberry are brine disposal sites of the Strategic Petroleum Reserve.

4.2 METEOROLOGICAL CLIMATE

4.2.1 Air Masses

The following air masses either influence or control the weather over the shelf (Orton, 1964):

- Maritime tropical (mT) air comes from the southeast and south with the circulation around the southwestern extension of the Bermuda high. This air mass dominates the region from March through October and is responsible for the subtropical climate of the Gulf of Mexico.
- Maritime Polar (mP) air comes from the Pacific Ocean, and mP fronts cross the shelf most frequently from late fall through spring. During July and August, the mP fronts are unable to cross the Rockies and push southward through the hot southwest. Most mP fronts have an average orientation when they reach the coast of 40 to 50° from north and, because of the orientation of the coast, they enter the Gulf between Corpus Christi and Galveston (Henry, 1979).
- Continental Polar (cP) air masses push into the Gulf in all months, but most frequently from October through April. Cold fronts, especially cP types, can become quasi-stationary after moving out over the Gulf in winter, a situation favoring cyclogenesis with accompanying low ceilings and continuous rain. According to Henry (1979), the cP fronts (except summer) have an orientation of 60 to 65° from north and enter the Gulf east of Galveston and west of 90°W. Fronts entering the Gulf during the summer are usually cP. They are oriented east-west, because this is their southern-most penetration, and enter to the east of 90°W.
- Continental Arctic (cA) air is very cold and dry, thus the term "blue norther." Invasions of the Gulf by cA fronts are confined normally to the months of January and February. The orientation of cA fronts is similar to that of cP fronts.

From March through September the influence of the Bermuda high dominates the climate over the shelf (Franceschini, 1961). The anticyclonic high pressure system builds during the spring and summer, its western end extending into the eastern Gulf, and mean wind conditions shift gradually from southeasterly to more southerly (Figure 4.2). (Note: the wind roses in Figure 4.2 are constructed using the oceanographic convention for direction. However, in this discussion, meteorological terminology is used, i.e., a southerly wind blows from the south). In the fall the high weakens and the western end retreats from the Gulf. Winds return quickly to southeasterly. In October, an anticyclonic cell, separated from the Bermuda High, lies over northeastern Texas, resulting in more frequent easterly winds. During the winter, as intrusions of polar air masses move out across the shelf, the frequency of strong northerly winds increases. Because the shelf is sufficiently far south to remain under the influence of the Bermuda high,

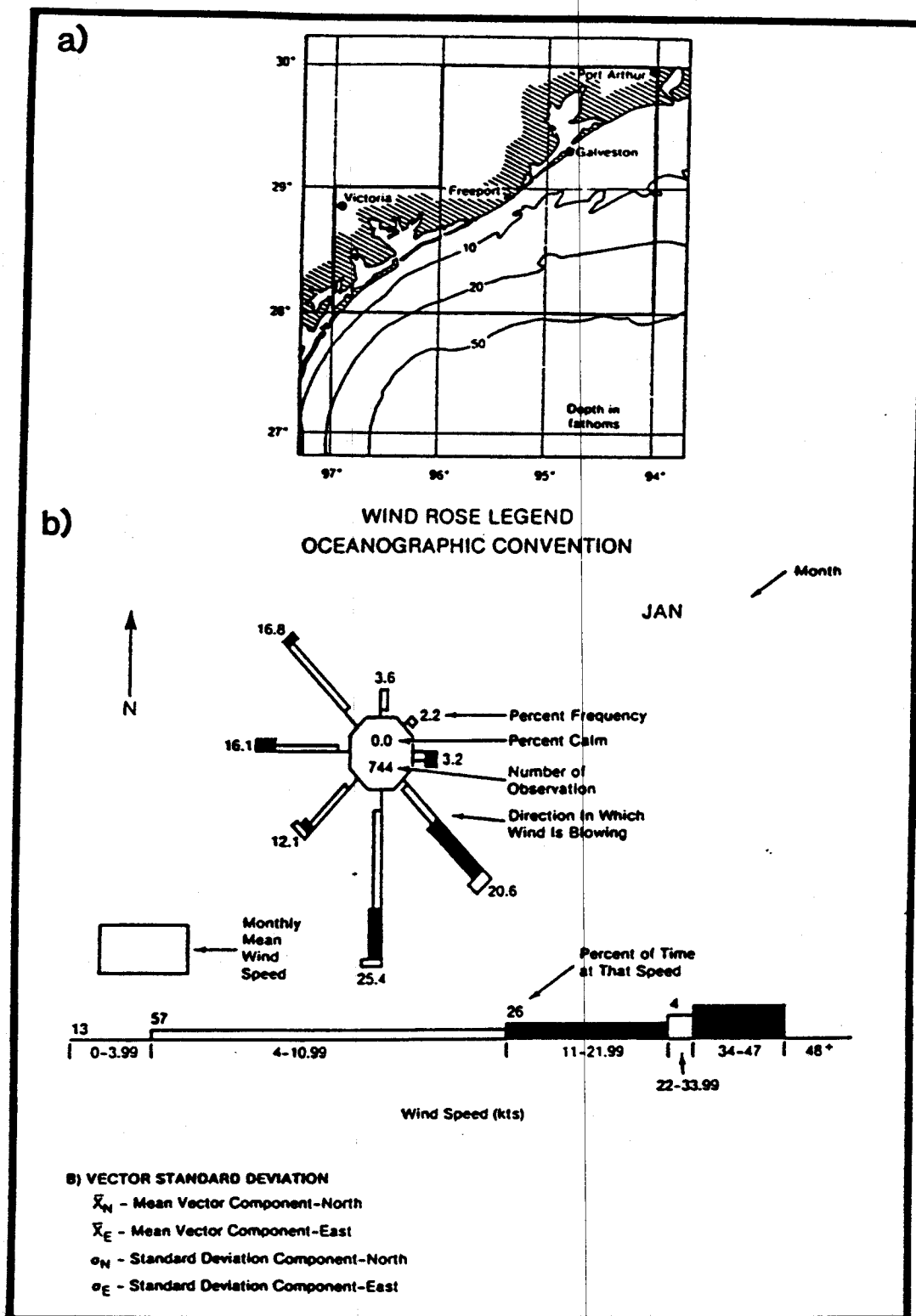
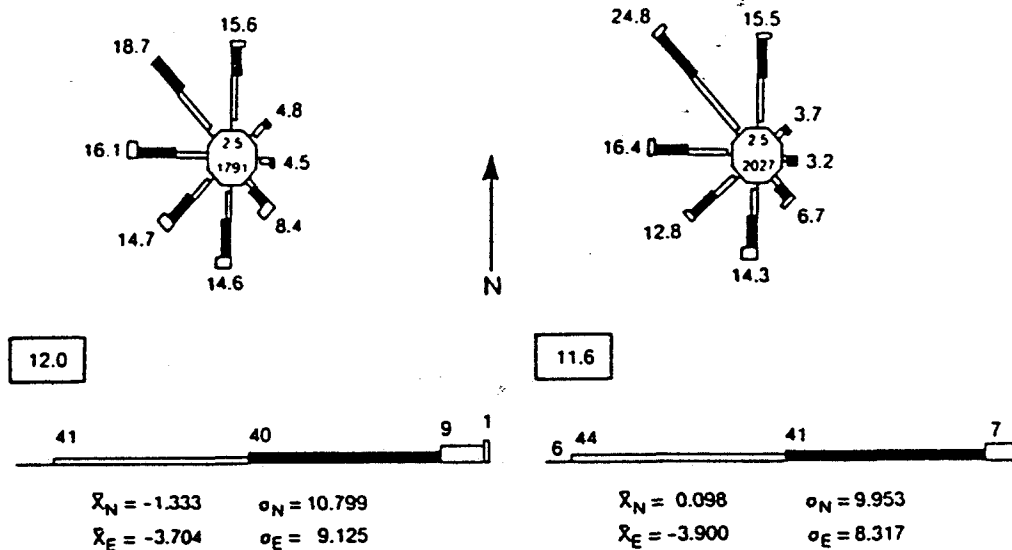


Figure 4.2. Summary of wind observations from transient ships for the period 1900 to 1970. Data were obtained from the Sea Surface Meteorological Observation data base of the NOAA National Climatic Center. a) Area of data collection. b) Wind rose legend. c) February-May data. d) June-September data. e) October-January data.

c)

FEB MEAN

MAR MEAN



APR MEAN

MAY MEAN

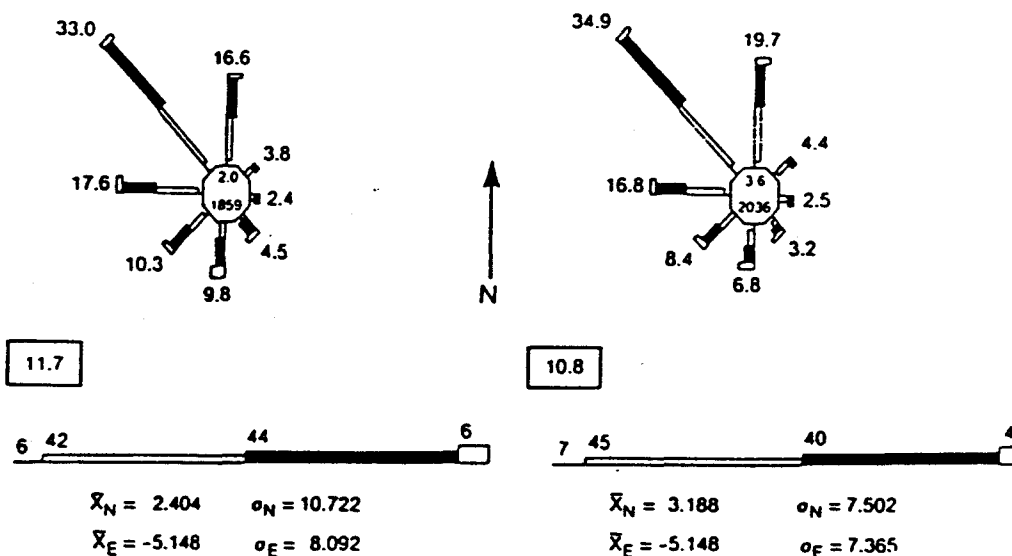


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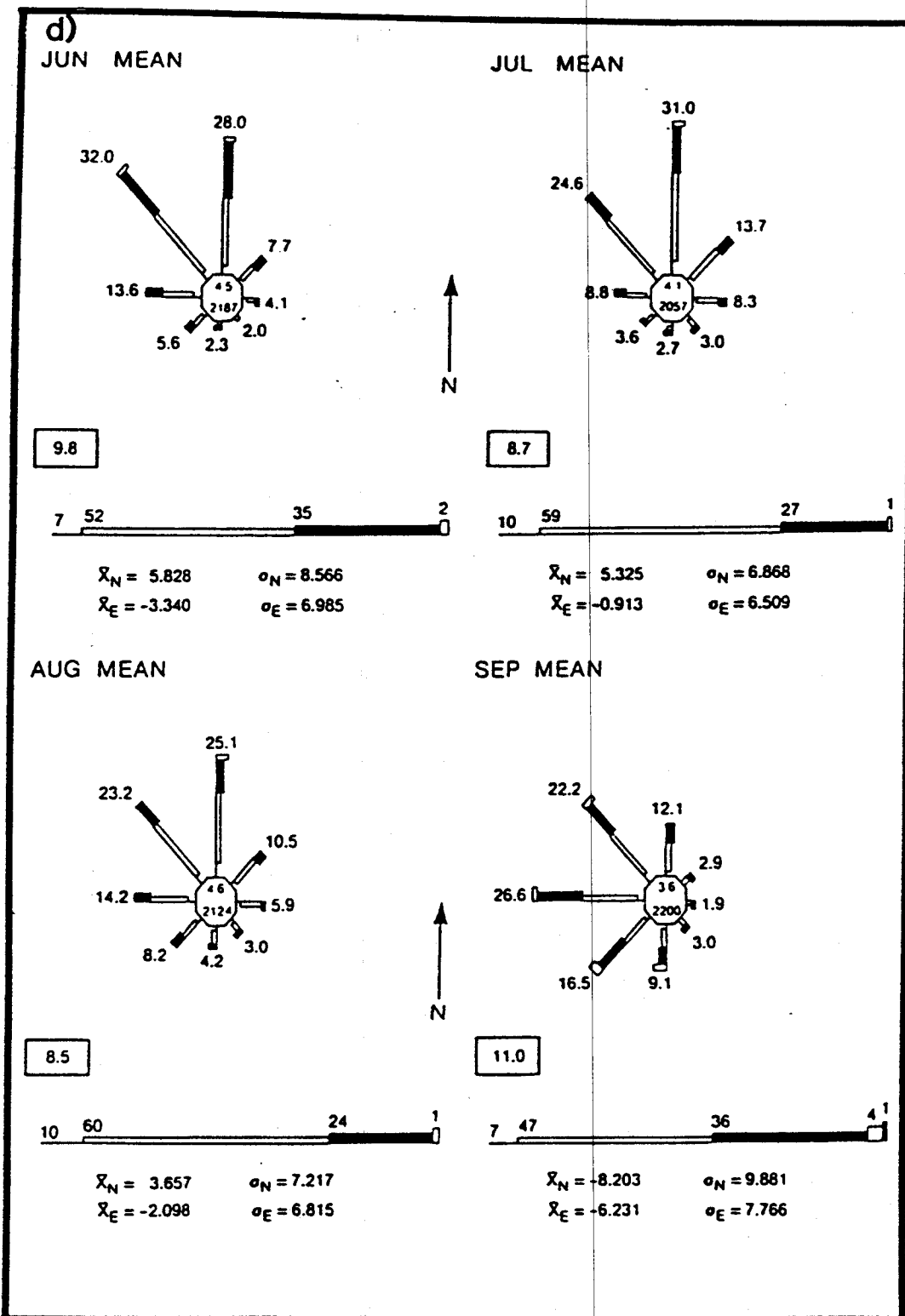


Figure 4.2. (Continued).

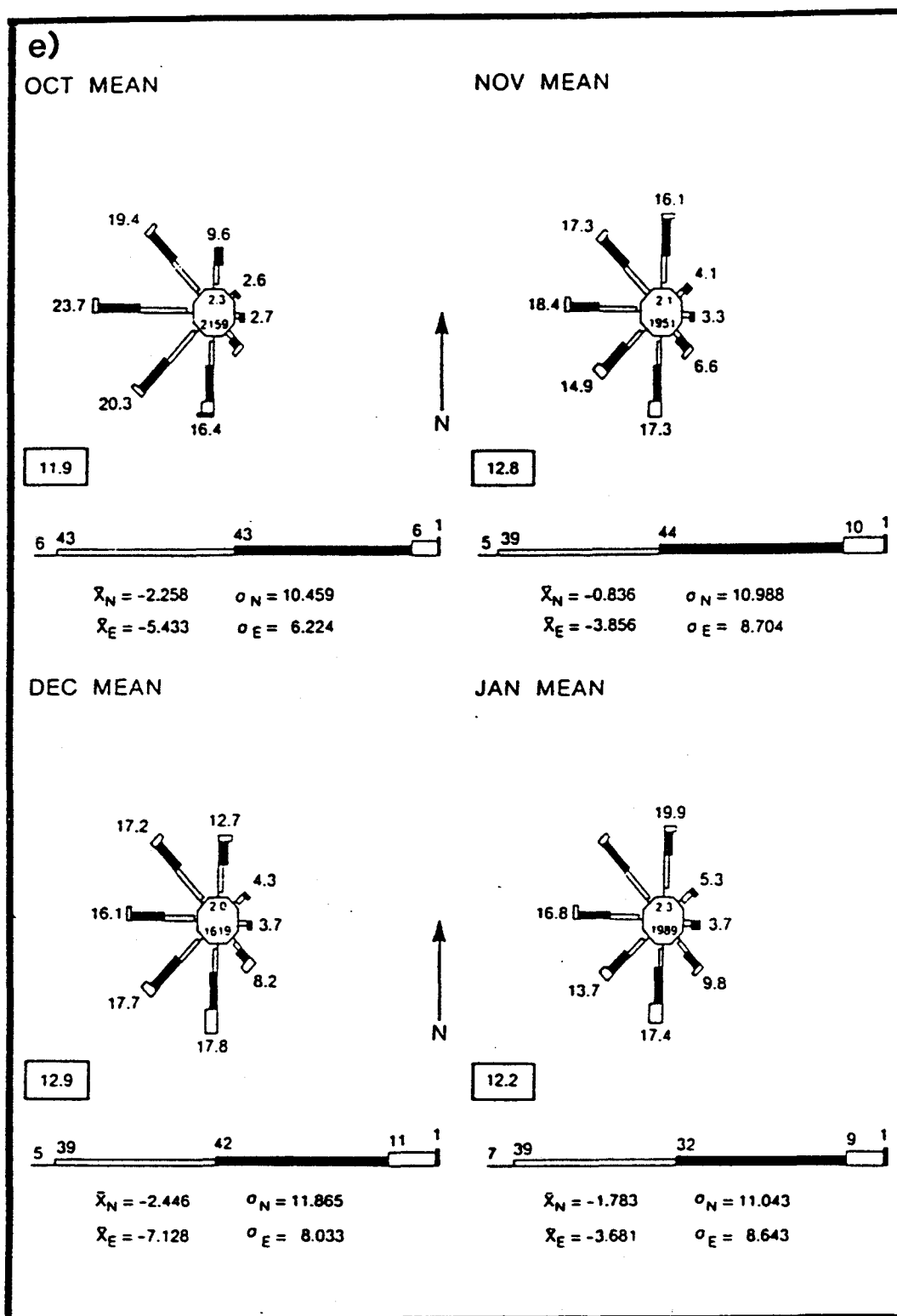


Figure 4.2. (Continued).

winds return to southeasterly after the passage of frontal systems. Therefore, during the winter months (November through February) northerly, northeasterly, easterly, southeasterly, and southerly winds occur with about the same frequency. Thus in Figure 4.2, which uses the oceanographic direction convention, the N, NW, W, SW, and S rose "petals" are about the same length.

4.2.2 Barometric Pressure

Average atmospheric pressure over the inner shelf (Figure 4.3) and outer shelf (Yamazaki and Herbich, 1985) is highest in winter because of continental high pressure systems; pressure averages about 1,019 mb in December. Mean pressure decreases in the spring as the equatorial trough migrates northward and the low pressure system over Mexico deepens. Minimum pressures of about 1,013 mb occur in May and June. In July there is a weak, relative maximum in atmospheric pressure, which Chew (1964) attributes to the maximum westward extension into the Gulf by the Bermuda-Azores high. In August and September there is a second minimum of about 1,014 mb, and then pressure rises as the equatorial trough migrates southward, the Mexican low pressure system fills, and the Bermuda high weakens and retreats.

4.2.3 Air Temperature

According to Orton (1964), mean monthly surface air temperatures during winter (December-February) are about 15.5°C over the inner shelf and 18°C over the outer shelf, and approximately 90% of the actual observations are within 6°C of the mean. For March, the means over the inner and outer portions of the shelf increase to 18°C and 21°C, respectively, and for April they are 21°C and 23°C. From May through September there is very little contrast in air temperature across the shelf from north to south. Also, variability decreases; 90% of the observations are within 3°C of the means. Mean values rise to 24.5°C for May, 28°C for June, reach a maximum of 29°C for July and August, and then decrease to 28°C for September. A north-south gradient reappears in the means for October, which are 24°C and 26°C, respectively, and variability increases. The values of mean air temperature for November are the same as for April.

4.2.4 Precipitation and Humidity

Quantitative measurements of the distribution of offshore precipitation do not exist. Based on ship reports (Orton, 1964), rain occurs most frequently in the Gulf in September. The "wet" season, July through September, is coincident with the season for easterly atmospheric waves and tropical storms. The "dry" season is March through May, with rain reported least frequently in May.

Dinnel and Wiseman (1986) estimated annual precipitation rates over the shelf, between the coast and the 91.5-m isobath and from the Mississippi River Delta to Brownsville, by extrapolating land-based coastal rainfall. They reduced the onshore values by 25% to account for the fact that the meteorological processes that produce rain result in greater precipitation toward the coast than farther offshore. Their

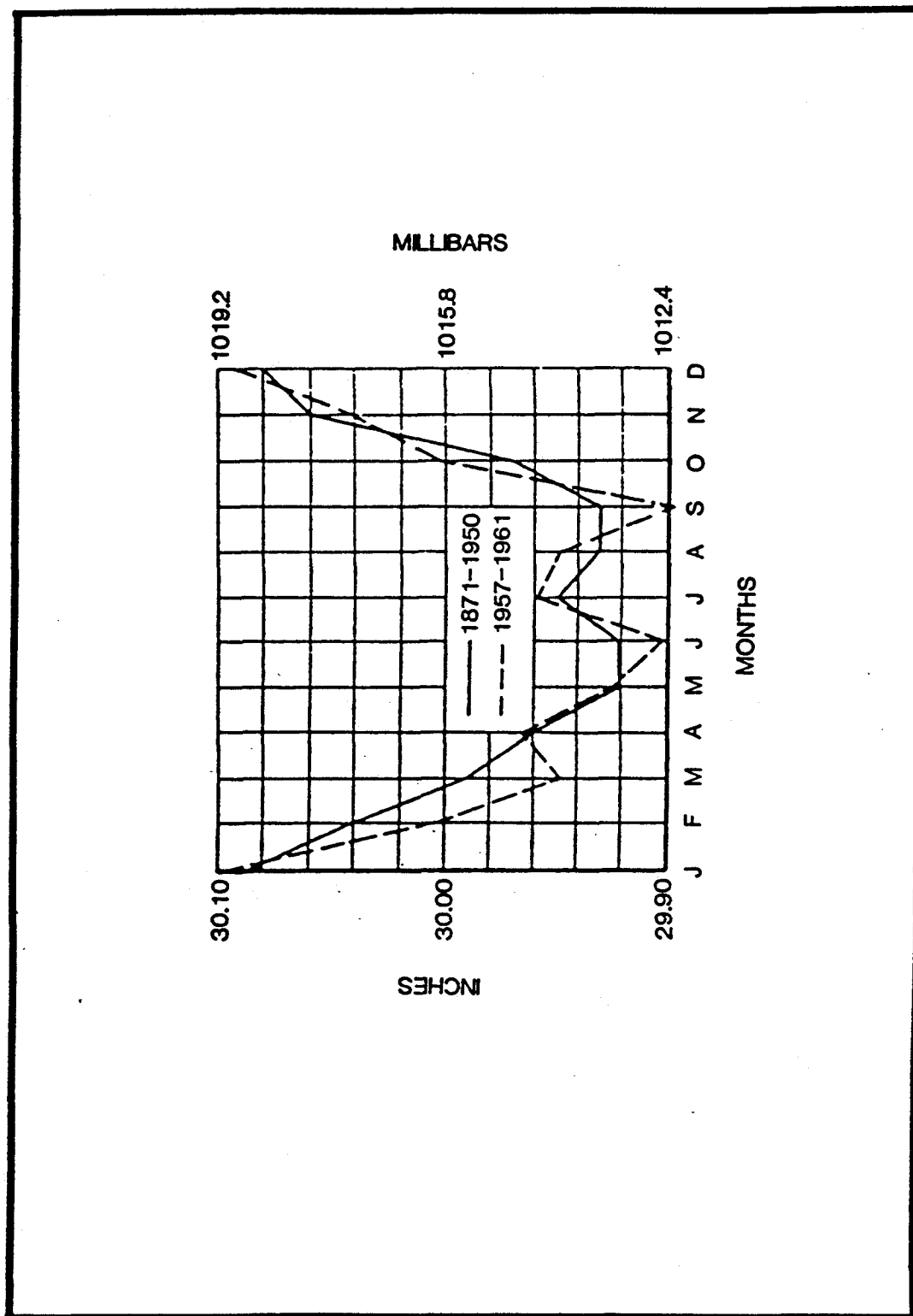


Figure 4.3. Monthly mean sea-level atmospheric pressure at New Orleans. Horizontal lines in legend are positioned at corresponding mean yearly pressures (From: Chew, 1964).

30-year mean value is $108 \text{ km}^3/\text{year}$, or, assuming a shelf area of $106,866 \text{ km}^2$, about $101 \text{ cm}/\text{year}$.

Fog is produced when warm, moist, Gulf air overrides cooler water. Advection-radiation fog, lasting about three or four hours, occurs most frequently from November through March in near coastal areas. Dense advection fogs (sea fogs) form out over the cold water surface during the winter months, December through February, and may persist for several days.

4.3 METEOROLOGICAL FORCING

4.3.1 Seasonal Variations of Wind Stress

The climate cycle described above produces an annual progression of wind stress patterns over the shelf that is responsible for much of the observed low-frequency circulation. Several estimates of the seasonal patterns of wind stress based on different sources of data are available for the Gulf on a 1° grid. For example, Elliott (1979) calculated his from all available ship observations up through 1972, and Rhodes et al. (1985) produced a wind set based on the Navy's twelve-hourly surface pressure analysis, which is available from 1967 to 1982. Figure 4.4 shows some results of the latter method. In the region of the Texas-Louisiana coast, the Navy Corrected Geostrophic Wind Stress data agree well with the observed wind stresses from coastal and nearshore meteorological stations. Figure 4.5 shows the observed monthly means (vectors) for Brownsville, Galveston, and the 5° quadrangle, 25° to 30°N , 90° to 95°W in the Gulf of Mexico. The sources of the data are given by Cochran and Kelly (1986a). The lines connecting the tips of the vectors form a "figure 8" and the chronological progression has the same sense.

Monthly mean alongshore components of wind stress are of considerable interest because of their importance to coastal currents (see Section 4.8). Figure 4.6 shows them for a number of fairly evenly spaced locations along the coast. The mean wind components are in the same direction along the coast over long distances. There is, however, a convergence in the wind components. It lies between Port Isabel and Port Aransas in January. It migrates upcoast in spring and early summer, reaching a position east of Calcasieu Pass in July. A downcoast retreat of the feature beginning in August is so rapid that by September it is, if it exists, south of Port Isabel. The convergence of the alongshelf component of wind results not from spatial change in the wind field so much as from the arcuate form of the Texas coast.

4.3.2 Frontal Passages

The maritime tropical atmosphere over the shelf may respond to cold fronts as much as 48 h before their passage. During this period, surface air pressure falls, the dew point rises, and temperature increases. Frontal passages across the shelf are accompanied by dramatic weather events such as gale-force winds, torrential rains and large temperature fluctuations. After the passage of cold fronts, about 36 to 48 h are needed for the air of polar origin to reach a new equilibrium

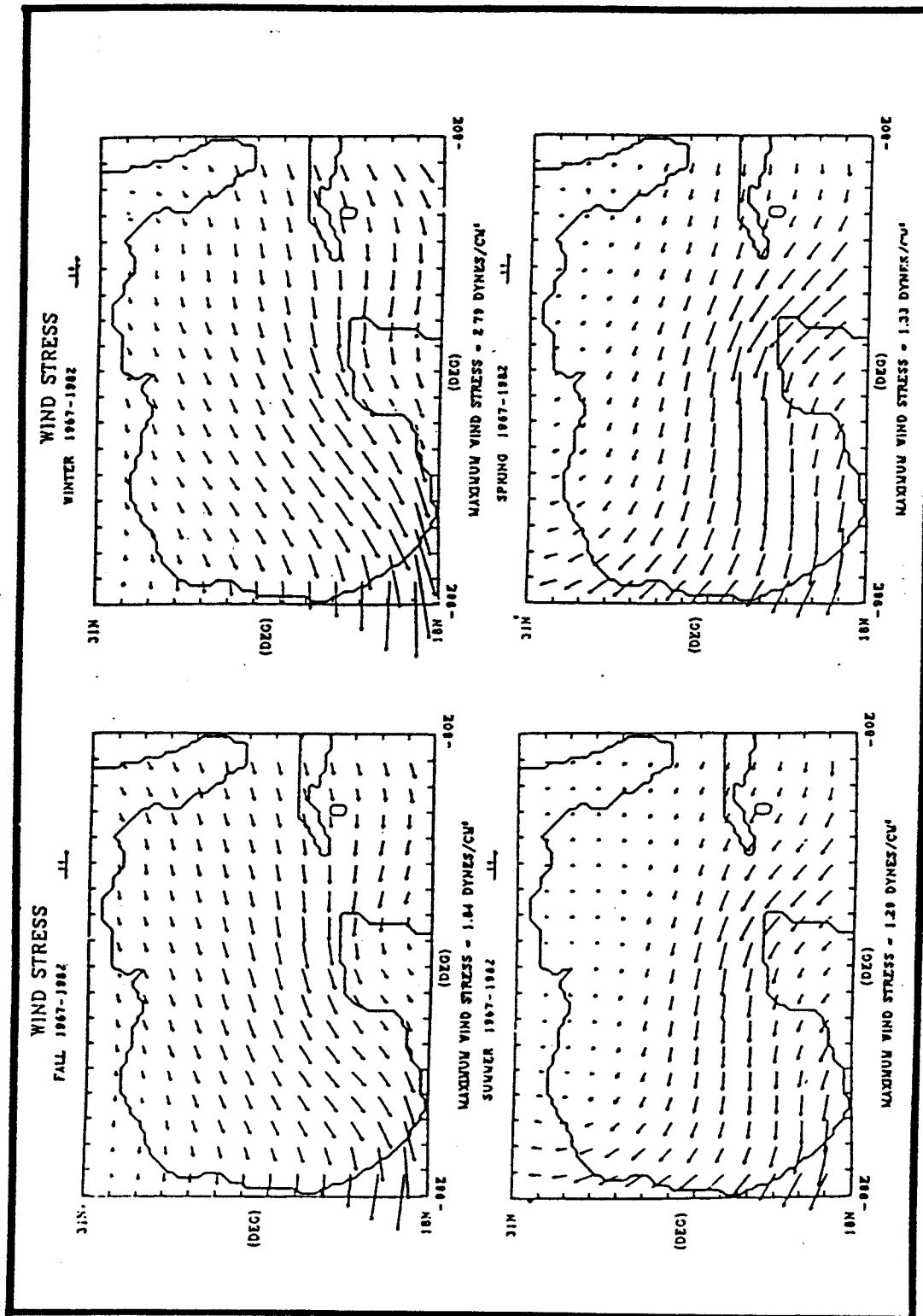


Figure 4.4. Spatial distribution of seasonal mean wind stress over the Gulf of Mexico based on the U.S. Navy's 12-hourly surface pressure analysis for the period 1967 to 1982 (From: Rhodes et al., 1985).

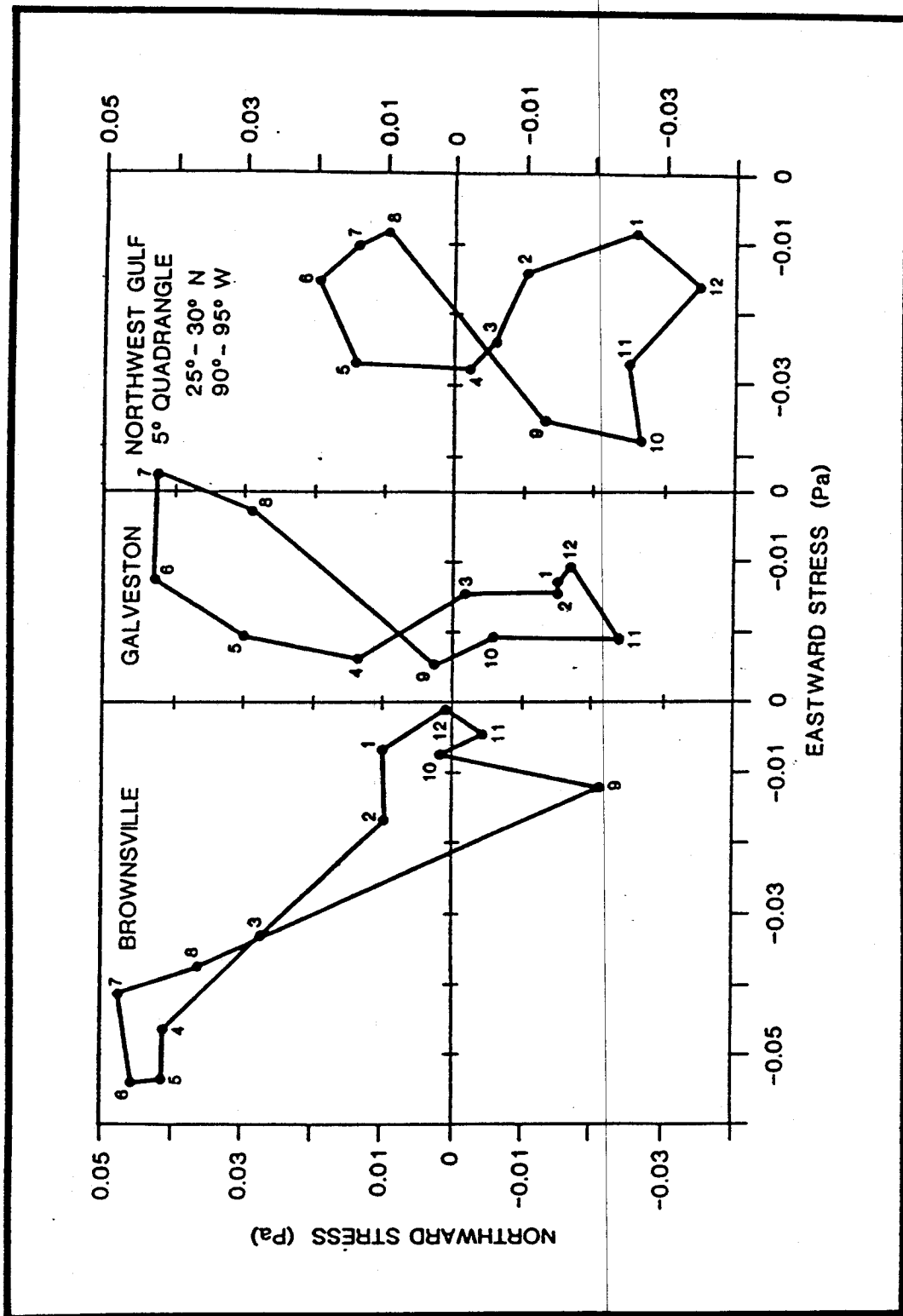


Figure 4.5. Monthly mean wind stress for Brownsville, TX; Galveston, TX; and the 5° quadrangle 25°-30°N, 90°-95°W (From: Cochran and Kelly, 1986a).

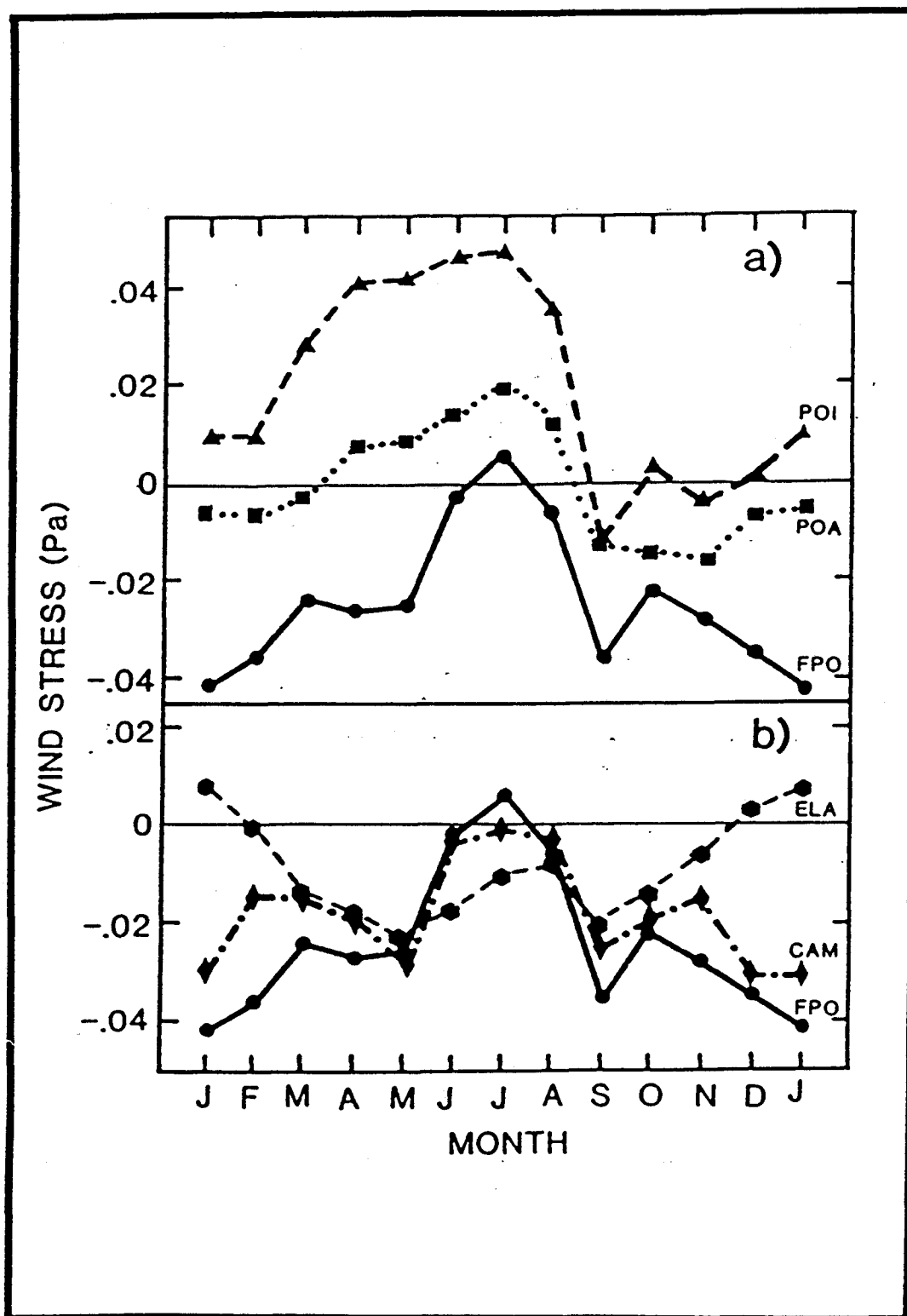


Figure 4.6. Monthly mean alongshore components of wind stress for the waters off a) Port Isabel (POI), Port Aransas (POA), and Freeport (FPO), TX; and off b) Freeport, TX (FPO), Cameron, LA (CAM), and the Louisiana coast between 92°W and the Mississippi River Delta (ELA) (From: Cochrane and Kelly, 1986a).

with the warmer, moister mT air mass. DiMego et al. (1976) analyzed all frontal passages into the Gulf for the period 1965 to 1972 and computed statistics of frequency and duration of frontal systems. Table 4.1 shows the results for the middle of the Texas-Louisiana shelf as interpolated from their maps. The transition from the low-frequency regime of summer to the high-frequency regime of winter is sharp in the fall, occurring between September and October. A more gradual decrease in activity occurs in spring. Henry (1979) also analyzed frontal activity in the Gulf and found a similar pattern, but less frequent frontal activity in winter.

4.3.3 Tropical Storms and Hurricanes

Tropical cyclones produce some of the most extreme meteorological conditions that occur over the Texas-Louisiana shelf. Meteorologists technically define a tropical cyclone (Neumann et al., 1981) as a non-frontal, low-pressure, large-scale system that develops over tropical or subtropical waters and has definite organized circulation. They further classify these cyclones according to wind speed: the sustained surface winds of a tropical depression are ≤ 61 km/h; a tropical storm, 62 to 117 km/h; and a hurricane, ≥ 118 km/h. Tropical cyclones draw their energy from the latent heat of condensation of water vapor over the warm waters of the Atlantic Ocean, the Caribbean Sea and the Gulf of Mexico. Tropical cyclones usually have a central "eye," with a geometric radius of 10 to 30 km or more. Winds increase in strength and become tangent to the eye wall (boundary updraft column); inside the eye, winds are light and cloud cover and air pressure are minimum. The high wind stresses, translational motion and low-pressure anomalies of tropical cyclones cause strong currents, high waves, large increases in sea level, intense mixing, and movement of larger sediment fractions. These effects are described in more detail in the appropriate sections.

In the Gulf of Mexico, tropical cyclones have a "season" in that they occur only between the months of May and October. Figure 4.7 shows the tracks of all tropical cyclones that occurred in the Gulf of Mexico in each of these months in the period from 1886 through 1980. They are most frequent in September. In June and July the cyclones tend to develop in western Atlantic, Caribbean and the Gulf of Mexico, but later in the season they tend to form in the eastern Atlantic near the Cape Verde Islands. Ebersole (1985) compiled, for the entire Gulf of Mexico and for the Atchafalaya Bay vicinity, which is centrally located on the shelf area of interest here, the historical frequency of occurrence for the following hurricane properties: heading, forward speed, central pressure and maximum wind (Figure 4.8). The number of occurrences for Atchafalaya Bay (within 100 miles to either side) are too few to define a "typical" hurricane.

As a rough estimate of the frequency of occurrence of landfall by a tropical cyclone of any category along the Texas-Louisiana coast, Figure 4.9 shows a portion of the results from the study of Ho et al. (1975) as presented by Neumann et al. (1981). For the approximately 400 NM of the portion of interest here, the figure suggests about 70 landfalls per 100 years (based on average of about 1.75 landfalls per 10 NM), i.e., not quite one per year. Because of their low frequency of

Table 4.1. Frequency and duration of frontal passages on the Texas-Louisiana shelf, 1965-1972 (From: DiMego et al., 1976).

Month	Passages/Month	Frontal Duration (h)
January	9	24
February	9½	21
March	8	24
April	6½	27
May	4½	30
June	2	24
July	2	24
August	2	42
September	3	48
October	6	30
November	7	24
December	9	30

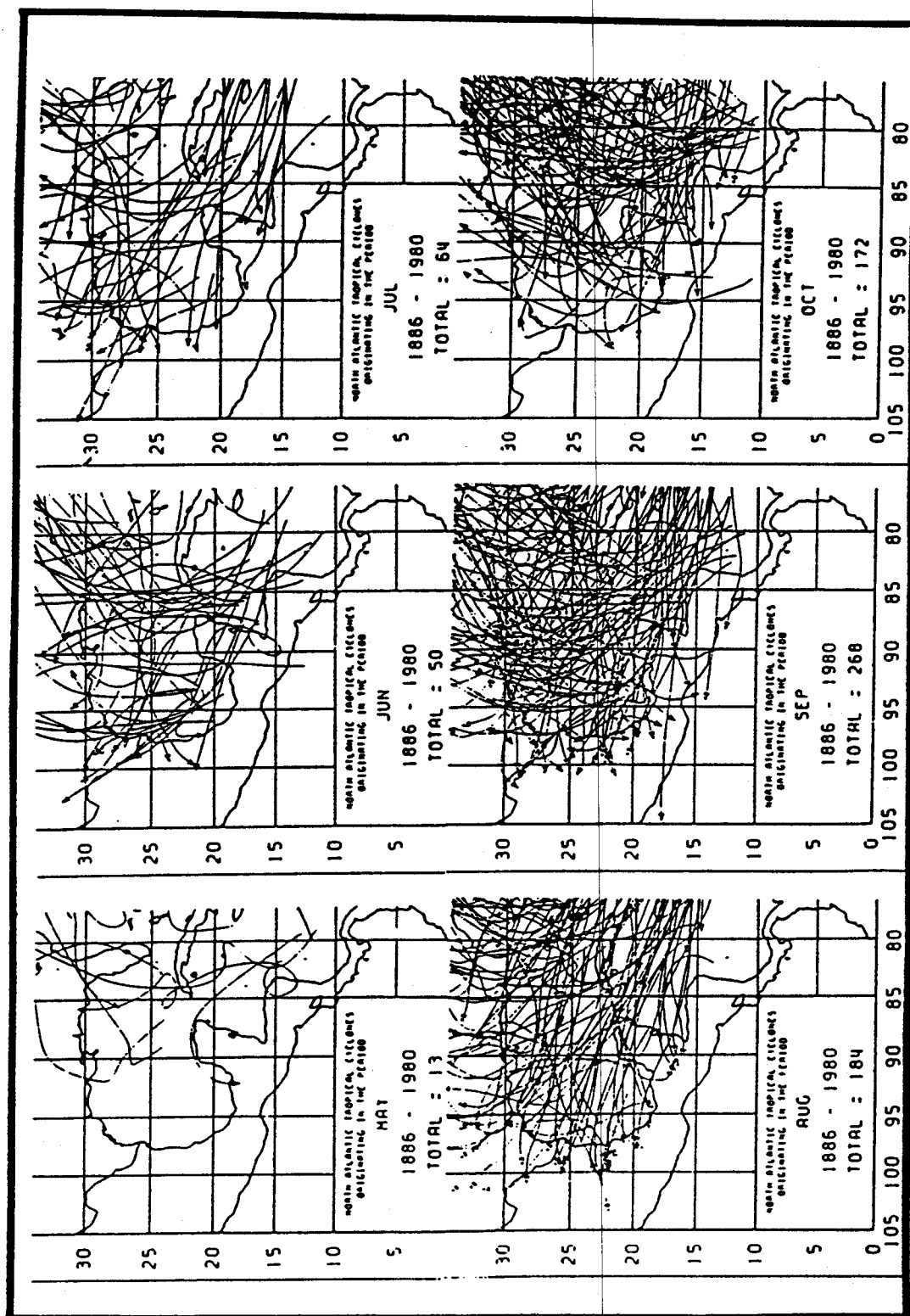


Figure 4.7. Tracks in the Gulf of Mexico of all North Atlantic tropical cyclones originating in each of the months May through September during the period 1886 to 1980.

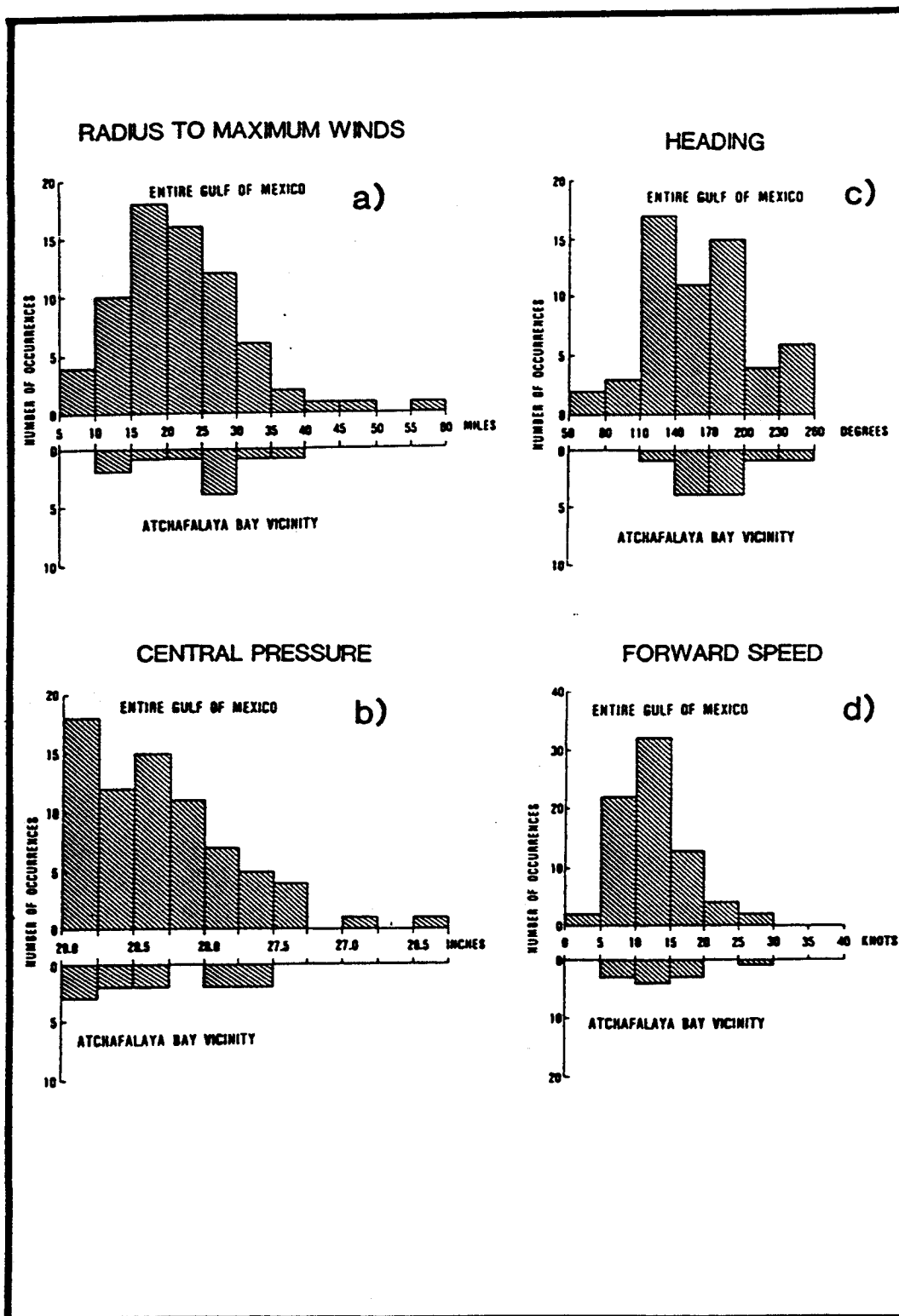


Figure 4.8. Historical (80-year population) frequency of occurrence for a) hurricane radius to maximum winds, b) hurricane central pressure, c) hurricane heading, and d) hurricane forward speed for the entire Gulf of Mexico and the Atchafalaya Bay vicinity (From: Ebersole, 1985).

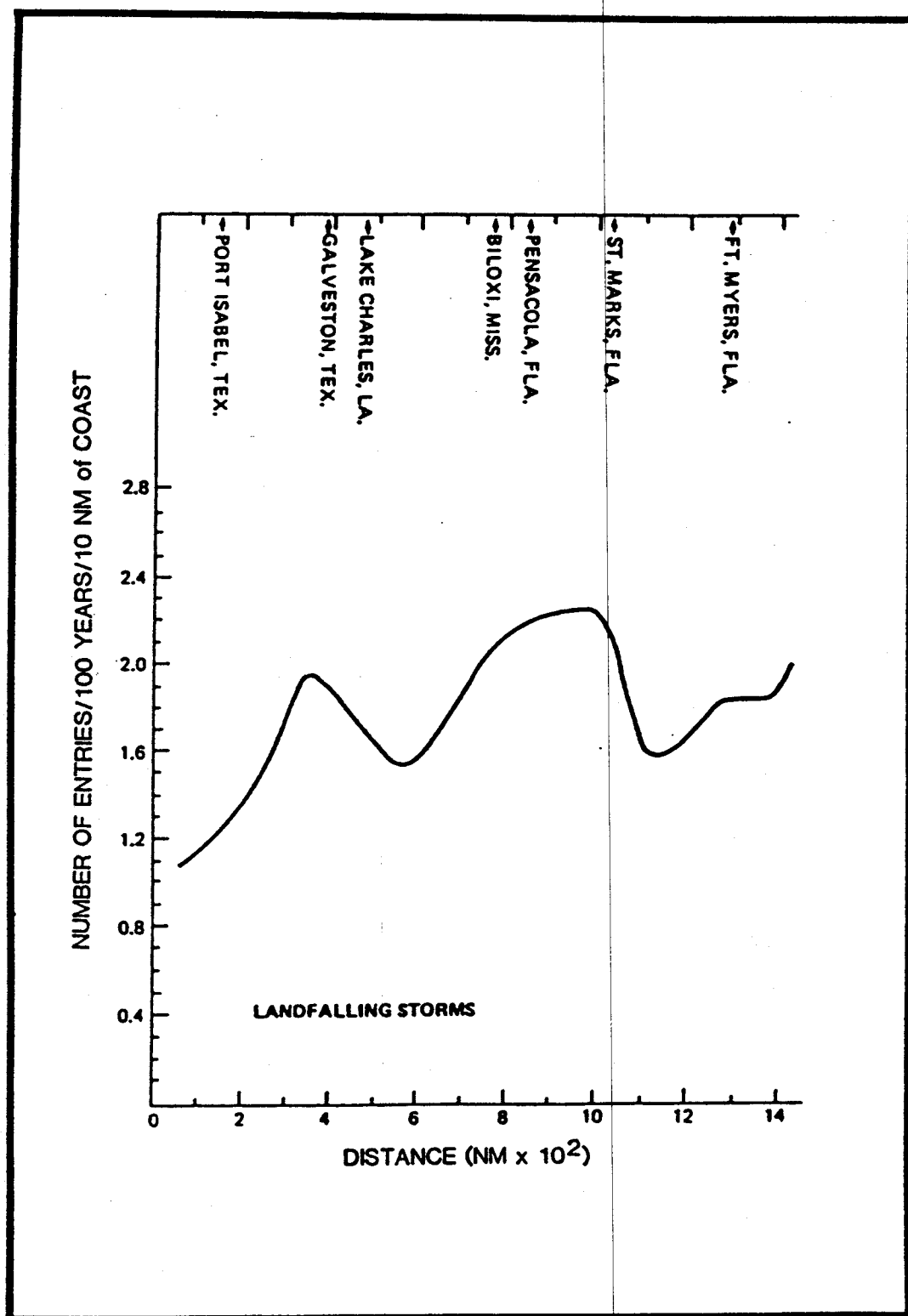


Figure 4.9. Smoothed frequency of landfalling tropical storms and hurricanes (1871 to 1973) for the Gulf of Mexico (Adapted from: Neumann et al., 1981).

occurrence, the oceanic effects of tropical cyclones, although extreme at times, probably are not a significant part of the shelf's current and hydrographic climatology, except perhaps in the area of sediment distribution.

Although the conditions that accompany some of the winter frontal passages are severe enough at times to be compared with tropical storms, northers do constitute part of the regional climatology because of their frequency. The cyclonic storms that sometimes form along the fronts of cP air masses over Gulf waters in winter are termed extratropical. These disturbances tend to be less severe and much larger, spatially, than tropical cyclones and are a distinctly different class of storms. Frey et al. (1981) describe the formation and effects of such winter storms in 1979 at the West Hackberry brine disposal site of the DOE/SPR.

4.3.4 Heat and Mass Exchange with the Shelf Waters

The exchange of heat and water across the sea surface may significantly modify water masses on the shelf, and thus, represents an important forcing factor. Etter et al. calculated a monthly, multi-annual mean heat budget for the Texas-Louisiana shelf, which in symbols may be expressed as

$$Q_R = Q_A + Q_T + Q_V$$

where Q_R is the radiation balance at the sea surface (incoming solar shortwave radiation minus outgoing longwave radiation), Q_A is the net turbulent heat flux from the sea surface (latent plus sensible heat flux), Q_T is the rate of oceanic heat storage, and Q_V is the heat flux divergence due to oceanic motions. They calculated monthly heat storage rates (Q_T) directly using Ulm's (1983) results of a volumetric temperature-salinity census of the shelf waters, and for surface heat exchange ($Q_R - Q_A$) they used unpublished monthly data of Bunker. The heat budget equation then yielded heat flux divergence values (Q_V) as a residual. Figure 4.10a shows their results. There is a net heat loss across the sea surface from September through February and a net heat gain from March through August. Net heat storage in the shelf waters actually begins in February and continues into September because of the contribution from heat flux convergence (i.e., negative divergence). The heat flux divergence term is negative throughout the year except in June and July, and Etter et al. (1985) argue that this pattern is consistent with the prevailing annual march of the alongshore component of wind stress (Figure 4.10b) and the upwelling and downwelling such a pattern would produce in coastal waters.

The extraction of heat from the water column by winter fronts proceeds in a stepwise manner (Nowlin and Parker, 1974). Each cold-air outbreak cools and deepens the surface mixed layer. This modification is more pronounced in the nearshore waters because the air mass rapidly gains heat and moisture from the ocean as it moves south. Figure 4.11 shows an example of the annual cycle of the cooling and warming of the water column at the West Hackberry DOE/SPR site from March 1981 through April 1982. The water depth is 10 m; instrument depths are 3.7 m (DT) and 7.3 m (DB). The periods of pronounced cooling, e.g., 18 to

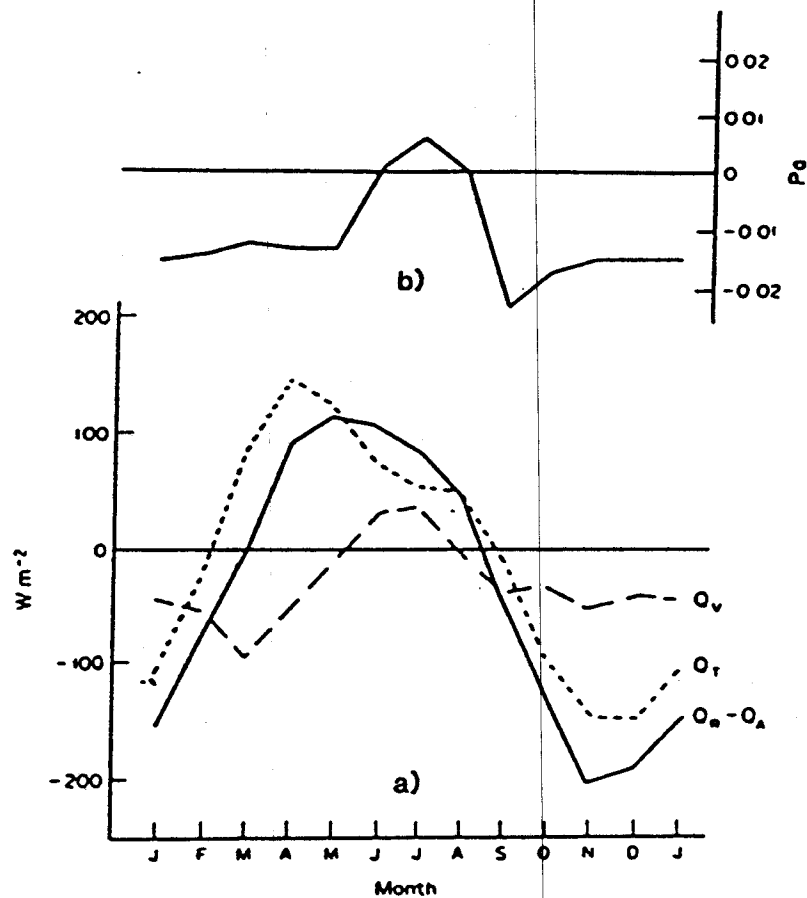


Figure 4.10. a) Annual marches (monthly means) of the rate of oceanic heat storage (Q_T), surface heat exchange (Q_R and Q_A), and flux divergence (Q_v) for Texas-Louisiana shelf waters. b) Annual march (monthly means) of the alongshore component of wind stress for the entire Texas-Louisiana coast from the Rio Grande to the Mississippi River Delta (From: Etter et al., 1985).

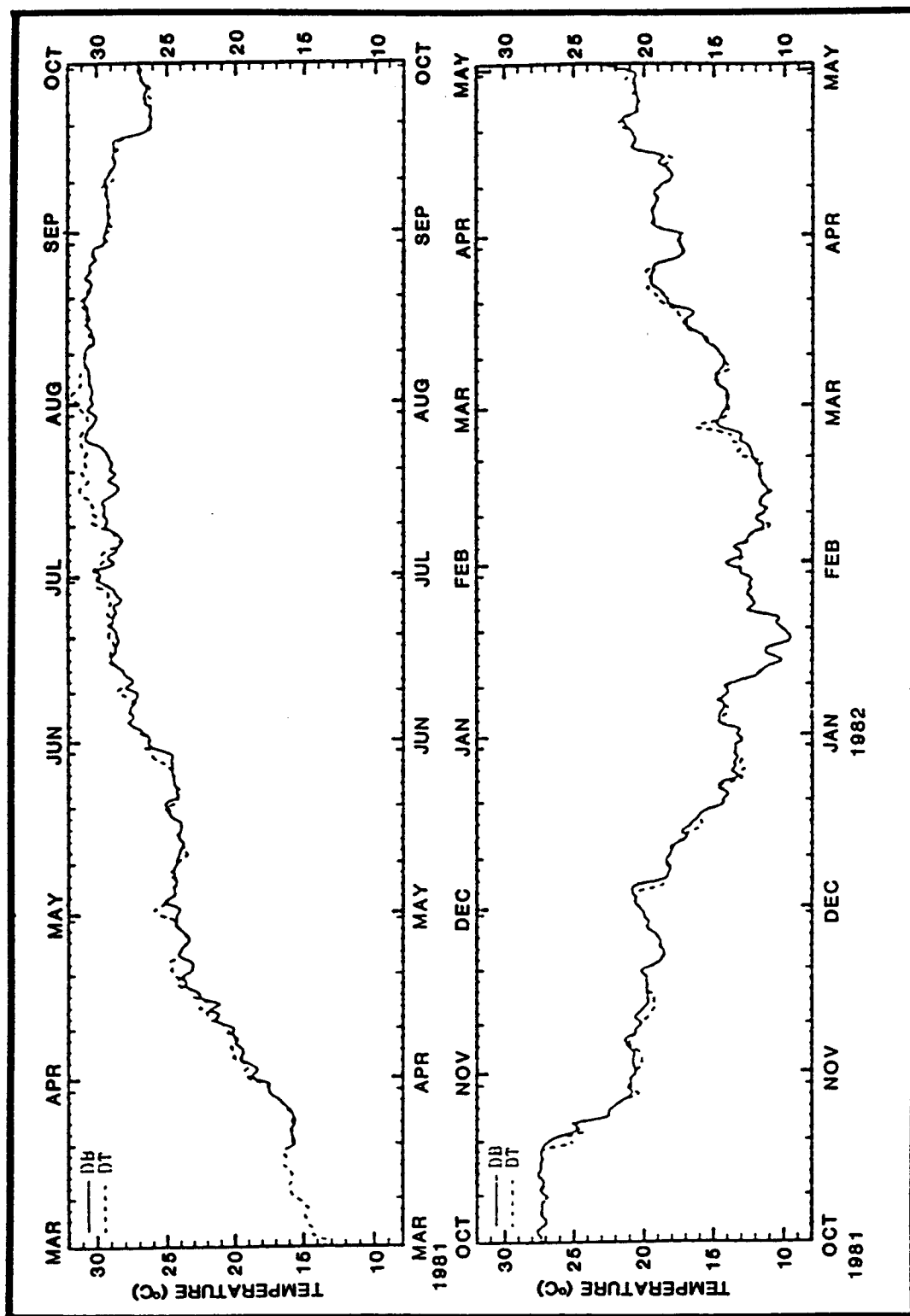


Figure 4.11. Forty-hour, low-pass-filtered time series of temperature measured at the West Hackberry brine disposal site of the Strategic Petroleum Reserve in 10 m of water, at 3.7 m below the surface (DT) and 7.3 m below the surface (DB) during the period March 1981 through April 1982 (From: Kelly et al., 1983b).

20 September, 19 to 28 October, 5 to 7 December 1981, and 10 to 14 January 1982 coincided with the passage of cold fronts through the area (Kelly et al., 1983b). During the cooling event of 19 to 28 October 1981, the temperature decreased by about 6.5°C , or $-0.72^{\circ}\text{C}/\text{day}$. Vertical temperature differences are small, less than 1°C most of the time, except in July and early August 1981 when a difference of up to 3°C was recorded. During the warming period, the near surface temperature is slightly warmer than the near-bottom temperature, and vice versa for the cooling period.

Nowlin and Parker (1974) studied the details of the heat extraction and water mass modification processes during one major outbreak of cold dry air that occurred in January 1966. During the 15-day period between their surveys of the middle portion of the Texas-Louisiana shelf, the rate of cooling of the water column was largest nearshore, about $-0.25^{\circ}\text{C}/\text{day}$, because the continental polar air mass had received a minimum of modification by the sea. The local rate of heat extraction, formed from the product of the local rate of change of vertically-averaged temperature, the local depth, mean density and specific heat, averaged about $400 \text{ cal}/\text{cm}^2/\text{day}$ within 50 miles from shore and generally increased to 700 to $1,500 \text{ cal}/\text{cm}^2/\text{day}$ at the offshore survey limits. They neglected advection. For comparison, Etter's mean monthly heat loss ($Q_R - Q_A$) for January (Figure 4.10A) is about $-150 \text{ W}/\text{m}^2$ or $-310 \text{ cal}/\text{cm}^2/\text{day}$ [negative values correspond to positive heat extraction values of Nowlin and Parker (1974)].

Evaporation and precipitation also modify the shelf waters. Nowlin and Parker (1974) estimated that salinity increased about 1‰ during the passage of the cold dry polar air. They concluded that the change in temperature-salinity (T-S) relationships before and after the outbreak indicated the possible formation over the shelf of subsurface water types generally found beneath the Subtropical Underwater core in the open Gulf. As noted in Section 4.2.4, observations of evaporation and precipitation over the shelf are generally lacking. Dinnel and Wiseman (1986) review the existing estimates of these parameters over the shelf and calculate their own. Tables 4.2, 4.3, and 4.4, reproduced from their study, summarize the various estimates and compare the characteristics of the Texas-Louisiana shelf with those of the Mid-Atlantic Bight and the South Atlantic Bight. Note that, where data are available, river input, precipitation and evaporation are all larger on the Texas-Louisiana shelf.

4.4 RIVERINE INFLUENCES

The Mississippi River is the principal source of fresh water on the Texas-Louisiana continental shelf. Its mean annual contribution is more than 15 times that of all of the other river sources combined (Cochrane and Kelly, 1986b) and almost four times larger than the contribution from precipitation (Dinnel and Wiseman, 1986). The variability of the flow of the Mississippi River contributes significantly to the variability of the currents, hydrography, nutrients, suspended sediments, chemical properties, and biological processes on the shelf, particularly on the inner third.

Table 4.2. Comparison of the evaporation rate calculated by Dinnel and Wiseman (1986) with global climatic evaporation rates prorated to this study area (Jacobs, 1951; Baumgartner and Reichel, 1975; Bunker, 1976), and with a regional evaporative rate estimated for the study area (Etter, 1975) (From: Dinnel and Wiseman, 1986).

Evaporation Rate (km^3/yr)	Study
148	Dinnel and Wiseman (1986)
128	Jacobs (1951)
160	Baumgartner and Reichel (1975)
214	Etter (1975)
187	Bunker (1976)

Table 4.3. Shelf precipitation rates calculated from onshore coastal data, 30-yr mean values, and climatic annual precipitation estimates for the Texas-Louisiana shelf (From: Dinnel and Wiseman, 1986).

Year	Onshore	Shelf
1963	105	79
1964	125	94
1965	125	94
30-yr Mean	143	108 [†]
Climatic Estimates		85 [†]
		107 [§]
		53 [#]

* 1963-1965 were drought years. Values are in km^3/yr .

[†] Jacobs (1951).

[§] Baumgartner and Reichel (1975).

[#] Dorman and Bourke (1981).

Table 4.4. Comparison of Mid-Atlantic Bight and South Atlantic Bight characteristics with those of the Texas-Louisiana shelf (From: Dinnel and Wiseman, 1986).

	Mid-Atlantic Bight	South Atlantic Bight	Texas- Louisiana Shelf
Area (km^2)	70,656	106,000	106,866
Volume (km^3)	3,108	3,610	3,746
River Input (km^3/yr)			
Long-Term	150	50	382 [†]
Short-Term			320 [§]
Precipitation (km^3/yr)			
Long-Term	49		108 [#]
Short-Term			89 ^{**}
Evaporation (km^3/yr)	71		148 ^{**}

* Mid-Atlantic data from Ketchum and Keen (1955), South Atlantic Bight data from Atkinson et al. (1983); values are prorated to a 50-fathom maximum shelf depth.

[†] Extrapolated value for major rivers.

[§] Mean 1962-1965.

[#] Estimate of shelf 30-yr mean.

^{**} Offshore mean 1963-1965.

The waters of the Mississippi River reach the Texas-Louisiana shelf through its two natural distributaries, the main river and the Atchafalaya River. The percentage carried by the Atchafalaya has increased significantly during this century (Gunter, 1979): between 1900 and 1919 the Mississippi proper carried about 85% of the river flow while the Atchafalaya carried about 15%; from 1920 through 1950 the Atchafalaya's share increased to about 30%, with a commensurate decrease in the Mississippi's flow. The U.S. Army Corps of Engineers (COE) now controls the partition by means of its Old River Control Structure. The COE maintains a split of approximately 30/70, but permits the percentage carried by the Atchafalaya to rise sometimes during floods; during the flood year of 1973 the Atchafalaya took over 37% of the flow (Gunter, 1979).

All of the flow from the Atchafalaya contributes to the fresh water budget of the shelf, but only an estimated 53% (Dinnel and Wiseman, 1986) to 65% (Scruton, 1956) of the discharge from the Mississippi Delta consistently flows onto the Texas-Louisiana shelf. Thus, the net contribution from the Mississippi distributary ($0.70 \times 0.53 = 0.37$) is now about the same as that of the Atchafalaya distributary. It is important to note that the total freshwater input to the shelf increased by more than 10% between 1920 and 1950, solely because of the change in the partition of the flow between the two distributaries.

Gunter (1979) analyzed the flows of the Mississippi-Atchafalaya River System for the 79-year period from 1900 through 1978 and found a mean annual flow of 646,304 cubic ft/s (cfs) ($1 \text{ cfs} = 2.832 \times 10^{-2} \text{ m}^3/\text{s}$; $1 \text{ cfs} = 8.932 \times 10^{-4} \text{ km}^3/\text{year}$). Table 4.5 shows that most flows were in the 500,000 and 600,000 cfs classes, with 53 of the 79 years, or 67%, below 699,000 cfs. A flood year has, by definition, an annual flow of 700,000 cfs or greater (Gunter, 1979). Flood years are not especially associated, and in several cases low flows and flood years are close together. Corresponding volumes of freshwater flow onto the shelf can be only estimated because of the variable partition between the two distributaries and the imprecise estimates of the percentage of water from the delta that flows onto the shelf; a rough conversion factor is 0.7.

Intra-annual variation, in terms of monthly mean discharges for the Mississippi-Atchafalaya System, follows a cycle that has a strong maximum (spring flood) in April of more than twice the annual mean discharge and a rather flat minimum in October (Figure 4.12). There is considerable inter-annual and short-term variability about this mean progression, as can be seen in Figures 4.13 and 4.14, which show the average daily stream flow for the water years 1977 to 1978 and 1979 to 1980, respectively. (A water year runs from October of the preceding calendar year through September of the given calendar year.) In addition to the April maximum, significant secondary maxima frequently occur in December, January and May. Minimum flow, in these example years, occurs near the month of October, and the lowest value reached varies by more than a factor of two. The annual means for 1977, 1978, and 1980 (541,000 cfs, 671,000 cfs, and 623,000 cfs, respectively) were normal according to Table 4.5. However, the 1979 annual mean of 1,012,000 cfs ranks as the third biggest flood since 1900.

Table 4.5. Annual flows of the Mississippi River for the twentieth century
(From: Gunter, 1979).

Class Ranges (Thousands of cfs*)	Number of Annual Flows in Class Range
300	3
400	10
500	17
600	23
700	15
800	6
900	3
1,000	1
1,100	1

* cfs = cubic ft/s

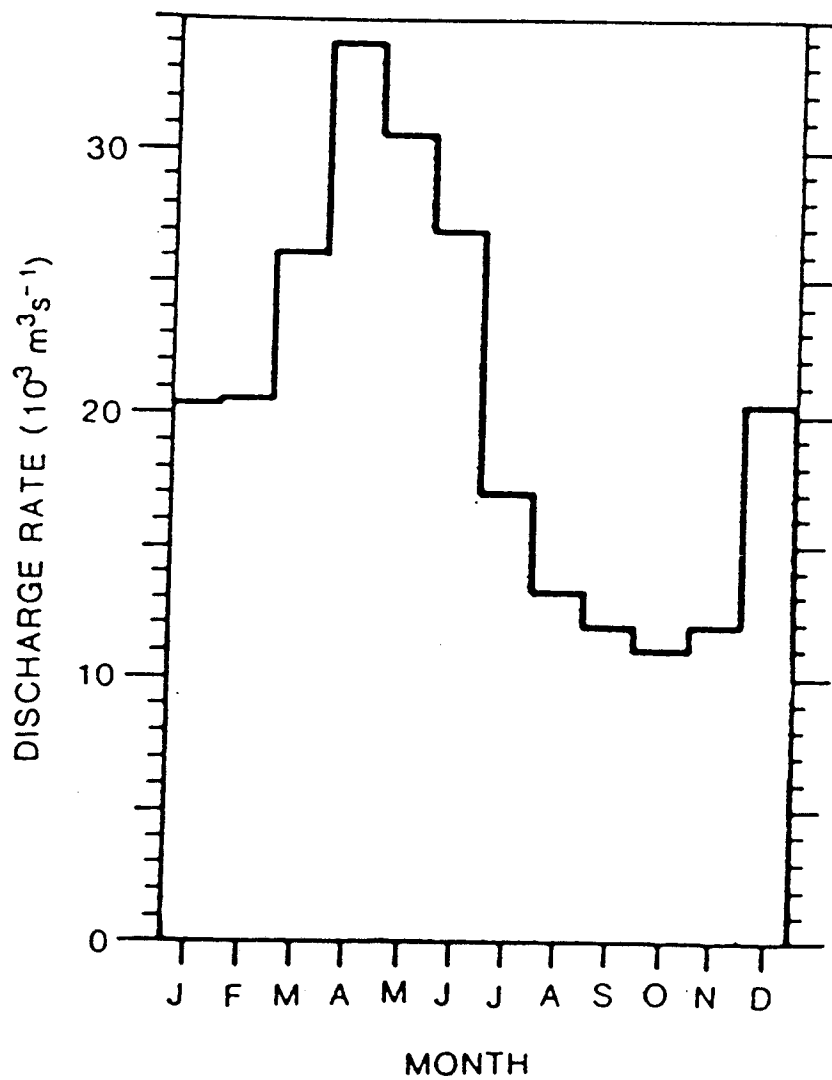


Figure 4.12. Monthly mean discharge rates for the Mississippi-Atchafalaya River system based on the water years 1977 to 1983 (From: Cochrane and Kelly, 1986b).

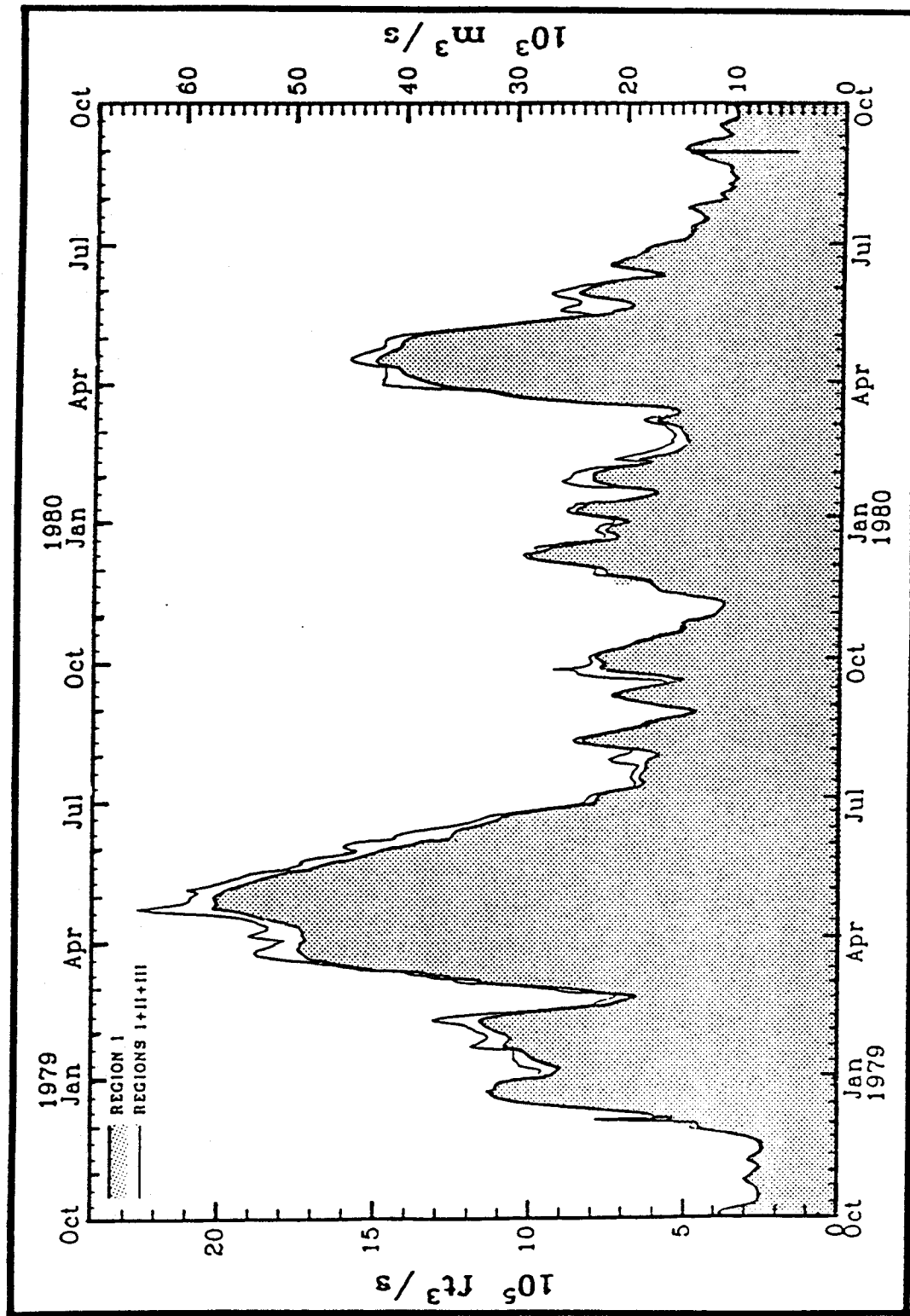


Figure 4.14. Average daily streamflow for water years 1979 and 1980 for the Mississippi-Atchafalaya River System (heavy line) and for this system plus all other rivers from the Vermilion to the San Bernard (light line) (From: Kelly et al., 1981).

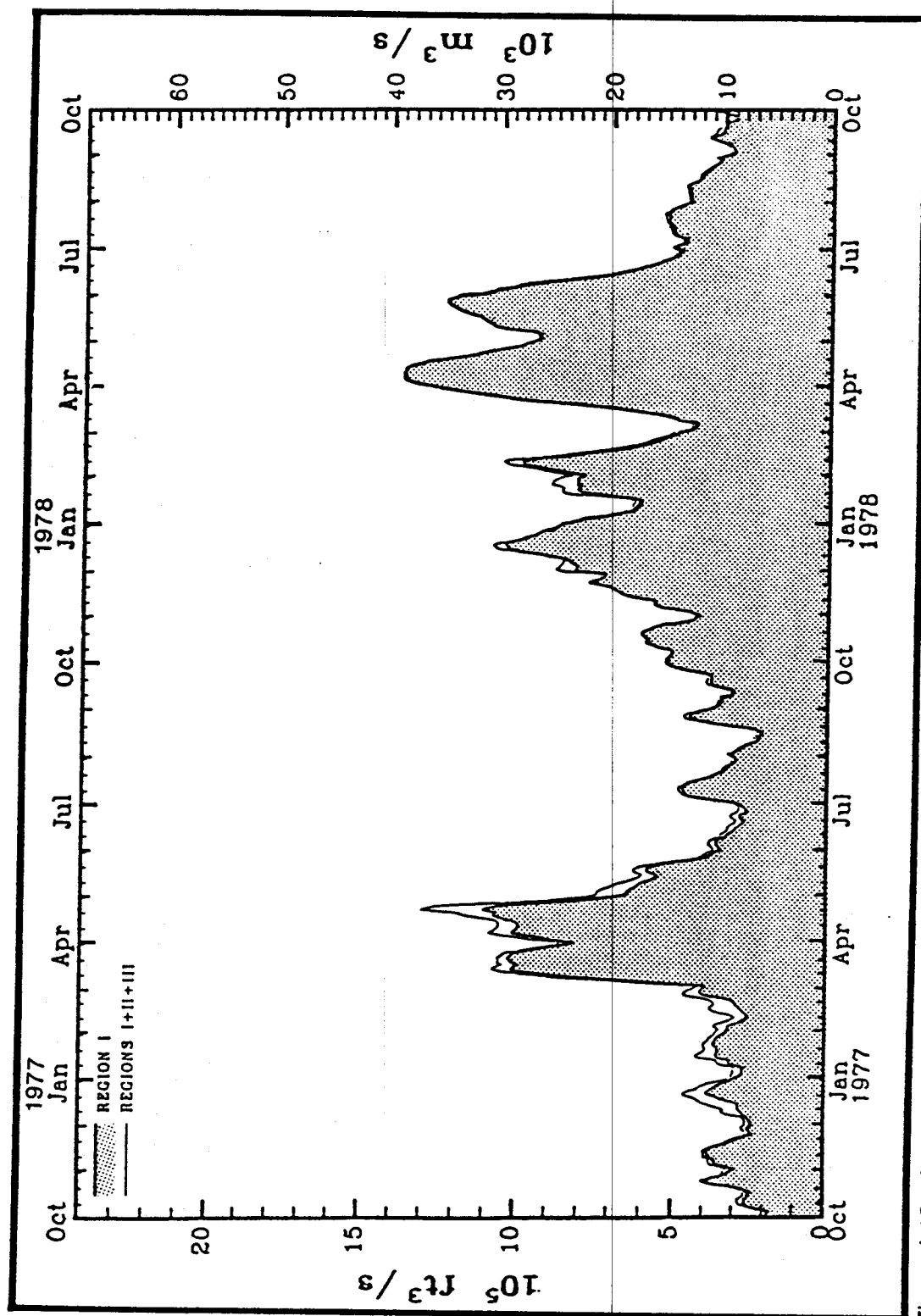


Figure 4.13. Average daily streamflow for water years 1977 and 1978 for the Mississippi-Atchafalaya River System (heavy line) and for this system plus all other rivers from the Vermilion to the San Bernard (light line) (From: Kelly et al., 1981).

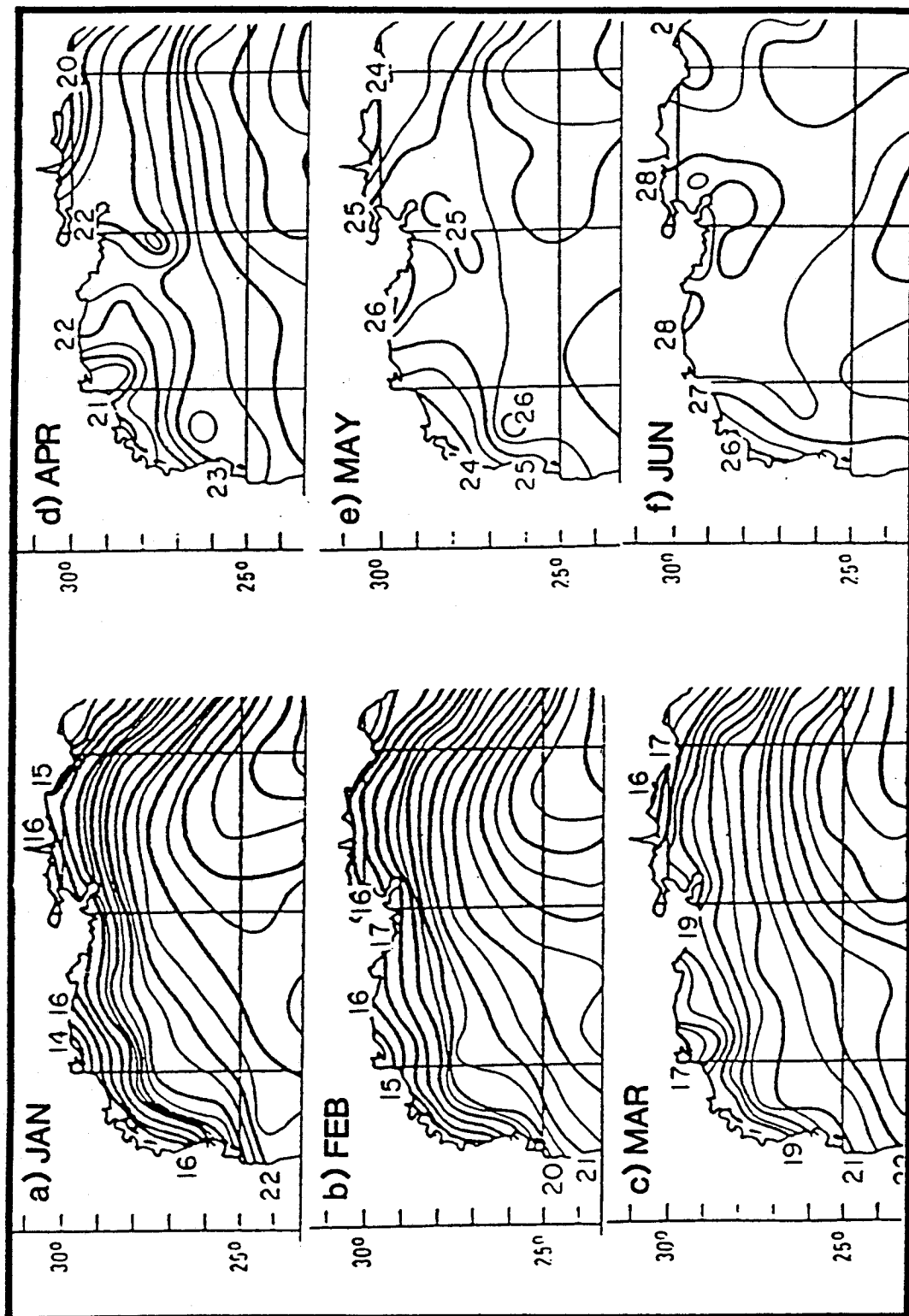


Figure 4.15. Monthly mean sea-surface temperatures for the northwestern Gulf of Mexico, a) January through
 1) December (Adapted from: Robinson, 1973).

The total additional discharge from all rivers and streams from the Vermilion in Louisiana to the San Bernard in Texas is also indicated in Figures 4.13 and 4.14 (light line minus heavy line). The degree to which the Mississippi-Atchafalaya flow dwarfs all other riverine input is evident.

4.5 HYDROGRAPHY

4.5.1 Water Temperature

The spatial and temporal variability of water temperature is caused by advection, turbulent and convective mixing, and the air-sea exchange processes discussed in Section 4.3.4. Temperature is the most abundantly measured of the hydrographic properties, and its variability on the Texas-Louisiana shelf is relatively well documented. For example, Robinson (1973) and Etter and Cochrane (1975) have produced maps of mean temperatures from independent data sets; Ulm (1983) prepared a volumetric T-S census for the waters of the shelf region; the multi-year studies at the various DOE/SPR brine disposal sites along the inner shelf and the MMS environmental studies at the Flower Gardens on the outer shelf have reported the results of continuous sub-surface measurements; and analyses of thermal imagery from satellites have described the synoptic scale spatial variability at the sea surface (e.g., Rezak et al., 1985).

Monthly mean sea-surface temperatures for the northwestern Gulf of Mexico from Robinson's (1973) atlas (Figure 4.15) illustrate the influence of climatology and advection. Shelf waters cool from summer highs near 29 to 30°C to inshore lows of about 14°C in January and offshore lows of about 20°C in February. Warming occurs from March through July. The eastern half of the shelf warms more rapidly and reaches higher temperatures than the western half because of upwelling that begins along the lower Texas coast in May and reaches the west Louisiana coast by July. On the eastern part of the shelf a warm region extends offshore in July and August. (The cold protrusions from Galveston Bay in March and April are probably an artifact of insufficient data, which gives undue weight to flushing events, and probably are not representative of mean conditions.)

Satellite infrared imagery for June through September 1982 (Figure 4.16) illustrates the development of the summer sea-surface temperature pattern over the shelf in greater detail. The June map indicates that upwelling extends as far northeast as Matagorda Bay. During late July and early August, the contrast in temperature between east and west increases markedly, and the warm region in the east grows westward and offshore. However, by mid-September the warm region is pressed against the coast and extends southwestward. The warm region coincides roughly with the summer region of low-salinity surface water (Section 4.5.2). The high stability of the upper layer of the water column in this region facilitates its heating. Figure 4.16 also demonstrates that in the summer months satellite imagery can still detect contrasts in sea-surface temperature on the shelf that are strong enough for useful interpretations to be made.

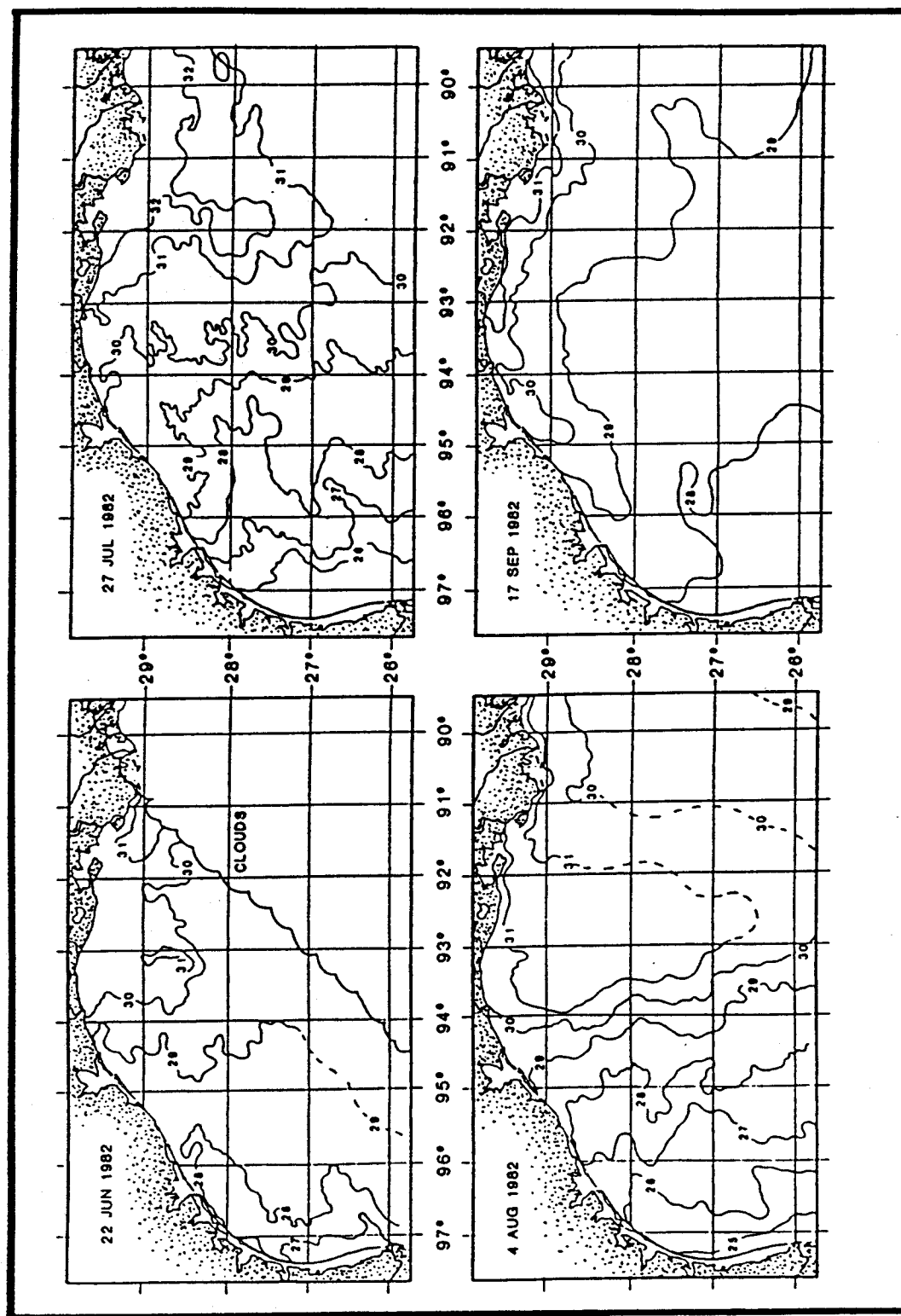


Figure 4.16. Sea-surface temperature based on infrared imagery for days in June, July, August, and September 1982 (From: Cochrane and Kelly, 1986b).

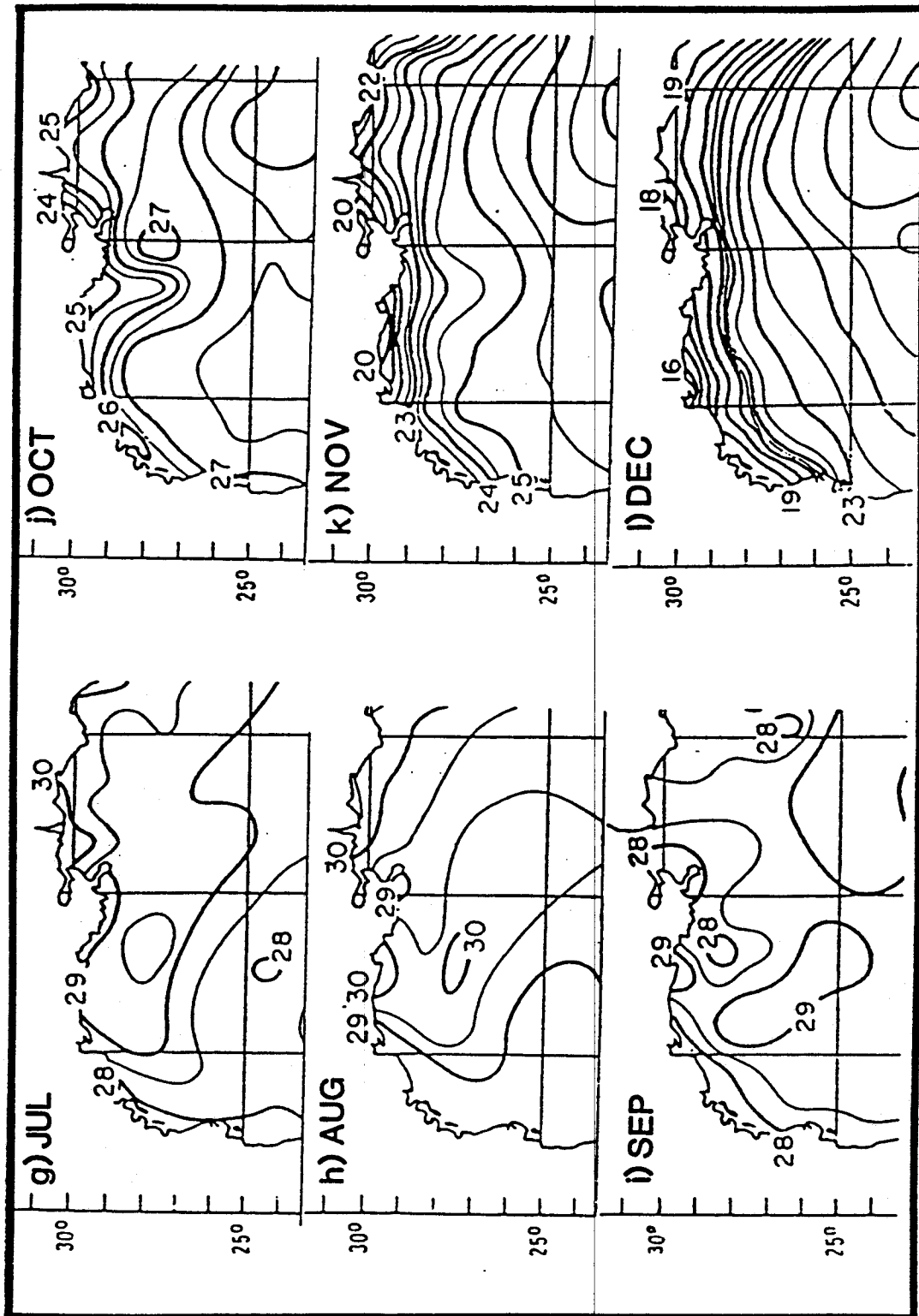


Figure 4.15. (Continued).

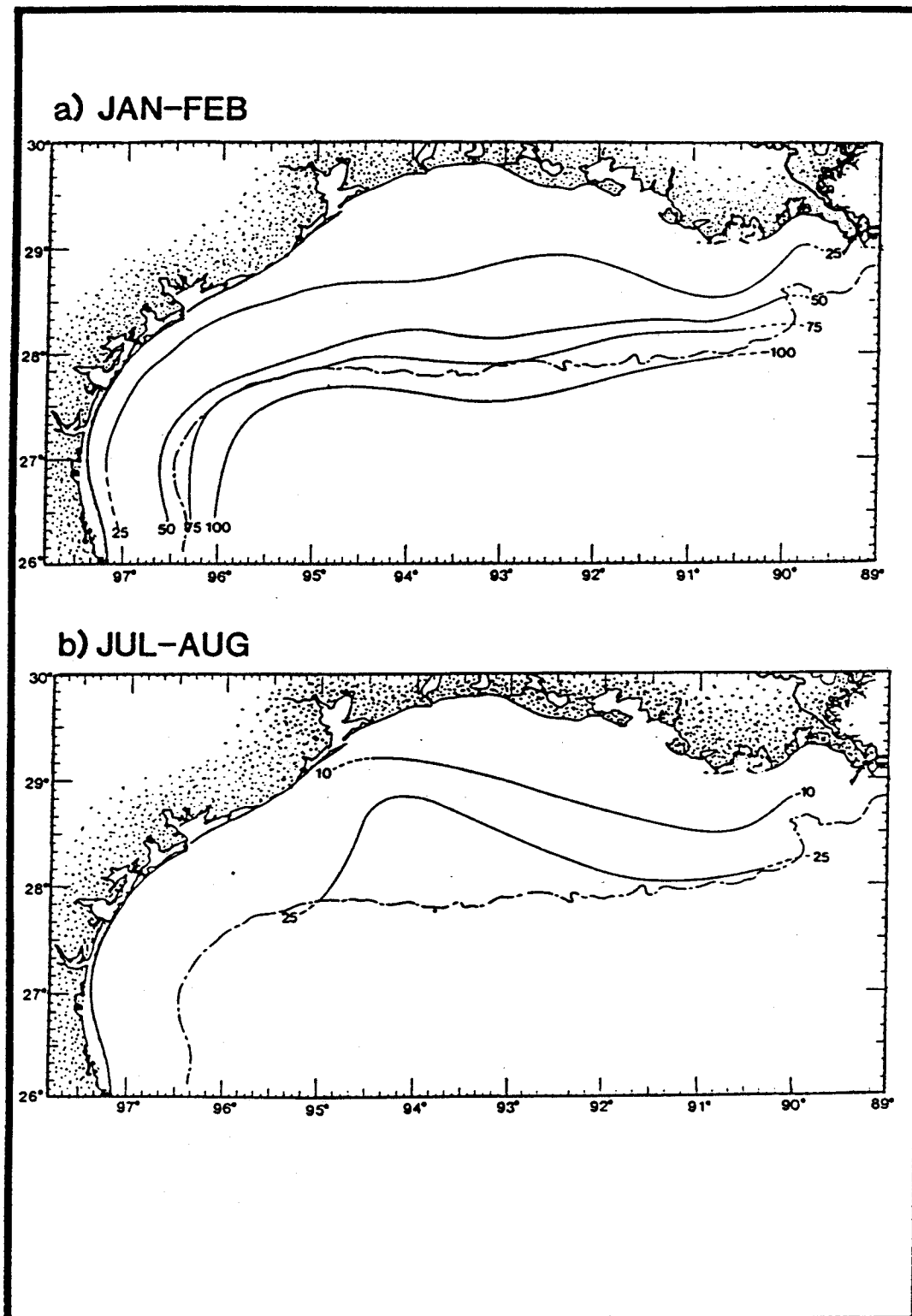


Figure 4.17. Map of mixed layer depths (m) for a) January-February and b) July-August. Dashed lines indicate lack of data, and curvature of these lines suggests most probable tendency (From: Etter and Cochrane, 1975).

According to Etter and Cochrane (1975), the depth of the mixed layer becomes shallow in the summer, ranging from 10 to 25 m, and deepens in the winter, with a maximum depth of about 75 m over the outer shelf between November and February (Figure 4.17). Within about 25 km of the coast, however, temperature is not a good indicator of mixed-layer depth because the distribution of salinity controls the density field much of the year. Thermocline strength over the outer shelf varies seasonally from a maximum of about 6°C in August to a minimum of less than 2°C in March (Etter and Cochrane, 1975). Rezak et al. (1985) report similar results for mixed-layer depth and thermocline strength near the Flower Garden Banks.

Bottom temperatures over the inner half of the shelf follow sea-surface temperatures to the extent that the mixed-layer depth reaches the bottom, while seaward of about 75 m, bottom temperatures reflect the off-shelf annual variations of the Gulf; seaward of about the 120-m isobath, bottom temperatures vary only slightly (Etter and Cochrane, 1975). Figure 4.18a shows that mean summer temperatures near the bottom are highest near shore and lowest offshore and that the isotherms closely parallel the isobaths. During fall and winter, mean bottom temperatures are low both nearshore and offshore with somewhat higher temperatures in between (Figure 4.18b).

4.5.2 Salinity

Wind-driven currents and freshwater influx from the Mississippi-Atchafalaya River System determine much of the spatial distribution and temporal variability of salinity on the shelf. The best coverage of salinity on the shelf is provided by a series of observations taken during 1963, 1964, and 1965 by the U.S. Fish and Wildlife Service aboard the M/V GUS III (Temple et al., 1977). Figure 4.19 shows the sea-surface salinity distributions for alternate months in 1964, and illustrates an annual sequence that is believed to be fairly typical. A band of low-salinity water lies along the coast from September through June, carried by the inshore limb of the gyre of cyclonic circulation that prevails on the shelf during this time. Minimum salinities occur all along the coast in May after the spring flood of the Mississippi in April. The May distribution also shows fresher water extending from the south Texas coast to the northwest along the outer edge of the shelf, a pattern that is attributed to convergence in the wind stress and current fields near the coast and to the offshore limb of the cyclonic gyre. The band recedes upcoast in June and disappears by August, although brackish water remains along the coast from the delta to about 92°W and in a region extending seaward over the shelf. The absence of low-salinity water west of about 92.5°W is caused by upcoast currents and upwelling driven by the upcoast wind stress in July. In September, currents all along the coast rapidly return to prevailing downcoast flow, bringing with them a coastal band of low-salinity water. The main source of the brackish water must be the extensive region of low-salinity water that lies off eastern Louisiana in July and August, because Mississippi discharge is near its annual minimum at this time.

Dinnel and Wiseman (1986) used the GUS III data set in a study of the freshwater budget of the shelf. They also included evaporation and

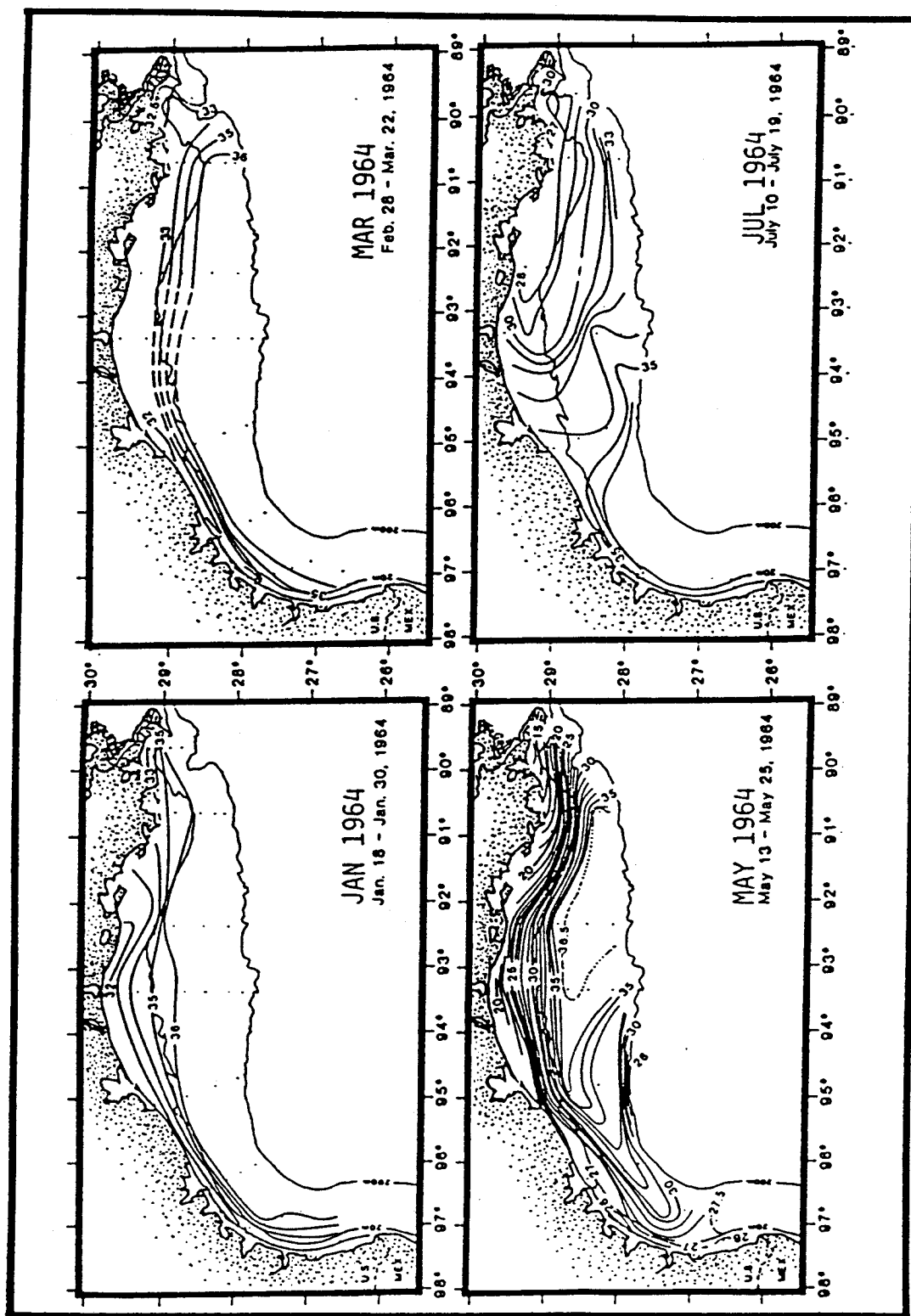


Figure 4.19. Sea-surface salinity (‰) for M/V GUS III cruises in 1964 (From: Cochrane and Kelly, 1982).

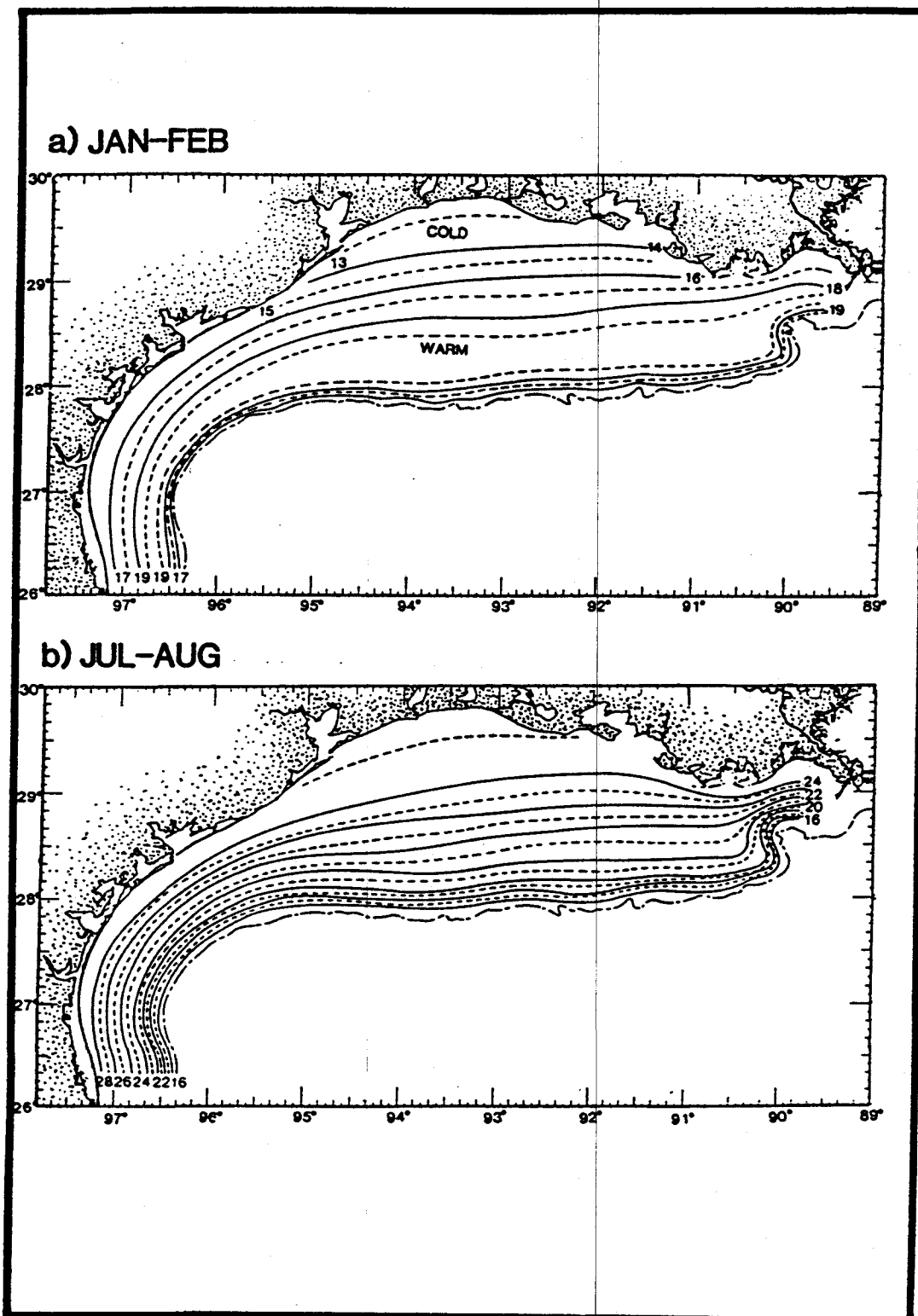


Figure 4.18. Map of mean bottom temperatures ($^{\circ}\text{C}$) for a) January–February and b) July–August. Solid lines represent even values of temperature, and dashed lines indicate odd values (From: Etter and Cochrane, 1975).

precipitation in the budget because evaporation-precipitation is not in balance on the shelf and is comparable to the river discharge term at times. The patterns of distribution of fresh water that they found are consistent with those of low salinity described above, except that the coastal band of maximum freshwater content is displaced slightly seaward of the coast because higher freshwater content is associated with a deeper water column. They found that there was generally a net dispersion of freshwater out of their study area, but that net gains in some spring flood months suggest some additional uncounted flow from the Mississippi Delta. Calculated fill times exhibit an annual cycle, and freshwater volume appears, in most cases, to have originated near the time of the previous spring flood. (Fill time is the length of time required for river runoff, precipitation, and evaporation to account for the volume of excess freshwater on the shelf.)

The slow downcoast change in coastal sea-surface salinity that is found in all of the months shown in Figure 4.19 except July cannot result simply from the local river discharges, which vary widely. Only downcoast advection of Mississippi-Atchafalaya water can explain the very gradual changing. However, floods in the other rivers and flushing of bays and estuaries by strong persistent northers can discharge pulses of relatively fresh water that briefly augment the inshore edge of the coastal brackish band in the near surface layers.

The upwelling current regime acting on the plentiful supply of brackish water in the coastal band produces a dramatically different vertical distribution of the water masses than occurs during downwelling conditions. A typical example of the hydrography of the inner shelf during the downwelling regime is shown by a transect (Figure 4.20) extending 30 km off Freeport, TX. The density field is determined primarily by the salinity field. The strongest part of the frontal zone lies between 10 and 20 km offshore, and the density interface has somewhat of an "S" shape. Contrast this with the conditions during 22 to 23 July 1983. Figures 4.21, 4.22, and 4.23 show the observations along transects running offshore through the Bryan Mound, Big Hill and West Hackberry DOE/SPR brine disposal sites, respectively. Salinity again determines the density field. High salinity water extends shoreward along the bottom almost to the coast, while in the upper layers low salinity water extends seaward past the limits of the transects. The result is extremely strong vertical stratification.

Such stratification also exists over the inner shelf just west of the Mississippi Delta, where Wiseman et al. (1976) observed intense temperature and salinity steps in vertical profiles collected in the river's plume (Figure 4.24).

To provide a more quantitative picture of the annual progressions of temperature and salinity in the coastal regions where the variability is greatest, monthly means for the surface and bottom are shown, together with a representation of both progressions in a T-S diagram, in Figures 4.25 and 4.26 for the Bryan Mound and West Hackberry sites, respectively.

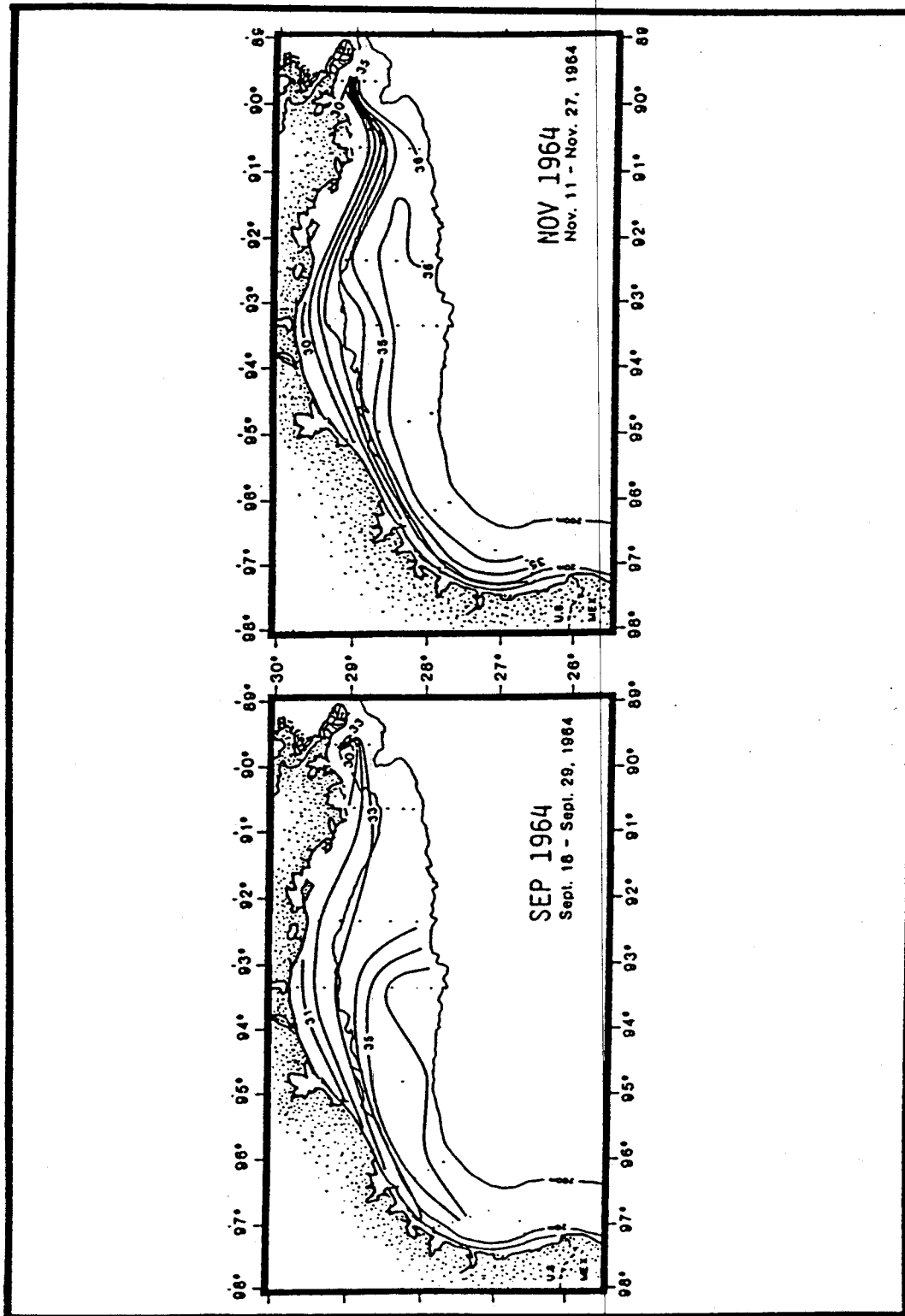


Figure 4.19. (Continued).

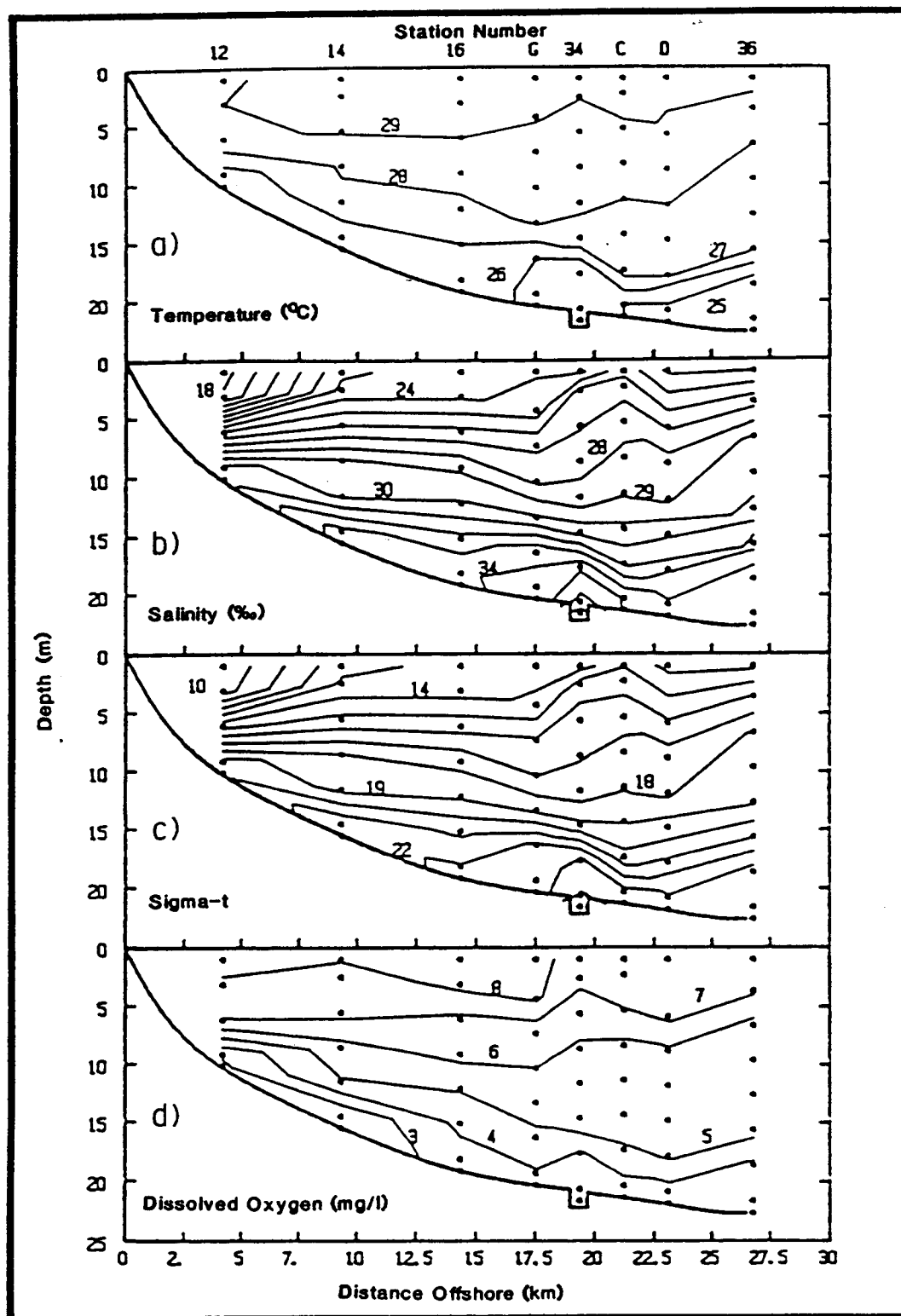


Figure 4.21. Hydrography for cross-shelf transect through the Bryan Mound site on 22 July 1983 (Note: the slightly higher salinity values near the bottom in the vicinity of station 34 were caused by the DOE/SPR brine disposal operations) (From: Kelly et al., 1984b).

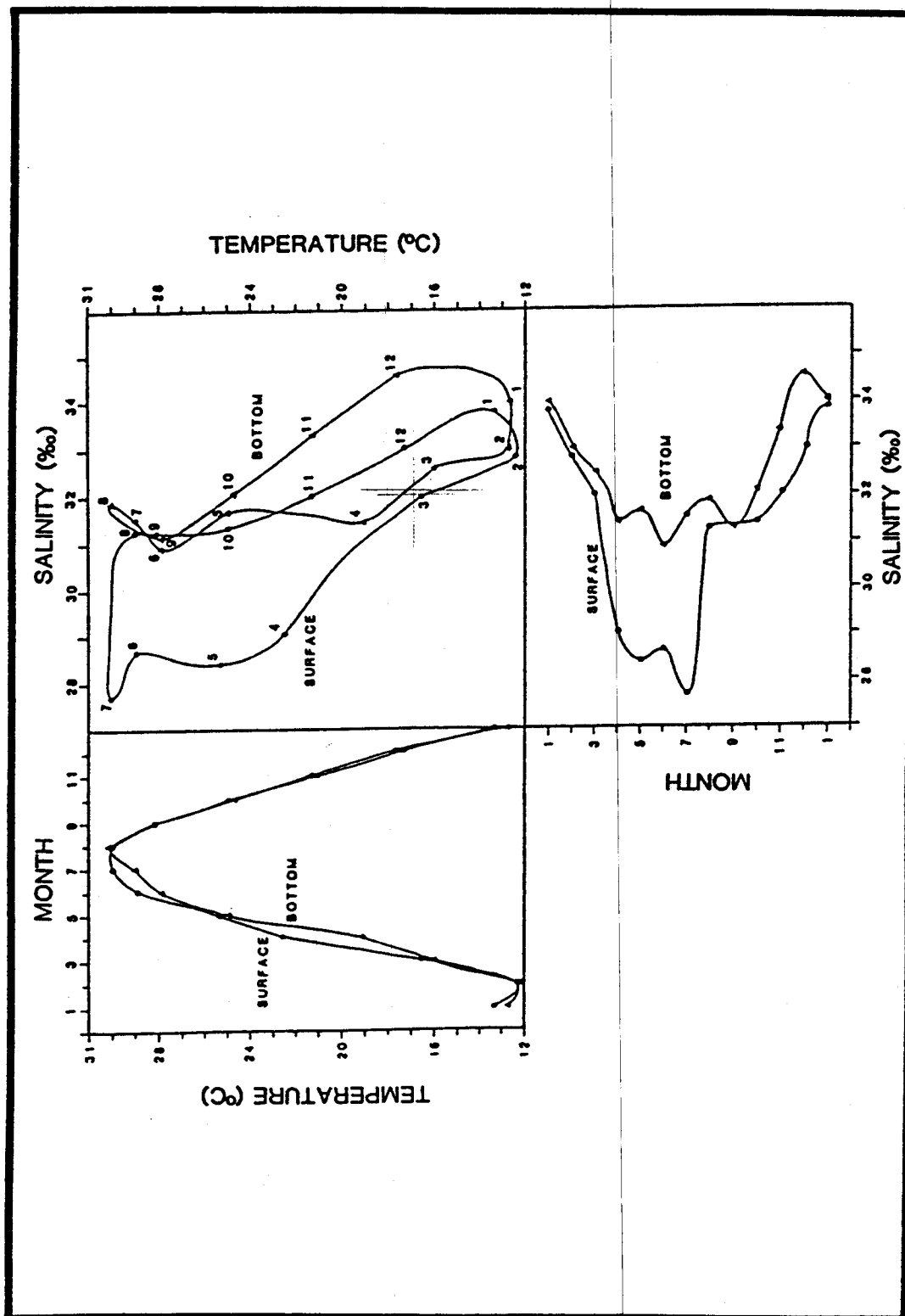


Figure 4.26. Annual progressions of mean salinity (‰) and mean temperature (°C) for near-surface and near-bottom depths for a region about 25 km off Cameron, LA in 10 m of water (West Hackberry station 22 and GUS III station E-1 data). Data are also shown in a T-S diagram in which sigma-t lines are entered (From: Kelly et al., 1983b).

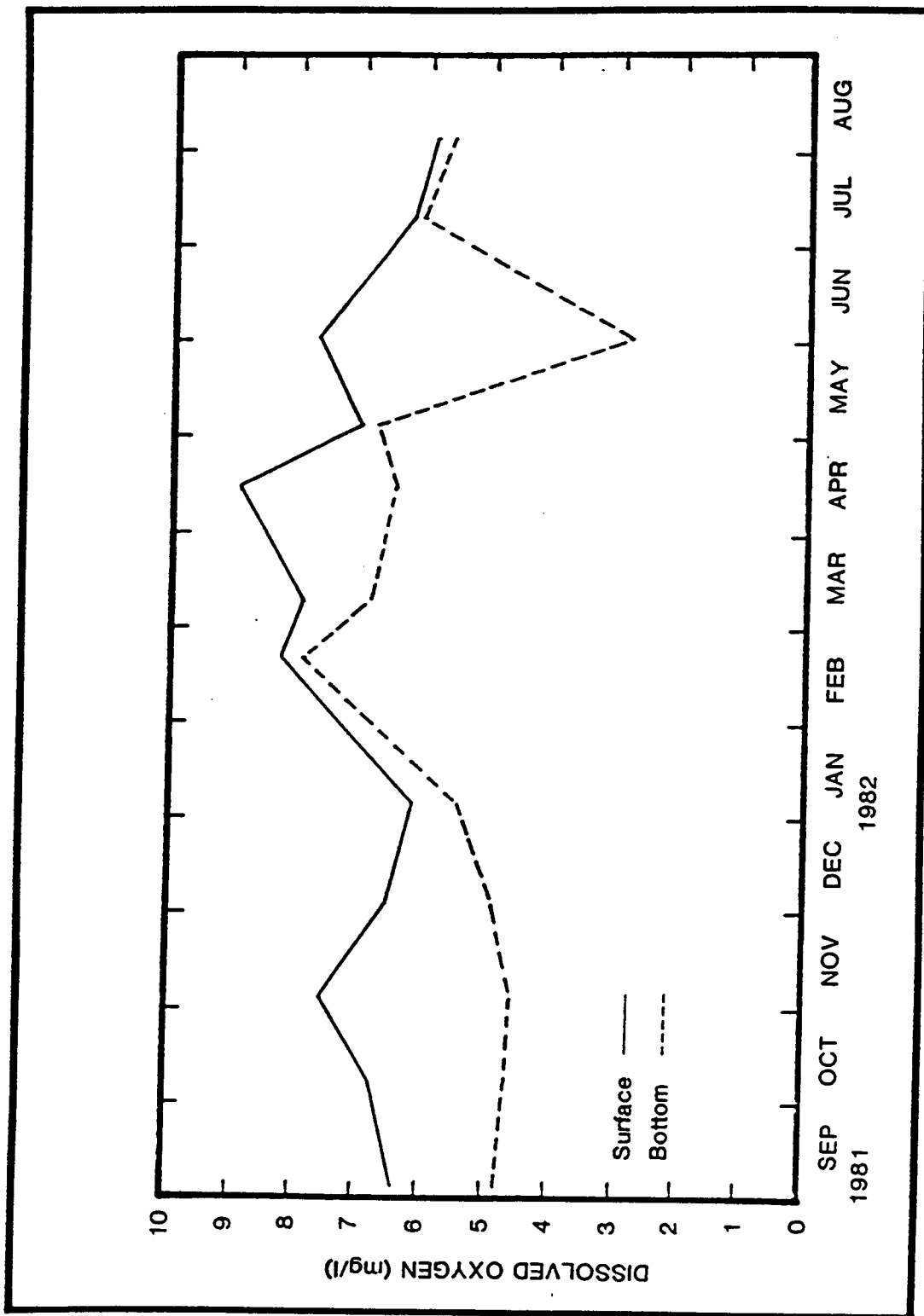


Figure 4.28. Dissolved oxygen concentrations measured at the surface and on the bottom about 25 km off Freeport, TX in 22 m of water by monthly hydrographic cruises (From: Kelly et al., 1983a).

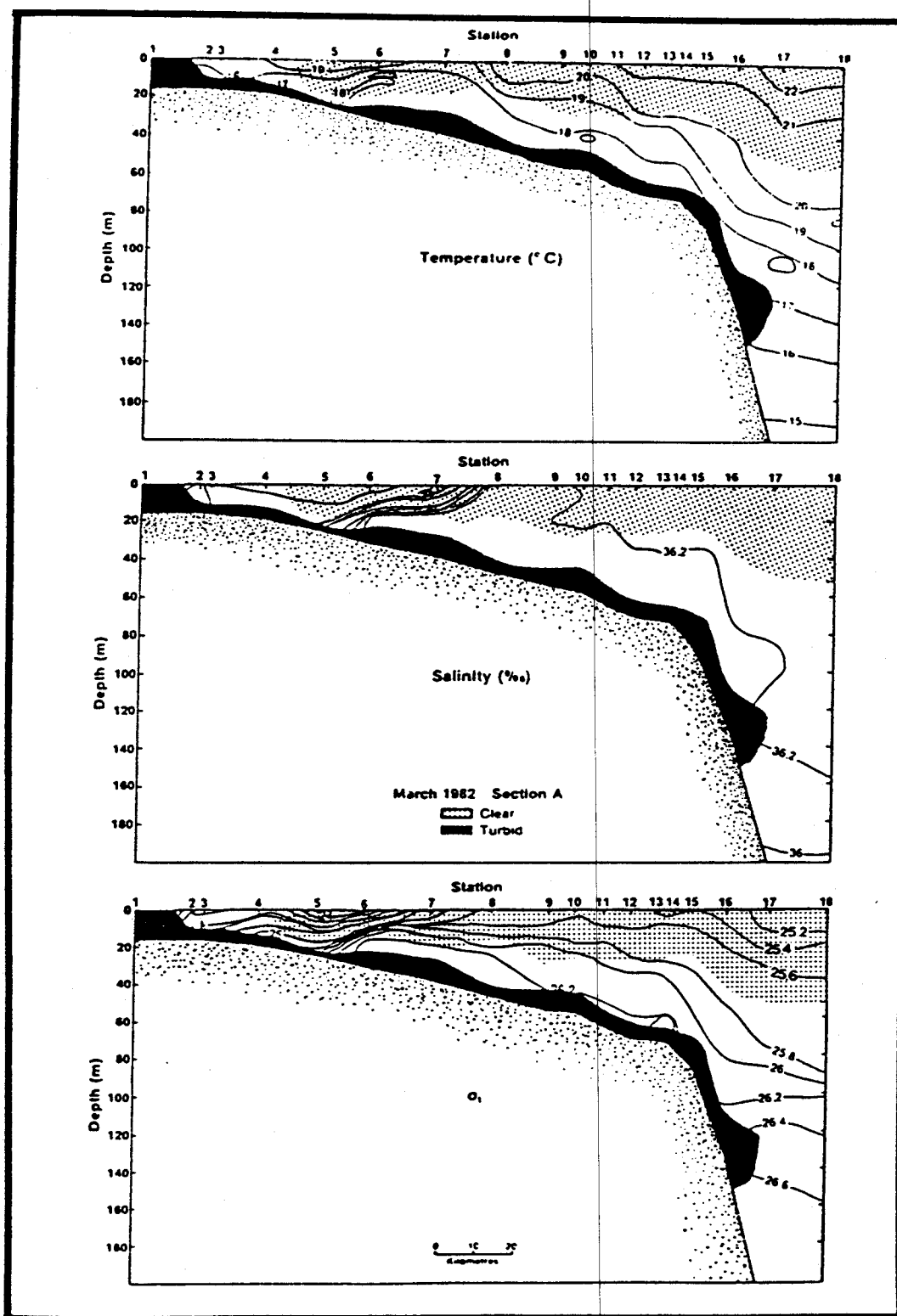
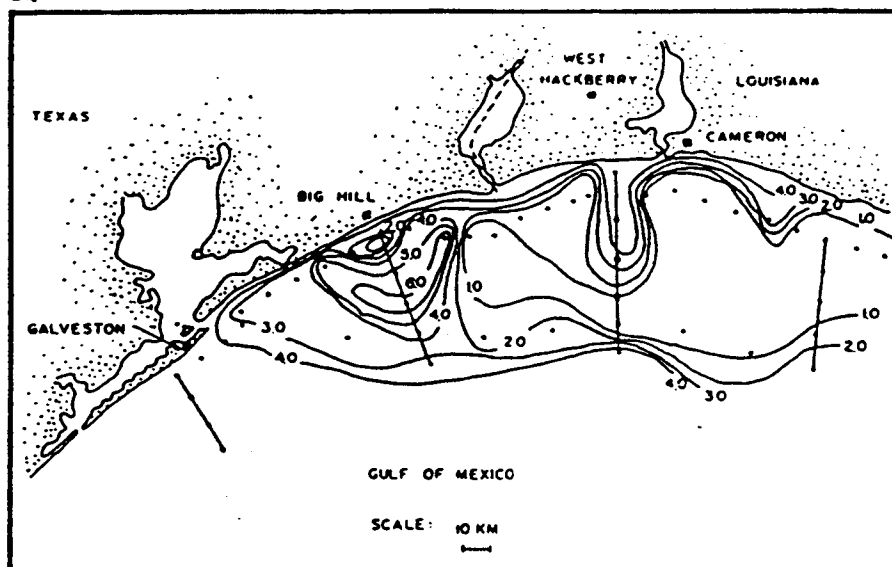


Figure 4.27. Cross-shelf vertical section of temperature, salinity, and sigma-t along a section running south-southeast from Galveston Island, TX, March 1982. Regions of turbid and clear water are indicated by dark and light shading, respectively; unshaded areas are of intermediate transparency (From: Rezak et al., 1985).

a)



b)

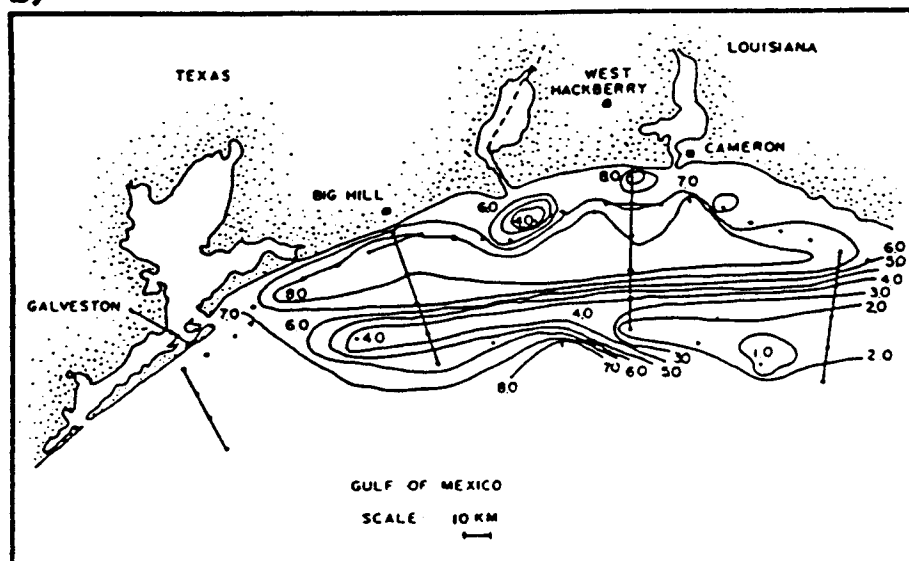


Figure 4.29. Dissolved oxygen concentrations (mg/l) at a) the bottom and b) 4 m above the bottom on 9 to 10 July 1984 (From: Pokryfki and Randall, 1987).

vertical mixing. Values typically range from 4 to 7 mg/l (Figure 4.28), but when strong vertical stratification develops (Figures 4.21 through 4.23), the bottom waters can become hypoxic (i.e., concentrations lower than 2 mg/l) and even anoxic.

Large areas of hypoxia develop almost annually in Louisiana coastal waters west of the delta according to the numerous reports in the literature that are reviewed by Pokryfki and Randall (1987). [Dennis et al. (1984), Rabalais et al. (1985) and Renaud (1985) have produced extensive bibliographies on hypoxia.] Hypoxia also occurs in Texas coastal waters as far west as Freeport, but less frequently. The major flood of the Mississippi in 1979 in conjunction with the summer advent of upwelling favorable winds and currents, produced hypoxic and anoxic bottom conditions in June and July off Freeport, TX that caused mass mortalities of benthic organisms (Harper et al., 1981).

Pokryfki and Randall (1987) measured the spatial extent of hypoxia in coastal waters from Galveston, TX to 74 km east of Cameron, LA in July 1974. Their results for concentrations on the bottom and 4 m above the bottom are shown in Figure 4.29. They note that the hypoxic mass lay entirely inshore of the 20-m isobath and was not an extension of the oxygen minimum layer that impinges on the outer shelf from the deep Gulf. Their study also developed a time-series model that uses river discharge and density gradient to predict concentrations of dissolved oxygen on the bottom during hypoxic events.

Boesch and Rabalais (1988) have systematically studied hypoxia on the northern shelves of the Gulf since 1985. Areas ranging from 7,000 to 10,000 km² were hypoxic during mid-summer surveys in 1985, 1986, and 1987. Bottom hypoxic conditions were found on the inner shelf deeper than 10 m and extending to about 30 m, from the Mississippi River Delta to the upper Texas coast. Hypoxic conditions formed as early as April and extended as late as October, depending on hurricane activity. They suggest that the massive enrichment of the Mississippi River discharge in levels of nitrate may have contributed to an intensification of shelf hypoxia since the 1960's. The increase in nitrate is attributable to increased use of agricultural fertilizers, waste discharges, and atmospheric precipitation.

Although the spatial distribution of hypoxia has been mapped with fair resolution at times, virtually nothing is known about its local time-rate-of-change or the advection of hypoxic water masses.

4.6 SURFACE WAVES

A large amount of the total wave energy of the sea surface is associated with gravity waves (gravity is the principal restoring force), which are generated by the action of the wind. Gravity waves are separated into two categories: seas, when the waves are under the influence of wind in a generating area, and swell, when waves move out of a generating area and are no longer subjected to significant wind action (U.S. Army Corps of Engineers, Coastal Engineering Research Center, 1975). Seas on the shelf generally have steep waves with short periods and lengths; the waves associated with swell have relatively long periods

and wavelengths and, until they reach shallow water, low amplitudes. The height, period, and length of wind waves are determined by fetch, wind speed, wind duration, and decay distance. Water depth, if shallow enough, will also affect the size of wave that is generated.

The winds from the southeast and south associated with the flow of MT air around the west side of the Bermuda High have considerable fetch, and are quite persistent at times, but wind speed is typically low to moderate. These winds generate waves that are usually in the range 0.5 to 1.5 m. The fetch of the strong winds from the north associated with frontal passages is small near the coast and increases toward the outer shelf. The northers typically generate 2- to 3-m waves nearshore and 4- to 6-m waves on the outer shelf. The most common periods associated with non-hurricane waves in the Gulf are between 4 and 6 s, with only a very few above 10 s. The extreme wind speeds of hurricanes generate waves of 7 to 10 m or more, with periods of 9 to 13 s.

Wave statistics and climatologies are generated from a variety of data sources, such as ship observations, hindcasts, and direct measurements by various types of instruments. Quayle and Fulbright (1977) gives the annual percent frequency of wave heights in half-meter increments for sea surface meteorological observation (SSMO) areas 27, 28, and 29 based on ship observations (Table 4.6). These three SSMO areas cover the Texas-Louisiana shelf. The COE Waterways Experiment Station has just completed a hindcast of wave climate for the U.S. coastal waters in the Gulf of Mexico as part of its Wave Information Study (WIS). WIS Report 18 (Hubertz et al., 1988), to be released about December 1988, summarizes twenty years (1956 to 1975) of hindcast significant height, peak period, and mean direction for various locations in four data products: percent occurrence tables, wave rose diagrams, mean and largest significant height and 20-year statistics tables, and return period table. Hurricane conditions are excluded. The report also statistically compares the hindcast data with direct observations by gauges and buoys from different periods of time. Hurricane hindcast methodology and wave statistics for the Atlantic and Gulf hurricanes from 1956 to 1975 are given in a separate report, WIS Report 19 (Abel et al., 1988), which is expected to be released about December 1988. Reese and Cardone (1982) also describe and compare examples of extreme wave events in the Gulf of Mexico computed from model hindcasts and measured by calibrated devices.

The wave climatology measured at the Pleasure Pier at Galveston is detailed by Thompson and Harris (1972) and Thompson (1977). Monthly means of significant wave height are less than 0.7 m, monthly maxima are 1 to 2 m, and wave periods are typically about 5 s. (Significant height is defined to be the average of the highest one-third of the waves in a record.) Another example of wave observations in inner-shelf waters is the work of Frey et al. (1981) at the West Hackberry DOE/SPR site in 10 m of water. Figure 4.30 from their work shows the effect on significant and maximum wave heights of the passage of three successive polar fronts. For estimates of wave characteristics near the outer shelf and slope the data from NOAA Data Buoy Office (NDBO) Buoy 42002 are probably applicable, even though the buoy is located about 200 km south of the shelf break. Yamazaki and Herbich (1985) present monthly wave

Table 4.6. Annual percent frequency of wave heights in various categories, based on ship observations in SSNO areas 27, 28, and 29 (From: Quayle and Fulbright, 1977).

Area Number	Wave Height (m)													Total OBS	Average Height (m)
	0	%	1	1%	2	2%	3	3%	4-4%	5-5%	6-6%	7-7%	8-9%		
27	14.2	25.6	29.7	17.3	7.1	3.4	1.3	0.7	0.6	0.1	0.1	*	*	77,800	1
28	13.9	26.1	30.4	17.1	6.7	3.3	1.2	0.6	0.5	0.1	0.1	*	*	47,507	1
29	8.3	24.2	33.3	18.3	8.2	4.2	1.7	0.9	0.7	0.1	0.1	*	0	13,371	1

* Less than 0.05% but greater than 0.

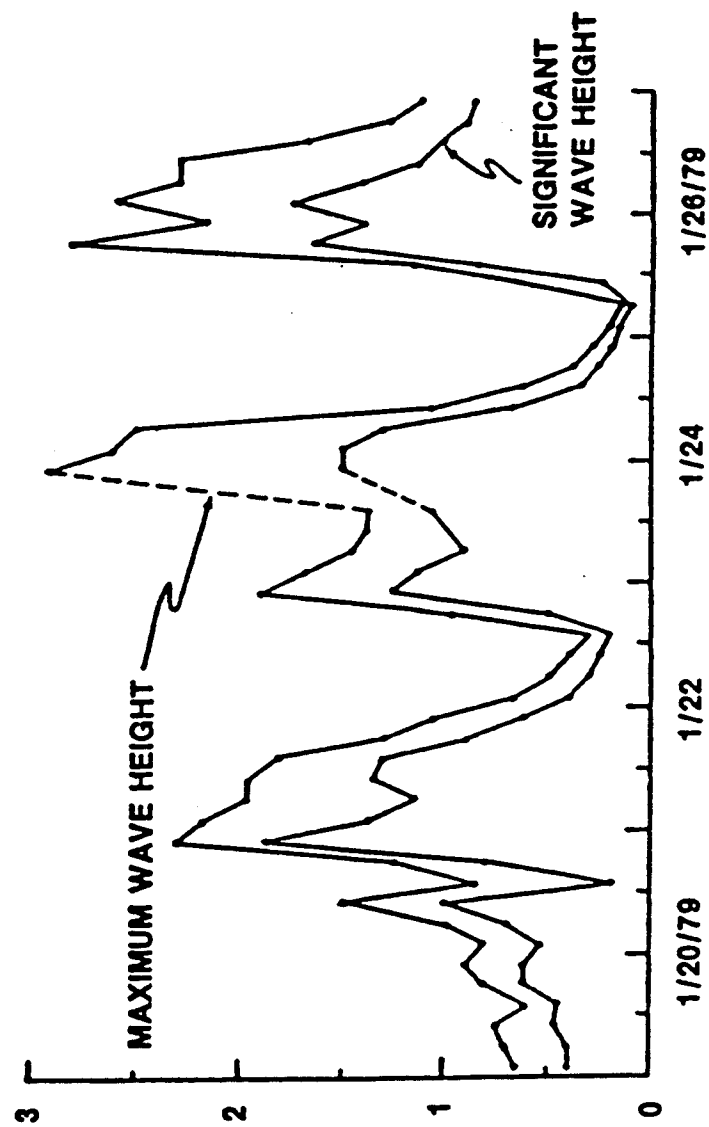


Figure 4.30. Significant wave height and maximum wave height observed at the West Hackberry site during the passage of three successive polar fronts from 19 to 27 January 1979 (From: Frey et al., 1981).

statistics for a 30-month period at NDBO Buoy 42002; monthly mean significant heights are typically 1 to 1.5 m, monthly maximum heights are usually 3 to 4 m but occasionally 6 to 8 m, and average wave periods range from 4.2 to 5.4 s.

4.7 TIDES AND SEA LEVEL

Sea level, the height of the sea surface relative to a fixed datum that is derived by averaging the instantaneous observations (e.g., hourly) of the height of the sea surface over periods of a day, month, year, etc., fluctuates because of the effects of long-period astronomic tides, local winds, local addition of mass, changes in atmospheric pressure, and changes in temperature and salinity, and thus, density, of the water column (steric effect). Marmer (1954) discussed the long-period fluctuations of sea level observed at tide stations around the Gulf of Mexico, and noted annual ranges and month-to-month variations. Along the Texas-Louisiana coast the annual range is about 25 cm. The annual variation has a semi-annual component; the absolute minimum and maximum values occur in January and September, respectively, and relative maximum and minimum values occur in May and July, respectively. Whitaker (1971) computed the seasonal variations of steric and recorded sea level at various stations around the Gulf, including Galveston, TX (Figure 4.31). For Galveston, he found that the regional steric sea levels relative to 150 db and 1000 db accounted for 55% and 77%, respectively, of the recorded sea-level range. However, the variations in steric sea level do not account for the July minimum in recorded sea level. Chew (1964) argued that the July minimum may be related to the upwelling coastal regime driven by local winds, which is consistent with the results of Cochran and Kelly (1986b).

For periods shorter than one month, coastal variability of sea level is driven substantially by wind forcing. Chuang and Wiseman (1983) examined the sea level response at Galveston and Eugene Island to the wind forcing associated with frontal passages. They found that at Galveston the alongshore component of wind, through Ekman convergence, drives sea-level fluctuations, but at Eugene Island, where a broad region of the shelf is much shallower, the cross-shelf component of wind, through direct wind set-up, drives the sea-level fluctuations.

Tropical cyclones produce extreme changes in sea level. Bunpamong and Reid (1985) investigated the gradual rise in water level along the coast that precedes the arrival of a hurricane (forerunner surge), and hurricane-induced storm surge has been modeled in various ways (Wanstrath, 1975; Ebersole, 1985).

Reid and Whitaker (1981) reviewed the tidal regime in the Gulf of Mexico and developed a numerical tide model that provides good agreement of the important tidal constituents with those deduced from open coastal measurements. Their numerical investigations show that the diurnal tides, K_1 and O_1 , have nearly uniform amplitudes and phases over the whole Gulf and are driven primarily by the in-phase volume transports through the Yucatan and Florida Straits. The contribution due to direct forcing is only about 15%. On the other hand, direct forcing accounts for about 55% of the signal of the semidiurnal tides, M_2 and S_2 . The M_2

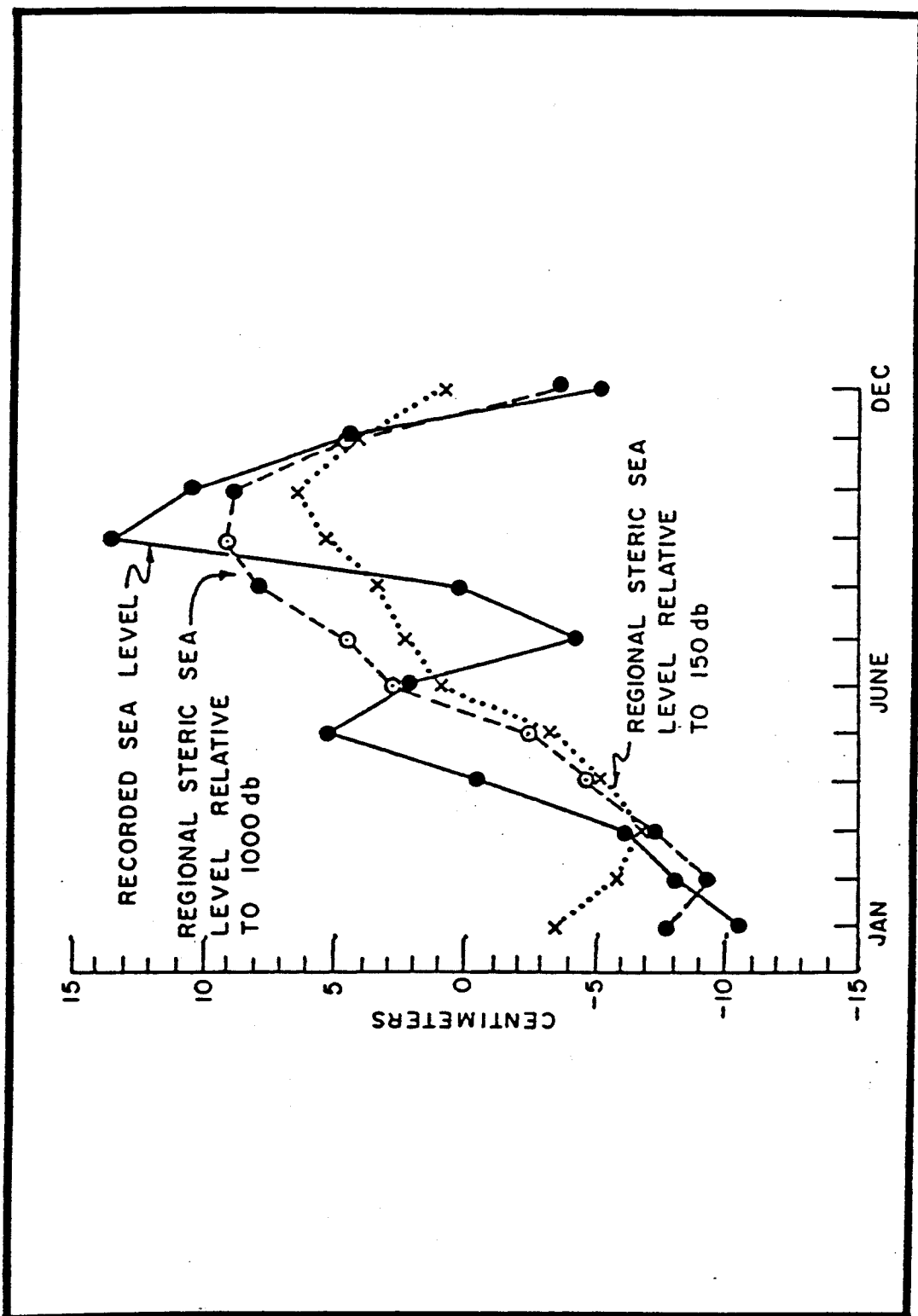


Figure 4.31. Annual variation of recorded sea level, regional steric sea level relative to 150 db, and regional steric sea level relative to 1000 db for Galveston, TX (From: Whitaker, 1971).

tide has negligible amplitude in the deep central Gulf and maximum amplitudes along the coast. Phase propagates counterclockwise around the amphidromic point. Table 4.7 gives the observed amplitudes and phases of the major tidal constituents at Southwest Pass, Point au Fer, Galveston, and Port Aransas (Reid and Whitaker, 1981). Along the Texas-Louisiana coast the amplitude of the M_2 tide is largest off Calcasieu Pass and decreases to the east and west. As a result of the relative magnitudes of the diurnal and semidiurnal constituents, the tide is characterized as mixed off Calcasieu Pass and becomes mainly diurnal off the Mississippi Delta and to the west of Galveston (Marmer, 1954; Hicks, 1977; Frey et al., 1981).

Figure 4.32 illustrates several of the processes noted above. It shows the sea-surface elevation for February 1984 at the Bryan Mound site, computed from pressure data measured at the bottom, time series of temperature and salinity at three depths, and atmospheric pressure (the inverted barometer effect is included). The colder water in winter decreased total steric sea level from the long-term mean by about 5 cm. During the last few days of February, winds were strong and directed upcoast and offshore (Kelly et al., 1985a); this wind pattern drove an upwelling event that decreased nearshore sea-surface elevation by about 70 cm (a 10 mb rise in atmospheric pressure forced an additional 10 cm decrease). Tides in Figure 4.32 are mainly diurnal. The maximum range, about 75 cm, occurs during tropic tides, when the K_1 and O_1 tides conspire.

4.8 CURRENTS

4.8.1 Inner Shelf

Wind stress and fresh water influx from the Mississippi-Atchafalaya River System are the dominant factors that control the long-period motion, hydrographic characteristics, and much of the short-period variability of the waters on the inner part of the Texas-Louisiana shelf. The basic features of the circulation on the inner Texas-Louisiana shelf have been known for quite some time. For example, Sweitzer (1898), in discussing the placement of harbor jetties on the Texas coast stated:

"Hence, it may be concluded that the currents flowing westward are on the northern shore of the Gulf, and those flowing north along the western shore meet at some point near Galveston, the location of which is variable, as it is dependent upon the direction and force of the winds."

Also, Leipper (1954) noted that this convergence is seen in the U.S. Navy pilot charts. Drift bottle studies by Kimsey and Temple (1963, 1964), Watson and Behrens (1970), Hunter et al. (1974), Hill et al. (1975), and Hill and Garrison (1978) provided a more detailed, but still qualitative, description of the inner-shelf circulation. In 1973, Smith (1975, 1977, 1978a, 1979, 1980b) initiated a series of studies off Aransas Pass to directly measure currents and quantitatively describe their temporal and spatial variability, vertical structure, and relation to wind stress.

Table 4.7. Observed amplitude (cm) and phase (relative to Greenwich) for the five principal tidal constituents at four locations along the Texas-Louisiana coast (From: Reid and Whitaker, 1981).

Station	Tidal Constituent				
	K_1	O_1	P_1	M_2	S_2
Southwest Pass	13.5 (20.4°)	13.4 (13.4°)	4.9 (19.5°)	1.7 (127.2°)	0.9 (111.7°)
Point au Fer	16.6 (23.7°)	15.3 (19.7°)	5.6 (18.1°)	9.9 (240.4°)	3.4 (234.7°)
Galveston	16.7 (28.3°)	15.5 (20.3°)	4.3 (13.5°)	13.5 (275.1°)	4.0 (273.0°)
Port Aransas	15.4 (26.4°)	15.1 (20.1°)	4.8 (26.0°)	7.7 (262.3°)	2.2 (264.6°)

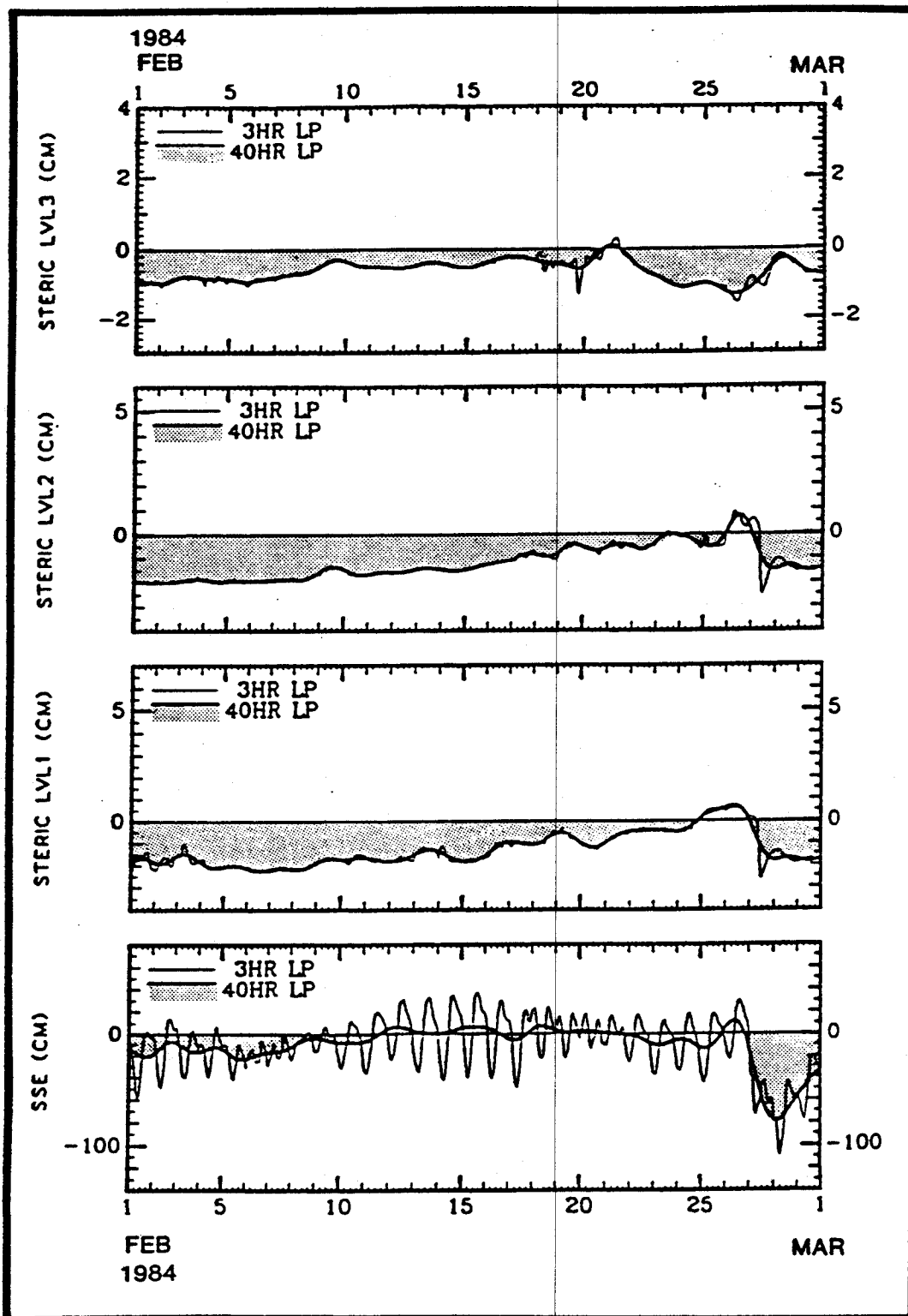


Figure 4.32. Time-series plots, for February 1984, of sea-surface elevation (bottom frame) and the contribution to sea-surface elevation by density fluctuations (cm) in the top, middle, and bottom layers of the water column (From: Kelly et al., 1985a).

Because the annual progression in alongshore wind stress (Section 4.3.1) is similar from Port Aransas to Cameron, the inner-shelf region between these points has a single current regime, which Cochrane and Kelly (1986b) have called the "Freeport regime," with downcoast (southeast) flow prevailing except for a period of upcoast (northeast) flow in spring and summer. Off Port Aransas the period is long, but off Freeport and Cameron it is limited to July.

Currents and Wind Stress off Freeport, TX

Numerous studies of currents along the Texas-Louisiana coast demonstrate the strong coherence between the alongshore components of wind stress and current (Smith 1978a, 1979; Kelly et al., 1982; Crout et al., 1984; Lewis and Reid, 1985; Cochrane and Kelly, 1986b) that results from the combination of wind stress, the Earth's rotation and the presence of the coastal boundary (e.g., Csanady 1982). A visual example of the dynamics comes from observations collected in 20 m of water about 20 km off Freeport, TX at the DOE/SPR Bryan Mound site. In Figure 4.33, the alongshore component of wind stress (Figure 4.33a) during the period 18 to 30 June 1980 is upcoast (toward Louisiana) and drives an offshore directed Ekman transport in the near-surface layer (Figure 4.33c), with a compensating onshore drift in the bottom layer (Figure 4.33d). Water level at the coast (not shown) decreases, and the cross-shelf pressure gradient drives the alongshore component of current (Figure 4.33b) in the upcoast direction. When the alongshore component of wind stress is downcoast (toward Mexico), everything reverses to a downwelling regime, as can be seen in Figure 4.33 for briefer periods earlier in the month. The dramatic effects of downwelling and upwelling on the hydrography of the inner shelf were described in Section 4.5.

The DOE/SPR studies off Freeport, TX and Cameron, LA produced the most statistically reliable evidence of the relation between the alongshore component of wind stress and the coastal currents because of the length of the observations. The seasonal variations of wind stress and the effect of these variations on both the alongshore component and the annual migration of a convergence zone were described in Section 4.3.1. Figure 4.34a shows the annual progressions of alongshore components of wind stress and current off Freeport, TX in terms of monthly means based on hourly values of 3-h low-passed data obtained between August 1978 and September 1984. Both components are positive (upcoast) in July, near zero in June and August, and negative (downcoast) during the rest of the year. By the criterion of standard error, the changes from May to June and from August to September are quite significant, and the values for July are significantly different from zero.

As Figure 4.33 suggests, there is considerable temporal variability about the monthly mean values. Cochrane and Kelly (1986b) examined this variability, and the relation between wind stress and currents for sub-inertial frequencies below 0.5 cycles per day. They applied the methods of cross-spectral analysis to one-year time series of wind stress and top and bottom currents collected at the Bryan Mound site during 1981. Their results agree in most respects with the above model. Alongshore stress drives currents effectively at all sub-inertial

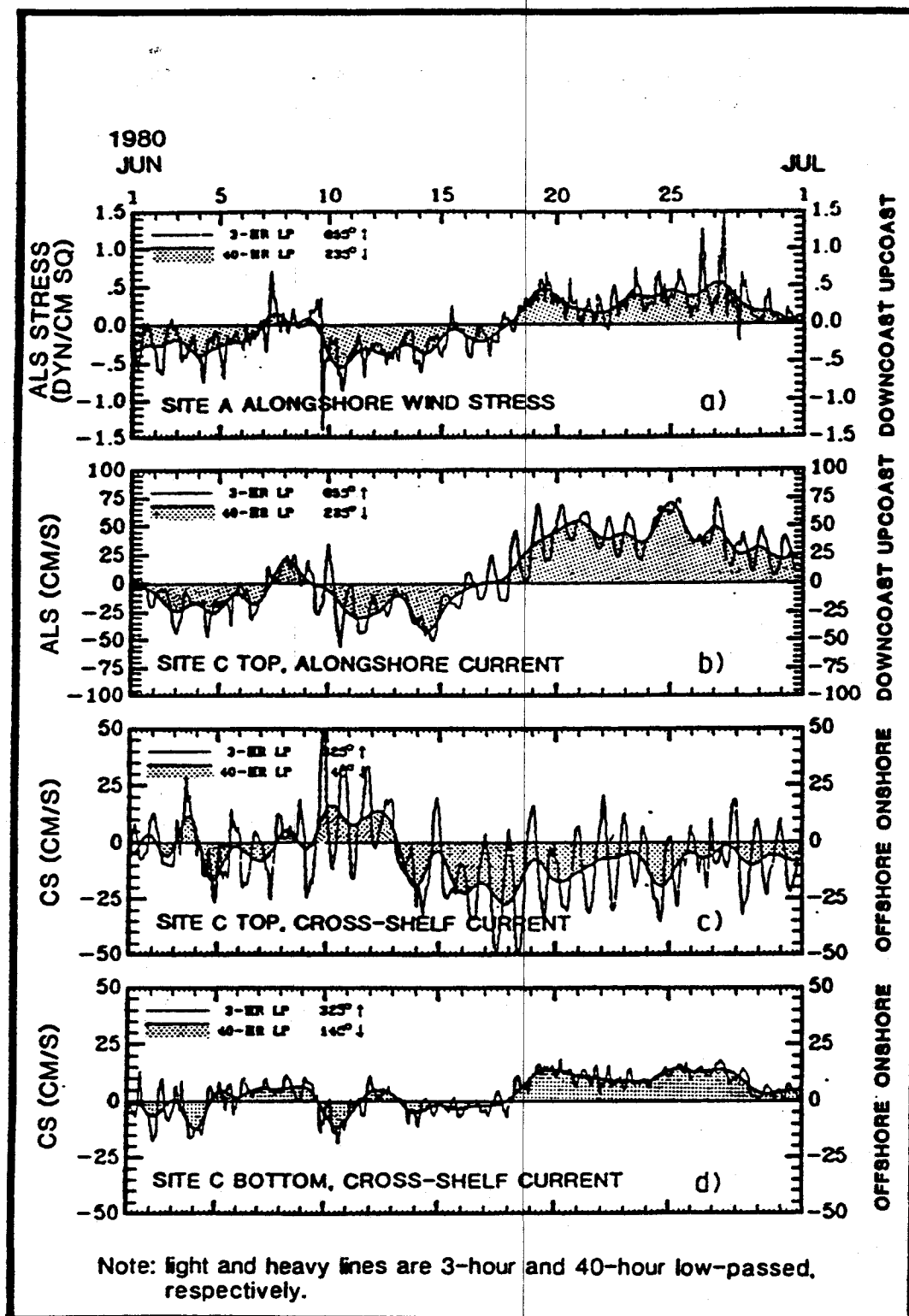


Figure 4.33. An example of the response of currents to the alongshore component of wind stress at Bryan Mound site, water depth 22 m: a) alongshore (positive toward 55° true) component of wind stress; b) alongshore component of near-surface current (3.7 m below the surface); c) cross-shelf (positive toward 325° true) component of near-surface current; and d) cross-shelf component of near-bottom current (1.8 m above the bottom). (Modified from: Kelly et al, 1982).

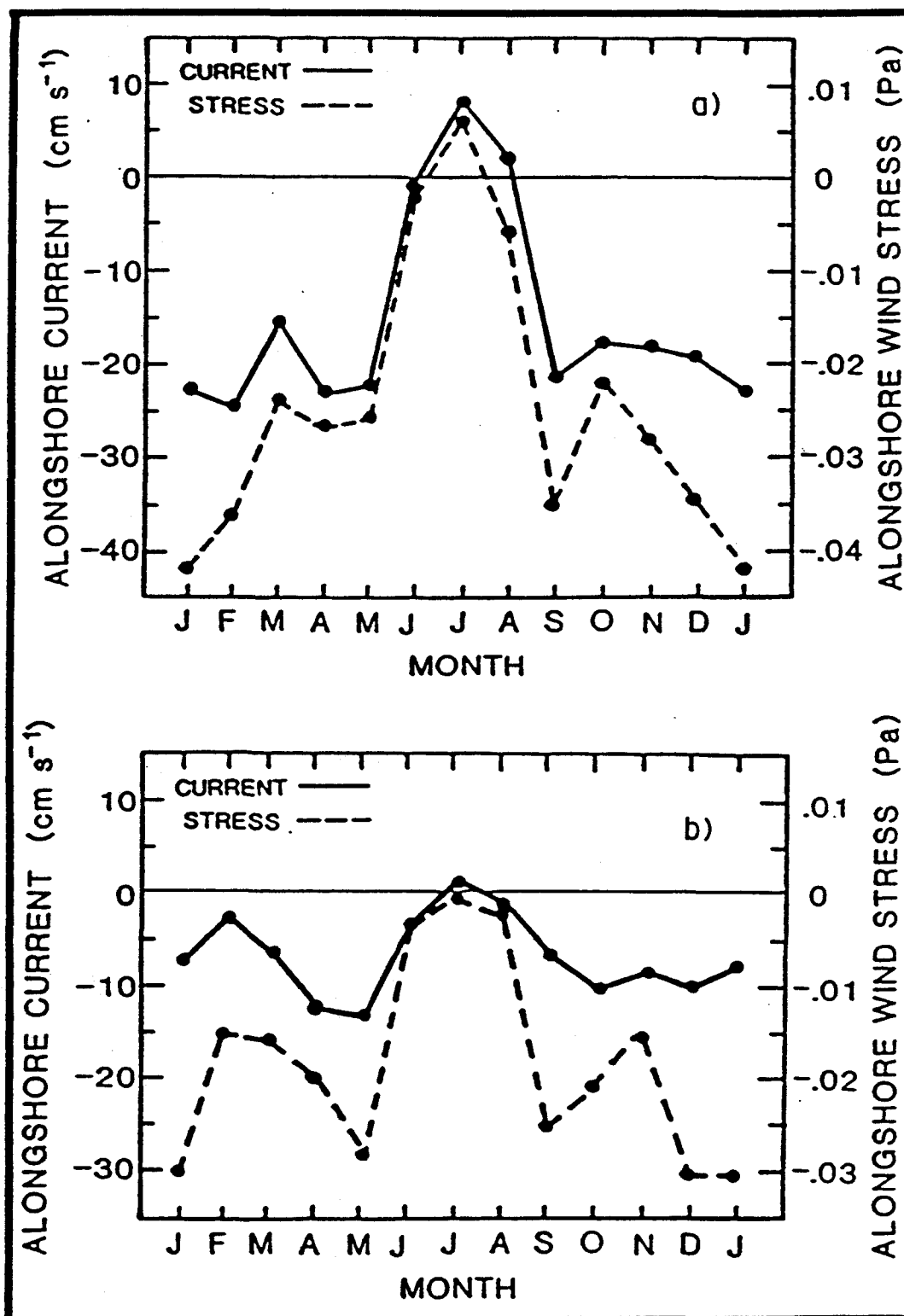


Figure 4.34. Monthly mean alongshore component of wind stress together with monthly mean alongshore component of surface current (3.7 m below sea surface); a) at the Bryan Mound site off Freeport, TX; the positive alongshore direction is 55° clockwise from north; and b) at the West Hackberry site off Cameron, LA; the positive alongshore direction is 86° clockwise from north (From: Cochrane and Kelly, 1986b).

frequencies, but cross-shelf stress does not. (Cross-shelf wind stress induces Ekman transport with a net alongshore direction, and so does not produce a coastal jet.) Alongshore wind stress is coherent with cross-shelf bottom currents, and the phase indicates flow to the left of stress, that is, in the direction expected for upwelling/downwelling regimes described above. The one disagreement with the model is the lack of coherence between cross-shelf currents at the surface and alongshore wind. However, the current meter mooring was located near the mean position of the intense frontal zone separating fresher coastal water from saltier offshore water, and they attribute the low coherence to the effects of the local baroclinic currents caused by the dynamics of the frontal zone.

The DOE/SPR studies amassed a very large data base concerning the characteristics of the inner-shelf region from about Galveston, TX to Cameron, LA. The results are described in a series of annual reports that present the data in a variety of useful formats. However, a comprehensive climatology synthesizing, for example, all of the current meter data does not exist. Therefore, to provide a quantitative example of the magnitudes of the near-surface and near-bottom currents, Table 4.8 shows the basic statistics in intervals of one month from September 1983 through August 1984 for the time series data from the current meters located 3.6 m below the sea surface and 1.6 m above the bottom. It is stressed that the interannual variability is significant at times (Smith, 1978a, 1980b; Kelly et al., 1985a,b).

Currents and Wind Stress off Cameron, LA

The annual progressions off Cameron, LA, from data collected between February 1981 and January 1985, are shown in Figure 4.34B. Similar to the results for Freeport, the means for current are positive (upcoast) only in July, and those for stress are very nearly zero although not positive in July. The two sets of means are clearly correlated. Both the stress and current means are smaller off Cameron than off Freeport, in large part because of the change in the orientation of the coastline in relation to the mean direction of the wind. The weaker response of current to alongshore wind in extensive regions of shallower waters (Chuang and Wiseman, 1983) and an increase in bottom friction (Chuang and Wiseman, 1983; Cochrane and Kelly, 1986b) are also factors.

A quantitative example of the basic statistics is provided for a small subset of the data base (Tables 4.9 and 4.10).

Coherence Between Currents off Freeport and Cameron

The West Hackberry and Bryan Mound sites are separated by a distance of about 200 km. Kelly et al. (1983b) investigated the coherence between the near-bottom currents of the two sites using common series of 231-days in length. Their results are shown in Figure 4.35 for the alongshelf components. High coherence exists between the alongshelf components at periods of about 10 days and 3.3 days. They also found significant coherence between the cross-shelf components at 10 days and weaker but still significant coherence at 3.5 days. Smith (1980b)

Table 4.8. Basic statistics by month for the hourly time series of 3-h low-pass filtered current velocity at depths of 3.7 m (CT) and 20 m (CU) at Bryan Mound site C in 22 m of water during the period 1 September 1983 through 31 August 1984 (From: Kelly et al., 1985a). Positive along-shelf direction is toward 55° true, and positive cross-shelf direction is toward 325° true. N = number of observations; MIN = minimum; MAX = maximum; MEAN = algebraic mean; SD = standard deviation.

SITE CT		ALONGSHELF (CM/S)					CROSS-SHELF (CM/S)				
START	STOP	N	MIN	MAX	MEAN	S.D.	N	MIN	MAX	MEAN	S.D.
09-01-83(0000)	09-30-83(2300)	719	-65.	45.	-19.3	21.7	719	-41.	32.	-2.3	13.4
10-01-83(0000)	10-31-83(2300)	744	-60.	23.	-17.7	18.7	744	-39.	38.	0.7	10.5
11-01-83(0000)	11-30-83(2300)	720	-46.	39.	-6.2	15.4	720	-27.	16.	-1.2	5.9
12-01-83(0000)	12-31-83(2300)	744	-58.	30.	-12.0	19.4	744	-36.	17.	-1.1	6.4
01-01-84(0000)	01-31-84(2300)	744	-79.	19.	-29.9	23.7	744	-36.	24.	1.0	8.0
02-01-84(0000)	02-29-84(2300)	696	-52.	78.	-11.7	24.6	696	-24.	51.	2.0	9.9
03-01-84(0000)	03-31-84(2300)	744	-62.	65.	-9.9	26.1	744	-33.	36.	3.0	10.7
04-01-84(0000)	04-30-84(2300)	720	-59.	54.	-8.8	26.7	720	-24.	30.	0.9	9.3
05-01-84(0000)	05-31-84(2300)	744	-71.	18.	-31.6	18.6	744	-37.	44.	1.5	13.9
06-01-84(0000)	06-30-84(2300)	720	-58.	56.	-0.7	30.0	720	-46.	38.	-0.6	14.9
07-01-84(0000)	07-31-84(2300)	744	-59.	62.	8.4	21.3	744	-47.	34.	-0.4	14.2
08-01-84(0000)	08-31-84(2300)	744	-38.	72.	7.6	22.1	744	-37.	38.	-0.8	10.9

SITE CU		ALONGSHELF (CM/S)					CROSS-SHELF (CM/S)				
START	STOP	N	MIN	MAX	MEAN	S.D.	N	MIN	MAX	MEAN	S.D.
09-01-83(0000)	09-30-83(2300)	720	-36.	18.	-9.8	11.8	720	-31.	21.	-3.9	9.7
10-01-83(0000)	10-31-83(2300)	744	-31.	25.	-2.8	12.2	744	-29.	22.	-1.1	7.5
11-01-83(0000)	11-30-83(2300)	720	-31.	32.	-2.4	16.1	720	-15.	25.	1.2	7.7
12-01-83(0000)	12-31-83(2300)	744	-41.	30.	-4.7	15.7	744	-21.	19.	2.8	6.0
01-01-84(0000)	01-31-84(2300)	744	-46.	26.	-8.7	14.8	744	-20.	19.	-0.4	6.8
02-01-84(0000)	02-29-84(2300)	696	-33.	39.	-0.2	14.1	696	-21.	63.	3.0	11.8
03-01-84(0000)	03-31-84(2300)	744	-28.	44.	-0.6	15.4	744	-25.	37.	1.2	11.3
04-01-84(0000)	04-30-84(2300)	720	-43.	28.	-5.8	17.3	720	-24.	20.	-0.3	9.7
05-01-84(0000)	05-31-84(2300)	744	-43.	25.	-3.4	13.0	744	-36.	18.	-1.6	10.5
06-01-84(0000)	06-30-84(2300)	720	-28.	23.	-0.6	12.5	720	-22.	19.	1.0	8.6
07-01-84(0000)	07-31-84(2300)	744	-34.	24.	-3.5	11.6	744	-22.	28.	-1.1	8.9
08-01-84(0000)	08-31-84(2300)	744	-25.	25.	3.3	12.2	744	-20.	23.	3.2	8.5

Table 4.9. Basic statistics in intervals of one month for the hourly time series of 3-h low-pass filtered current velocity at a depth of 3.7 m (DT) at West Hackberry site D in 10 m of water during the period 1 May 1982 through 30 November 1983 (From: Kelly et al., 1984c). Positive along-shelf direction is toward 86° true, and positive cross-shelf direction is toward 356° true. N = number of observations; MIN = minimum; MAX = maximum; MEAN = algebraic mean; SD = standard deviation.

SITE DT		ALONGSHORE (CM/S)					CROSS SHELF (CM/S)				
START	STOP	N	MIN	MAX	MEAN	S.D.	N	MIN	MAX	MEAN	S.D.
05-01-82(0000)	05-31-82(2300)	744	-62.	35.	-11.6	16.5	744	-30.	31.	-0.3	9.6
06-01-82(0000)	06-30-82(2300)	720	-23.	43.	7.3	13.0	720	-19.	22.	1.9	6.7
07-01-82(0000)	07-31-82(2300)	744	-23.	24.	2.8	8.7	744	-22.	26.	0.5	6.6
08-01-82(0000)	08-31-82(2300)	744	-50.	24.	-0.6	12.2	744	-15.	17.	-0.6	5.1
09-01-82(0000)	09-30-82(2300)	720	-76.	19.	-11.8	14.4	720	-28.	26.	2.4	6.9
10-01-82(0000)	10-31-82(2300)	744	-51.	21.	-11.8	12.5	744	-22.	22.	2.3	5.7
11-01-82(0000)	11-30-82(2300)	720	-40.	25.	-10.7	11.5	720	-19.	27.	3.0	7.0
12-01-82(0000)	12-31-82(2300)	744	-67.	31.	-13.4	18.3	744	-18.	30.	3.0	8.4
01-01-83(0000)	01-31-83(2300)	744	-58.	23.	-5.2	17.3	744	-23.	27.	3.1	8.3
02-01-83(0000)	02-28-83(2300)	672	-61.	45.	-0.7	19.1	672	-24.	27.	2.7	9.3
03-01-83(0000)	03-31-83(2300)	744	-85.	28.	-5.3	18.5	744	-18.	28.	1.7	6.9
04-01-83(0000)	04-30-83(2300)	720	-60.	73.	-6.8	22.6	720	-21.	25.	4.0	8.2
05-01-83(0000)	05-31-83(2300)	744	-80.	30.	-20.1	21.6	744	-35.	46.	2.9	12.0
06-01-83(0000)	06-30-83(2300)	720	-60.	24.	-8.5	16.1	720	-33.	29.	-0.3	11.4
07-01-83(0000)	07-31-83(2300)	744	-58.	30.	0.4	14.6	744	-27.	27.	0.4	9.9
08-01-83(0000)	08-31-83(2300)	744	-43.	42.	0.6	12.9	744	-19.	23.	1.6	8.0
09-01-83(0000)	09-30-83(2300)	720	-37.	18.	-5.3	8.9	720	-17.	19.	2.4	5.6
10-01-83(0000)	10-31-83(2300)	744	-27.	26.	-1.7	10.4	744	-16.	21.	2.0	7.0
11-01-83(0000)	11-30-83(2300)	720	-56.	26.	-9.2	13.6	720	-10.	28.	2.9	5.0

Table 4.10. Basic statistics by month for the hourly time series of 3-h low-pass filtered current velocity at a depth of 8 m (DB) at West Hackberry site D in 10 m of water during the period 1 May 1982 through 30 November 1983 (From: Kelly et al., 1984c). Positive along-shelf direction is toward 86° true, and positive cross-shelf direction is toward 356° true. N = number of observations; MIN = minimum; MAX = maximum; MEAN = algebraic mean; SD = standard deviation.

SITE DB		ALONGSHORE (CM/S)					CROSS SHELF (CM/S)				
START	STOP	N	MIN	MAX	MEAN	S.D.	N	MIN	MAX	MEAN	S.D.
05-01-82(0000)	05-31-82(2300)	744	-57.	21.	-4.8	12.5	744	-37.	25.	-2.1	9.1
06-01-82(0000)	06-30-82(2300)	742	-17.	25.	4.0	7.7	742	-27.	20.	0.5	6.9
07-01-82(0000)	07-31-82(2300)	766	-22.	21.	4.4	8.2	766	-18.	16.	0.3	5.7
08-01-82(0000)	08-31-82(2300)	766	-26.	22.	-1.0	8.8	766	-18.	14.	0.4	5.1
09-01-82(0000)	09-30-82(2300)	742	-57.	26.	-6.4	9.4	742	-39.	16.	-0.4	6.5
10-01-82(0000)	10-31-82(2300)	766	-29.	18.	-6.7	9.8	766	-22.	17.	2.0	6.1
11-01-82(0000)	11-30-82(2300)	742	-32.	25.	-6.0	9.8	742	-23.	26.	1.6	6.8
12-01-82(0000)	12-31-82(2300)	766	-49.	24.	-7.4	12.8	766	-42.	25.	1.5	9.2
01-01-83(0000)	01-31-83(2300)	766	-38.	28.	-2.7	11.8	766	-25.	27.	2.5	9.3
02-01-83(0000)	02-28-83(2300)	672	-38.	34.	1.7	11.9	672	-19.	22.	2.2	7.4
03-01-83(0000)	03-31-83(2300)	222	-37.	37.	-4.5	15.8	222	-24.	36.	0.3	9.1
04-01-83(0000)	04-30-83(2300)	742					742				
05-01-83(0000)	05-31-83(2300)	653	-62.	20.	-10.3	13.4	653	-29.	24.	-0.6	9.2
06-01-83(0000)	06-30-83(2300)	742	-34.	32.	-2.0	11.1	742	-27.	18.	-1.0	7.7
07-01-83(0000)	07-31-83(2300)	766	-31.	28.	-1.2	9.5	766	-46.	20.	0.0	10.6
08-01-83(0000)	08-31-83(2300)	766	-36.	26.	-0.8	9.9	766	-26.	22.	-0.2	7.6
09-01-83(0000)	09-30-83(2300)	742	-31.	22.	-0.4	8.6	742	-22.	29.	0.6	6.8
10-01-83(0000)	10-31-83(2300)	766	-26.	22.	-0.4	9.5	766	-21.	19.	1.1	7.1
11-01-83(0000)	11-30-83(2300)	742	-39.	18.	-5.8	10.9	742	-16.	28.	2.9	6.5

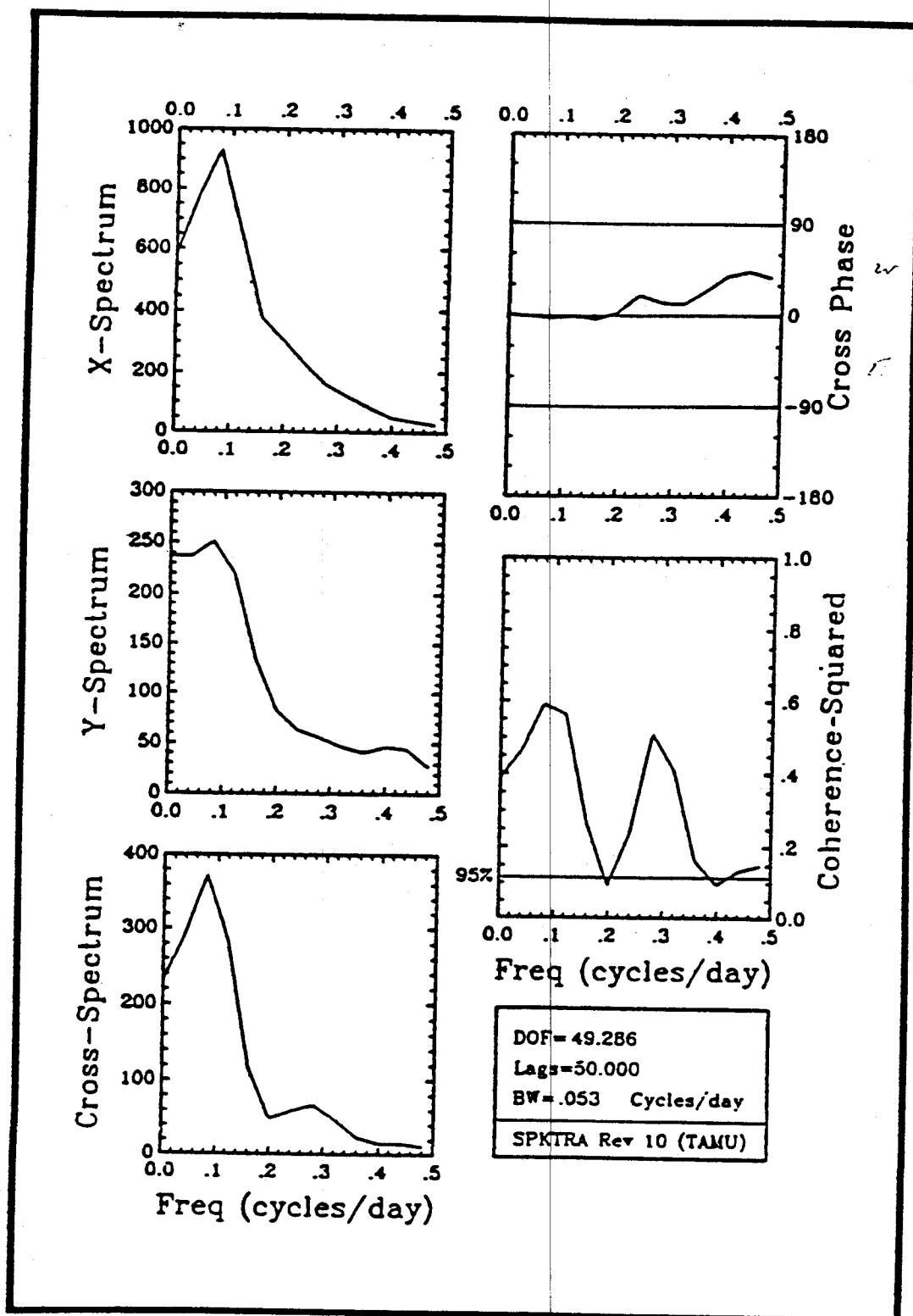


Figure 4.35. Autospectra, cross-spectrum, cross-phase and coherence squared for Bryan Mound site C (X) versus West Hackberry site D (Y) alongshelf components of near-bottom current for the 231-day period from 6 June 1981 through 22 January 1982. Ordinate units are $\text{cm}^2/\text{s}^2/\text{cycle/day}$. Negative phase means Y lags X (From: Kelly et al., 1983b).

investigated alongshore coherence from a 43-day period in the summer of 1977 at three sites along the central and lower Texas coast. Mid-depth current data in about 20 m of water exhibited an alternating alongshore motion that was coherent with the alongshore component of wind stress and had a net north-northeasterly displacement. Alongshore currents were highly coherent at the longest period resolved (12.5 days), and very nearly in phase over spatial scales in excess of 100 km.

Circulation East of Cameron

East of Cameron, LA there is much less combined wind and current meter data. Most of the observational programs have focused on the bays and estuaries and on the Mississippi Bight. The circulation in the bight is dominated by an anticyclonic drift of the plume from Southwest Pass that appears to be present approximately 70% of the time (Wiseman et al., 1976; Rouse and Coleman, 1976). Daddio (1977) found poor correlation between alongshore wind stress and current in the bight.

East of 92.5°W, the area of shallow depths increases markedly. According to Chuang and Wiseman, a large increase in bottom friction is to be expected in such shallow regions. As the coefficient of friction increases, the vertical-mean current required to balance a given alongshore stress decreases. Thus, even a constant alongshore stress at the coast may produce alongshore divergence in the current. Cochrane and Kelly (1986b) found that flow was directed onshore in this region as a result of divergence in the alongshore current because of a change in bottom frictional resistance, as well as wind stress divergence.

Bottom Ekman Veering

In a fluid with friction, the influence of the earth's rotation appears in the change of the direction of flow with depth, a phenomenon called Ekman veering. That Ekman veering is usually present can be seen in progressive vector diagrams of currents at various depths plotted from a common origin. Figures 4.36 and 4.37 (Kelly et al., 1984a,b) show progressive diagrams of currents at the Bryan Mound and the West Hackberry sites during the period 1 September through 30 November 1982. The various depths are indicated in the figures. From CM down to CZ there is counterclockwise veering in the bottom Ekman layer. In the bottom Ekman layer, the veering is counterclockwise, or to the left looking down, because the stress of friction is directed oppositely to the direction of flow in the wall layer just above the bottom. The mean veering angle at the Bryan Mound site is 52°. At the shallower West Hackberry site the surface and bottom Ekman layers are partially merged, but the counterclockwise veering is still evident and has a mean value of 20°.

Smith (1979) described the vertical structure of currents in 33 m of water off Aransas Pass based on current measurements at four levels above the bottom (2, 7, 16, and 23 m). With increasing distance above the bottom, both the increase in current speed and the clockwise angle (looking down) between the resultant current vector direction and the local alongshore direction were nearly linear when plotted against the

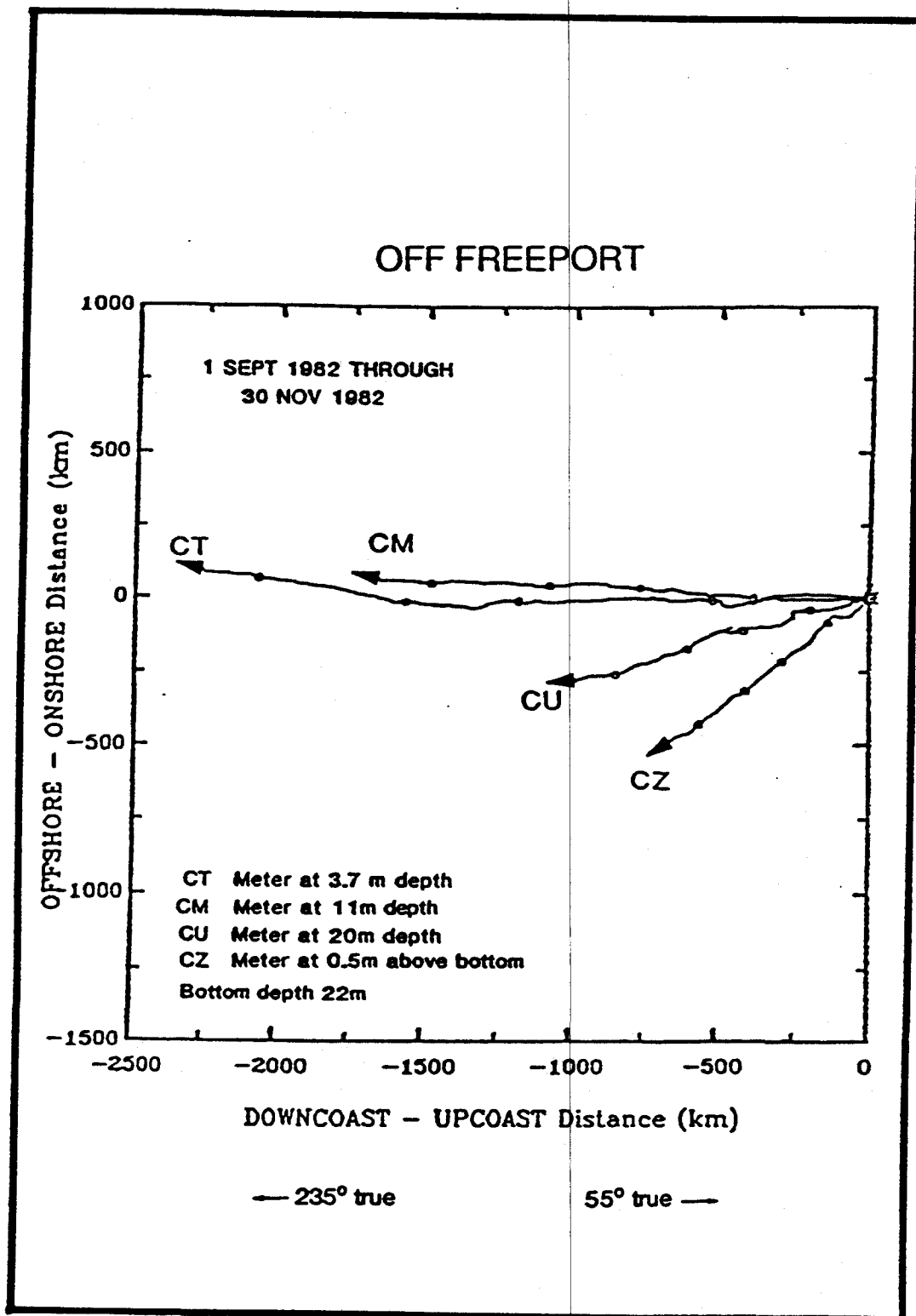


Figure 4.36. Progressive diagrams for currents observed during the period 1 September through 30 November 1982 at Bryan Mound site C at the depths indicated in legend (From: Kelly et al., 1984a). Tic marks are plotted at 20-day intervals.

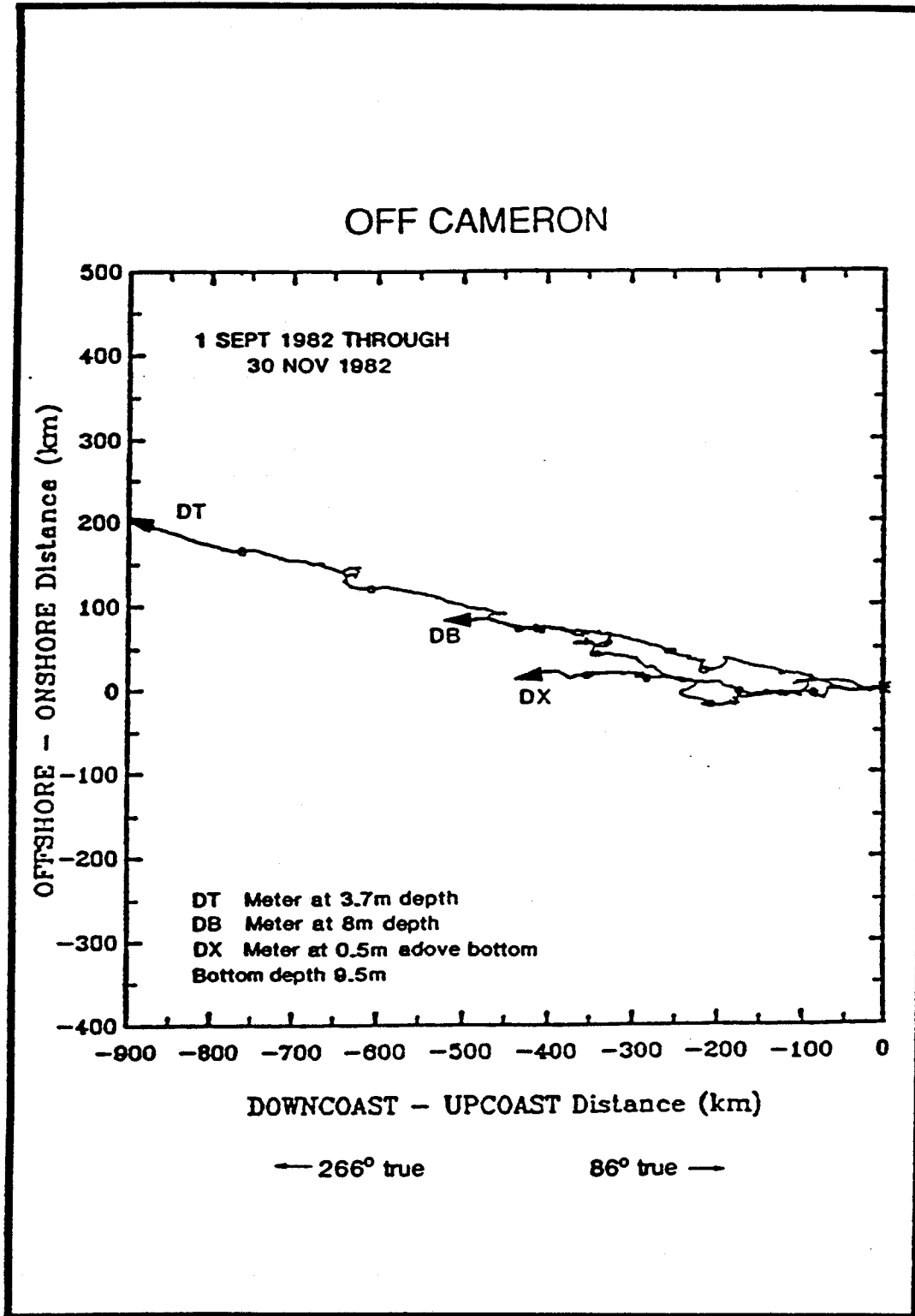


Figure 4.37. Progressive diagrams for currents observed during the period 1 September through 30 November 1982 at West Hackberry site D at the depths indicated in legend (From: Kelly et al., 1984b). Tic marks are plotted at 20-day intervals.

natural log of the height above the bottom, which would suggest a spiral pattern.

Kelly et al. (1985a) evaluated veering at the Bryan Mound site using six current meters between the surface and bottom in 22 m of water during the period 20 July to 8 August 1983. Meters were located at the following heights above the bottom: 0.5, 1.2, 2.1, 5.0, 11.0, and 18.1 m. There were two intervals of relatively steady current conditions, one with an upcoast wind component from 24 through 28 July and the other with a downcoast wind component from 30 July to 1 August. The mean current during the first time interval is shown as a function of depth in Figure 4.38a. The cross-shelf component is directed offshore (shown as positive) at the top current meter and onshore (negative) at all other meters, as might be expected for a wind with an upcoast component (upwelling). The curve is drawn so as to balance offshore and onshore flow. The alongshelf velocity component is quite strong; even at the bottom meter only 0.5 m above the bottom, it is more than 10 cm/s.

To reveal the turning of the current with depth, the mean velocities for the current meters at various depths are plotted in Figure 4.38b on a single polar coordinate grid. In both cases, the bottom two current meters (Z and W) show little turning, and, in the upcoast (July) case, even the third meter from the bottom (U) indicates only a very small turning. The layer with little turning seems to be the wall layer of more or less constant stress in which the velocity is expected to have a logarithmic profile. Above the quasi-logarithmic layer the current turns strongly clockwise with height. For the upcoast case, the turning from the direction at the deepest meter to the direction of the local isobaths (55° to 235°) is 39° . For the downcoast case the turning is 46° . The comparison with the local isobath orientation is made in order to avoid in so far as possible the Ekman veering due to wind stress. The current reaches this orientation somewhere above the current meter (M) at mid-depth. There appears to be no clear separation of the flow into surface and bottom Ekman layers with a geostrophic layer sandwiched between.

4.8.2 Outer Shelf

The prevailing currents in the upper layers on the outer shelf and continental slope flow toward the northeast or east, according to the orientation of the isobaths. The evidence for this comes from hydrographic data and direct measurements of current. Cochrane and Kelly (1986b) review various hydrographic data sets in which vertical sections crossing the outer shelf have isopycnals sloping down in the seaward direction. This implies eastward flow if the currents are quasi-geostrophic.

Direct measurements of currents in the upper part of the water column are available from the following locations along the outer shelf (Brooks et al., 1984): Green Canyon (GC), Flower Garden Banks (FG), East Breaks (EB), and Baker Bank (off Mustang Island) (MI) (see Section 4.8.3). Figures 4.39 through 4.42 display all available data for the respective locations in the form of a current rose and its associated

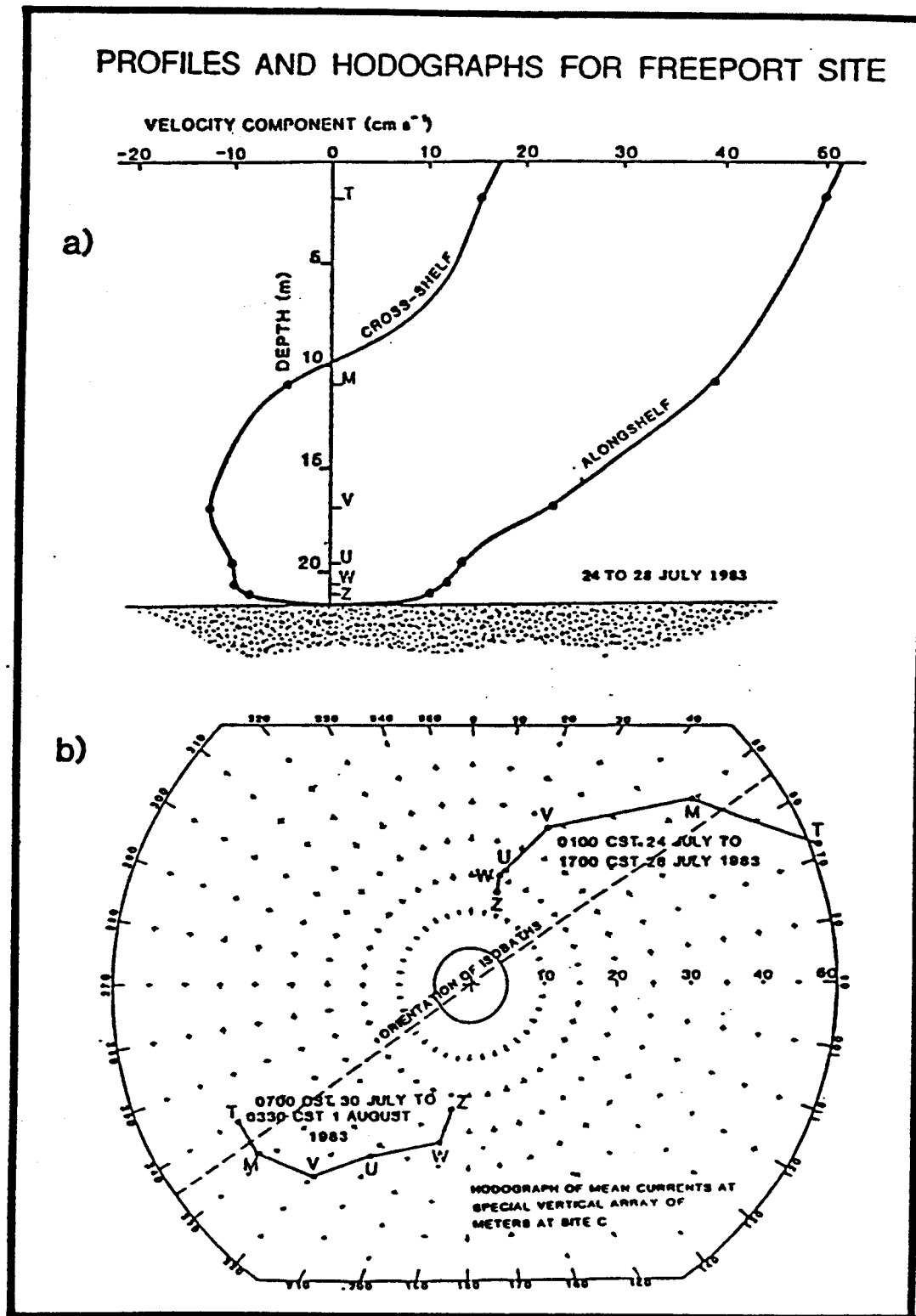


Figure 4.38. a) Mean current for the special vertical array of meters at Bryan Mound site C, 24 to 28 July 1983. Cross-shelf velocity is plotted as positive offshore and alongshelf velocity as positive upcoast (northeastward). b) Time-hodographs (connected points) of mean velocity for two periods. Each point is the head of a vector (starting at the origin) that represents the mean velocity at one of the meters (From: Kelly et al., 1985a).

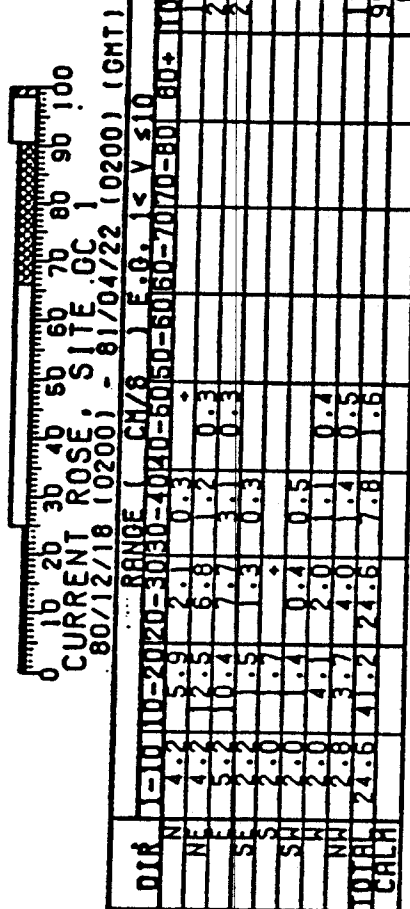
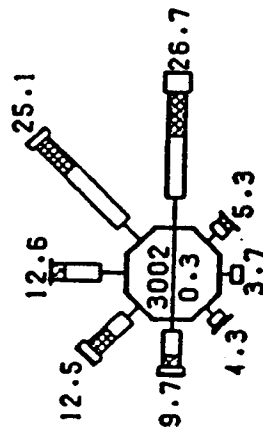
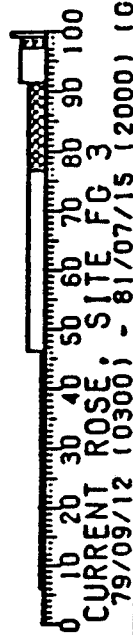
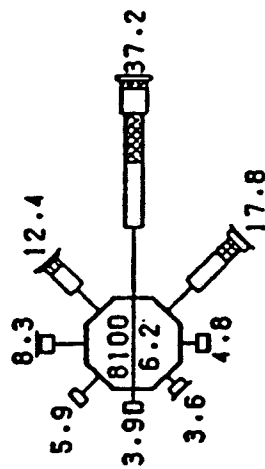
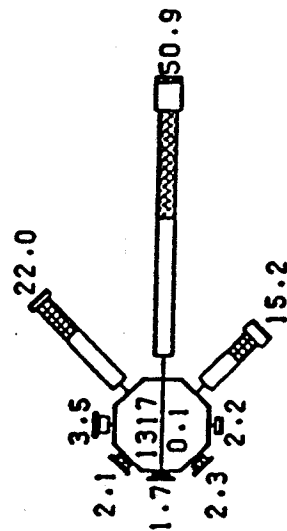


Figure 4.39. Current rose and associated table of joint frequency distribution of speed and direction from 3,002 hourly values of 3-h, low-pass-filtered time-series collected during the period 18 December 1980 to 22 April 1981 at 35 m depth in about 400 m of water in Green Canyon vicinity (From: Brooks et al., 1984). See Figure 4.1 for location.



DIR	1-10	10-20	20-30	30-40	40-50	50-60	60-70	70-80	80-90	90-100	Avg
N	5.4	2.1	0.7	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5
NE	5.5	4.6	1.7	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5
E	11.9	10.9	8.2	4.1	1.4	0.7	0.7	0.7	0.7	0.7	0.7
SE	6.0	6.9	3.3	1.1	0.4	0.4	0.4	0.4	0.4	0.4	0.4
S	2.8	6.8	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2
SW	1.9	1.4	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2
W	2.3	1.3	0.3	0.3	0.3	0.3	0.3	0.3	0.3	0.3	0.3
NW	4.2	1.5	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2
TOTAL	40.1	30.4	14.8	5.7	1.9	0.9	0.9	0.9	0.9	0.9	0.9
CALC											6.2

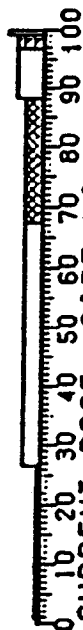
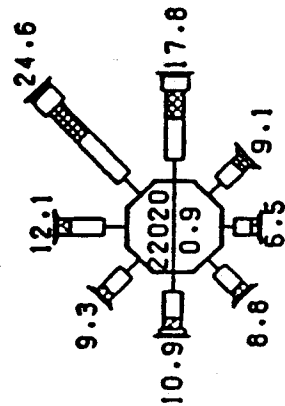
Figure 4.40. Current rose and associated table of joint frequency distribution of speed and direction from 8,100 hourly values of 3-h, low-pass-filtered time-series collected during the period 12 September 1979 to 15 July 1981 at 55 m depth in about 100 m of water just northwest of the West Flower Garden Bank (From: Brooks et al., 1984; data courtesy of D.W. McGlail). See Figure 4.1 for location.



0 10 20 30 40 50 60 70 80 90 100
CURRENT ROSE, SITE EB
80/06/19 (0400) - 80/08/12 (2300) (GMT)

DIR	RANGE (CH/S)				E.O. 1 < V ≤ 10				AVG
	1-10	10-20	20-30	30-40	40-50	50-60	60-70	70-80	
N	0.1	0.4	0.4	0.6	0.6	0.6	0.6	0.6	0.6
NE	2.2	11.2	7.4	0.8	0.5				7.2
E	3.9	22.6	18.7	4.6	1.1				22.0
SE	2.2	6.5	4.3	1.7	0.5				50.9
S	1.1	0.8	0.3						15.2
SH	0.7	0.2	0.9	0.3	0.2				2.3
H	0.5	0.2	0.5	0.2	0.2				2.3
NH	0.5	0.4	0.7	0.4	0.2				1.7
TOTAL	12.0	43.4	33.1	8.6	2.8				99.9
CALC									19.8

Figure 4.41. Current rose and associated table of joint frequency distribution of speed and direction from 1,317 hourly values of 3-h, low-pass-filtered time-series collected during the period 19 June 1980 to 12 August 1980 at 63 m depth in about 280 m of water in East Breaks vicinity (From: Brooks et al., 1984). See Figure 4.1 for location.



CURRENT ROSE, SITE MI
78/03/02 (0500) - 81/03/20 (0700) (GMT)

DIR	1-10	10-20	20-30	30-40	40-50	50-60	60-70	70-80	80-90	90-100	TOTAL	AVG
N	3.8	5.3	2.2	0.4	0.9	0.2	0.2	0.2	0.2	0.2	12.1	4.6
NE	4.1	9.6	6.7	3.0	0.9	0.2	0.2	0.2	0.2	0.2	24.6	20.1
E	3.6	6.7	4.5	2.3	0.5	0.2	0.2	0.2	0.2	0.2	17.8	19.4
SE	2.9	3.7	1.6	0.5	0.2	0.1	0.1	0.1	0.1	0.1	9.1	5.9
S	2.1	2.6	1.2	0.4	0.1	0.1	0.1	0.1	0.1	0.1	6.5	5.2
SW	2.7	4.1	1.3	0.5	0.2	0.2	0.2	0.2	0.2	0.2	8.8	5.3
W	3.3	4.5	1.9	0.8	0.3	0.3	0.3	0.3	0.3	0.3	10.9	6.8
NW	3.2	4.1	1.5	0.4	0.1	0.1	0.1	0.1	0.1	0.1	9.3	4.4
TOTAL	25.7	40.8	20.9	8.4	2.5	0.8	0.8	0.8	0.8	0.8	99.1	17.1
CRCH											0.9	

Figure 4.42. Current rose and associated table of joint frequency distribution of speed and direction from 22,020 hourly values of 3-h, low-pass-filtered time-series collected during the period 2 March 1978 to 20 March 1981 at 12 m depth in about 73 m of water in Baker Bank vicinity (off Mustang Island) (From: Brooks et al., 1984). See Figure 4.1 for location.

table of joint frequency distribution of speed and direction. The preponderance of flow to the east and northeast is obvious.

Cochrane and Kelly (1986b) argue that the flow along the edge of the outer shelf is a countercurrent or return flow to the downcoast current that dominates the inner shelf except during summer (see Section 4.8.3). The flows on the inner and outer parts of the shelf are connected by cross-shelf flows caused by convergence along the south Texas Coast and divergence in the east.

Near-bottom currents on the outer shelf are less well documented. They were not considered in the study of Brooks et al. (1984). Recently, near-bottom currents on the south Texas continental shelf were monitored during 145 days in the summer, fall, and early winter of 1984 (Science Applications International Corporation, 1986). Current meters were located in water depths of 12, 18, 34, 74, and 140 m along a line perpendicular to Corpus Christi Bay. At the lower frequencies, coherence between the overlying wind field and the bottom currents decreased in an offshore direction. However, there was still some coherence even at the shelf break. Net along-isobath displacements for July to October and October to December periods were downcoast at all locations. Net cross-isobath displacements suggested patterns of convergence and divergence across the shelf. Rezak et al. (1985) and Halper and McGrail (1988) found that at the Flower Garden Banks, near-bottom currents tend to flow toward the east or southeast, and that the stronger the flow, the more southerly it tends to be. They found an overall average of 1.6 cm/s toward the east-southeast for records in the 11 to 18 m above-the-bottom group. These results are not consistent with Ekman veering, which predicts counterclockwise (looking down) veering to the northeast for eastward flowing geostrophic currents above the Ekman bottom boundary layer. Rezak et al. (1985) offer several explanations: orographic effects from the banks, off-shelf flow of cold, dense water formed during frontal passages, and baroclinic shearing.

4.8.3 Prevailing Circulation on the Shelf

To provide a first approximation to the prevailing currents over the whole Texas-Louisiana shelf, Cochrane and Kelly (1986b) computed monthly mean geopotential anomalies (relative to the 70 db) from monthly mean temperatures and salinities for the data from the GUS III cruises. Their charts with streamlines are reproduced in Figure 4.43. In months other than June, July, and August, an elongated region of low geopotential dominates the shelf. On the inner-shelf side of the cyclonic feature, downcoast flow prevails, in agreement with the downcoast wind component discussed above. There is a counterflow (eastward or northeastward) along the shelf break. During the summer months, there is upcoast flow along much of the inner shelf to about 92.5°W, and a high in geopotential lies off the Louisiana coast. Although there is a suggestion of eastward or northeastward flow along the shelf break, it is very weak. The summer situation is, thus, very different from the situation during the rest of the year when there is a distinct "low" in geopotential over the outer shelf. The annual progression of the patterns of geopotential anomaly provide a conceptual framework that is consistent with the various aspects of shelf

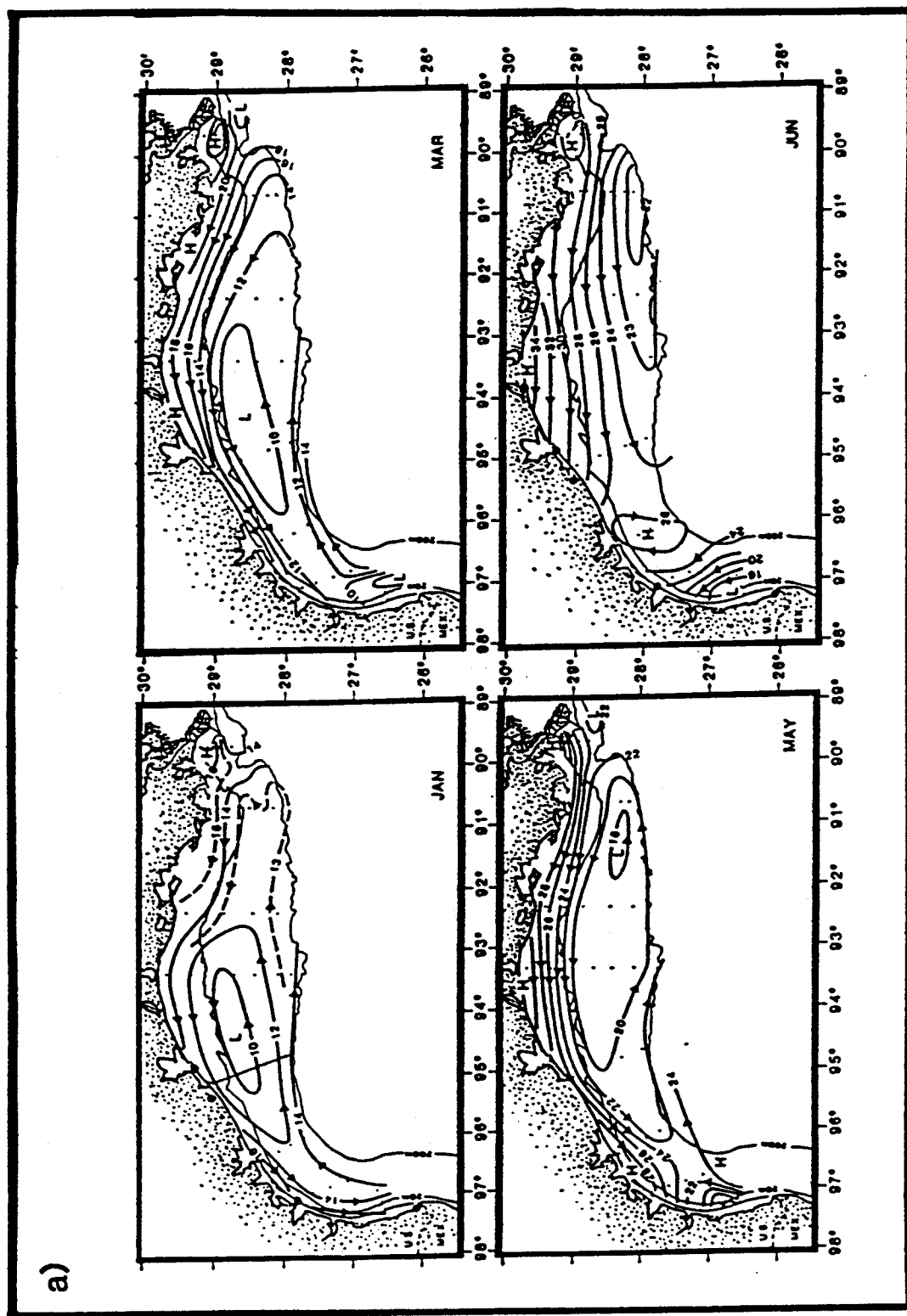


Figure 4.43. Monthly mean geopotential anomaly (dyn cm or $10^{-1} \text{ J kg}^{-1}$) of the sea surface relative to 70 db or 0.70 MPa based on data taken aboard M/V GUS III in 1963, 1964 and 1965 (From: Cochran and Kelly, 1986b). a) Data for January, March, May, and June. b) Data for July, August, September, and November.

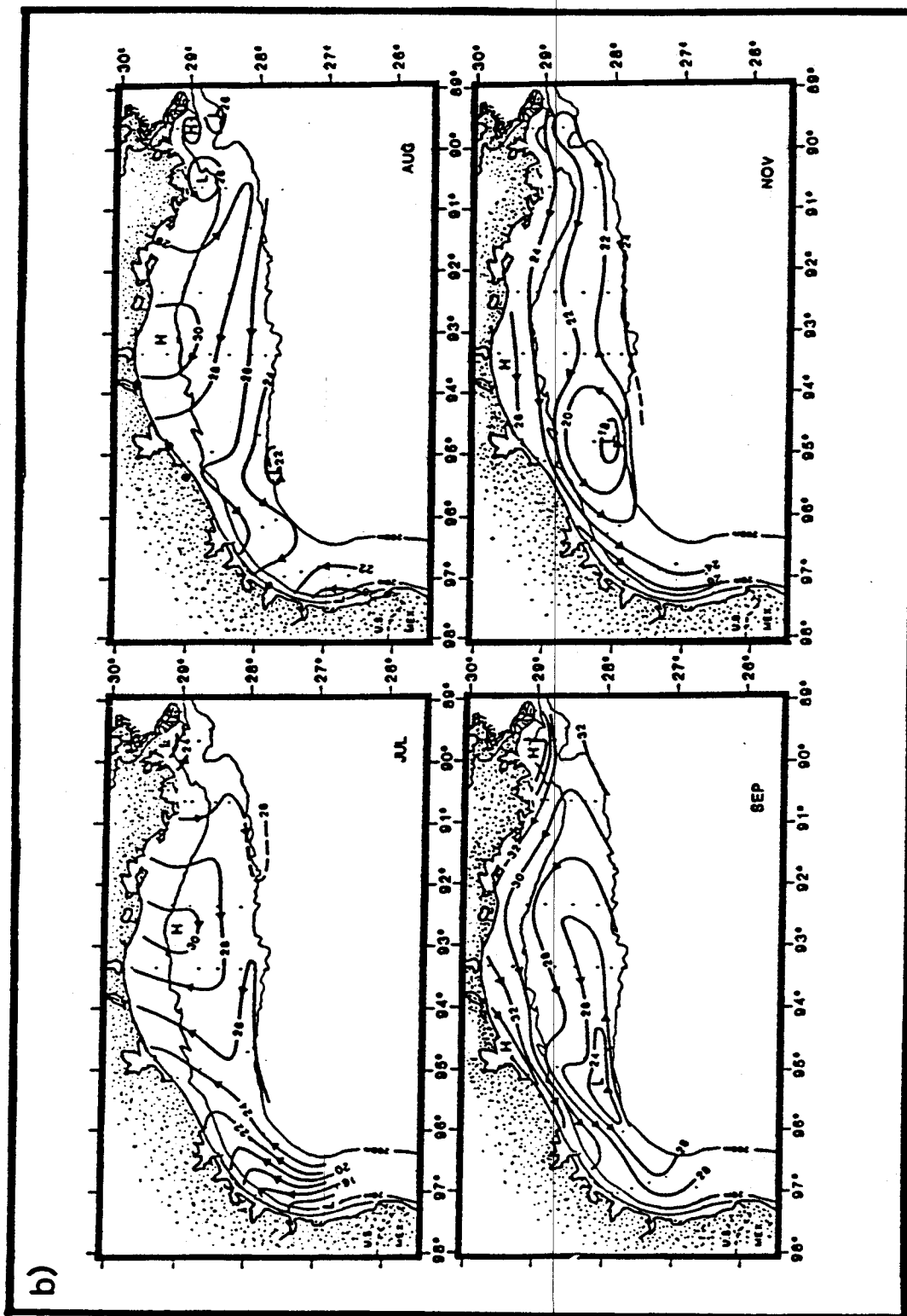


Figure 4.43. (Continued).

circulation and hydrography that have been described in the preceding sections.

4.8.4 Hurricane Currents

Currents generated by tropical storms and hurricanes have been observed at various locations on the shelf. Forristall et al. (1977) describe the effects of Tropical Storm Delia in September 1973 at the Buccaneer platform located in 20 m of water 50 km south of Galveston. Currents exceeded 2 m/s toward the west-southwest (approximately parallel to the local isobaths) at 17 m above the bottom. Peak currents at 3 m above the bottom were 75% of those at the upper level. During part of the time, currents were sufficiently large that the total particle velocities, including the oscillatory wave motion, never reversed direction, which means that steady currents were stronger than the orbital velocities of the highest waves. Forristall et al. (1978) describe these orbital velocities in detail and compare them with theoretically predicted velocities. Maximum observed horizontal wave velocities at 17 m above the bottom were on the order of 150 to 200 cm/s.

Tropical Storm Debra passed about 20 km west of the West Hackberry DOE/SPR site (where the water depth is 10 m) on 28 August 1978 (Frey et al., 1981). At 3 m above the bottom, maximum current speed reached 77 cm/s, and at 1 m above the bottom they were 57 cm/s. Hurricane Anita, during late August and early September 1977, moved west-southwest across the Gulf from its origin, about 250 km south of the Mississippi Delta, to landfall, about 250 km south of Brownsville. Despite the distance of Anita from the central Texas Coast, speeds of up to 80 cm/s were observed on the inner shelf (Smith, 1978b, 1980c; Maresca and Carlson, 1980).

At the Flower Garden Banks, Rezak et al. (1985) observed the effects of Tropical Storm Claudette at 60 m and 90 m in 100 m of water. They conclude that the passage of a hurricane over the banks elicits a three-layer response: a direct wind-driven current down to about the thermocline, an indirectly forced flow extending from the thermocline to about 10 m of the bottom, and a near-bottom flow, probably compensatory to a large degree to that at the surface. Strong inertial oscillations and 2-day shelf waves were also observed.

Hurricane Alicia passed directly over the Bryan Mound DOE/SPR site in 22 m of water (Kelly et al., 1984b). Current meters were moored at nominal depths, i.e., before mooring tilt caused by strong currents, of 3.6 m (CT), 10.9 m (CM), and 20.1 m (CU). Figure 4.44 shows the alongshelf component of currents at these depths after filtering with a 3-h low-pass filter. The effects of Hurricane Alicia were strong but short-lived. Peak currents were downcoast and offshore at all depths and occurred between 1800 and 2400 h CST on 17 August 1983 when the eye passed just east of the mooring. Peak values of the alongshelf component were 160 cm/s at CT, 125 cm/s at CM, and 100 cm/s at CU. Current speeds dropped to less than 25 cm/s at all depths by the end of 18 August and became quite weak for several days. Kelly et al. (1984b) also note that large scale homogenization of the coastal waters did not occur. Prior to Hurricane Alicia, on August 16, the water column at the mooring was very

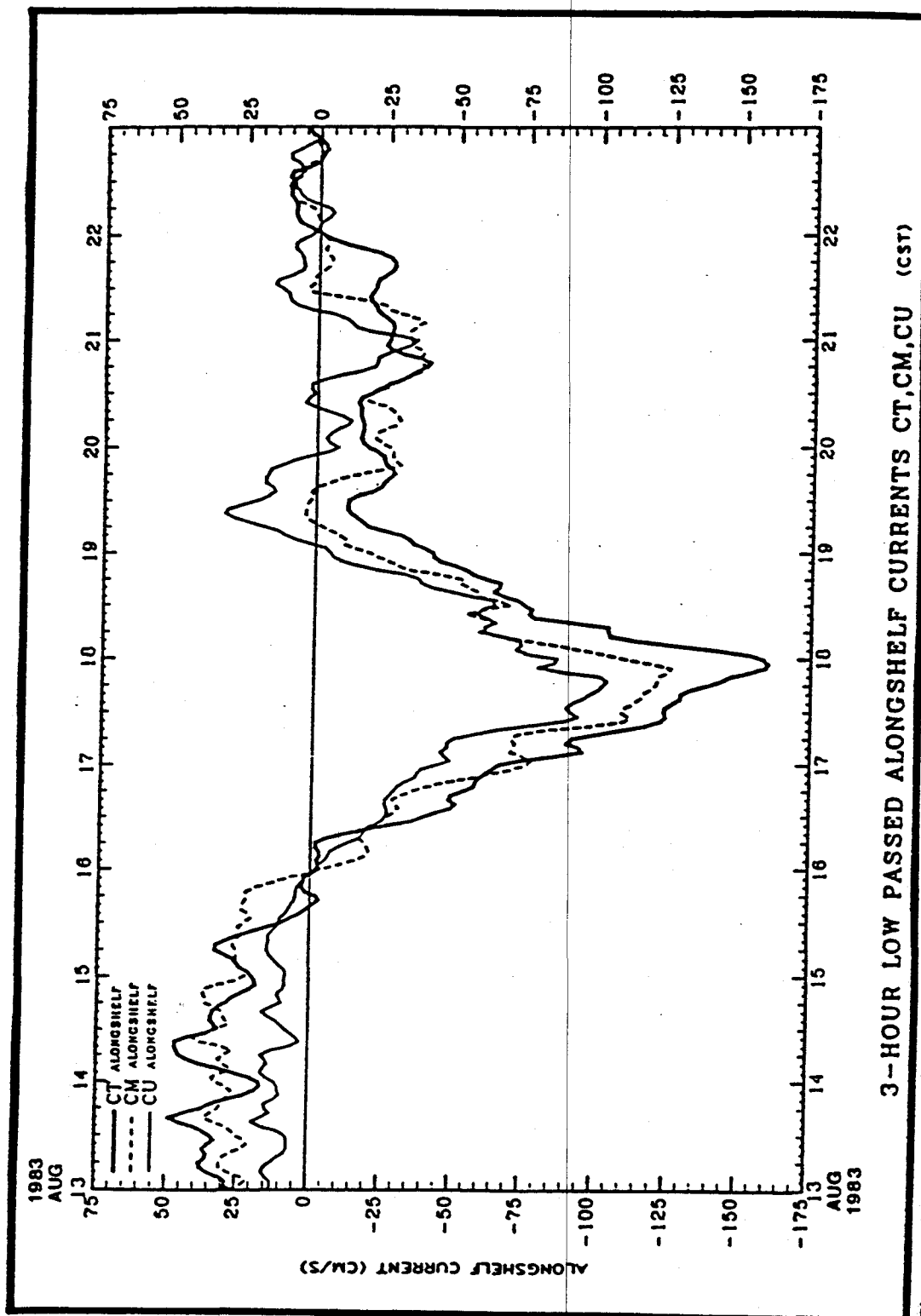


Figure 4.44. Alongshelf components of currents recorded as the eye of Hurricane Alicia passed a few kilometers to the east of a current meter mooring in 22 m of water at the Bryan Mound site. Current meter depths, before any mooring tilt caused by the strong currents, were 3.7 m (CT), 11 m (CM), and 20 m (CU). Original 2-min data were smoothed with a 3-h low-pass filter.

stratified. A thick layer of fresher, warmer water (30 to 32‰ and 29 to 30°C) overlay a thinner bottom layer of saltier, cooler water (35‰ and 25°C). The water column at the location of the mooring became homogeneous during the hurricane, but on 19 August salinity stratification had returned, with a top-to-bottom difference of 5‰. Advection from nearby areas, relatively unaffected by the hurricane, restored the stratification.

4.8.5 Eddy-Shelf Interaction

Large (diameters greater than 200 km), strong (swirl speeds greater than 50 cm/s) eddies, both anticyclonic and cyclonic, are frequently observed near the outer or deep part of the continental slope in the northwestern part of the Gulf of Mexico, but the degree to which they influence the circulation of the Texas-Louisiana shelf and the continental slope out to a depth of about 500 m is little known and poorly understood. Events of strong persistent currents on the upper slope off Louisiana and Texas that suggest the influence of eddies have been reported by offshore industry operations and recorded by current meters. Usually, however, observations of eddies lack concurrent current meter and sub-surface hydrographic observations on the shelf and slope, and vice versa, making it impossible to draw definitive conclusions. For example, Figure 4.45 shows a period of persistently eastward flow at the Flower Gardens during May 1980, when wind forcing at the time was weak and variable, the signature one might expect on the north side of an offshore anticyclonic eddy. Brooks and Legeckis (1982) and Brooks (1984) report that in April 1980 an anticyclone was centered near 25.5°N, 92°W, and its diameter was about 250 km. But if it moved due westward, its northern periphery would have extended only to the 500-m isobath, which is about 50 km south of the Flower Gardens, and so, a connection between the eddy and the currents in Figure 4.45 cannot be proved.

The anticyclonic eddies originate in the eastern Gulf where they separate from the Loop Current. Vukovich and Crissman (1986), in a study of 12 years of satellite infrared data, showed that, after separation, they follow three characteristic paths of westward migration across the western Gulf: a northern one, a mid-Gulf one, and a southwestern one. When an anticyclone follows the northern path its northern side brushes against the Texas-Louisiana slope, and the outer edge of its circulation frequently lies inshore of the 2000-m isobath. The studies during the MMS Gulf of Mexico Physical Oceanography Program: Year 3 (Science Applications International Corporation, 1988) observed an eddy that illustrates this case. Figure 4.46 shows the depth of the 8°C temperature surface, based on a mid-July 1985 expendable bathythermograph (XBT) survey, in an anticyclone that had just separated from the Loop Current. (The depth of the 8°C temperature surface indicates the pattern of the geostrophic flow; deep regions correspond to regions of anticyclonic circulation, and shallow regions to cyclonic circulation.) The eddy's subsequent movement was tracked by satellite infrared imagery, ARGOS drifters and ship surveys. Figure 4.47 shows that by mid-November 1985 the eddy was interacting with both the western and northern slopes and that several small scale cyclonic and anticyclonic features were located around its periphery. During December 1985 and January 1986

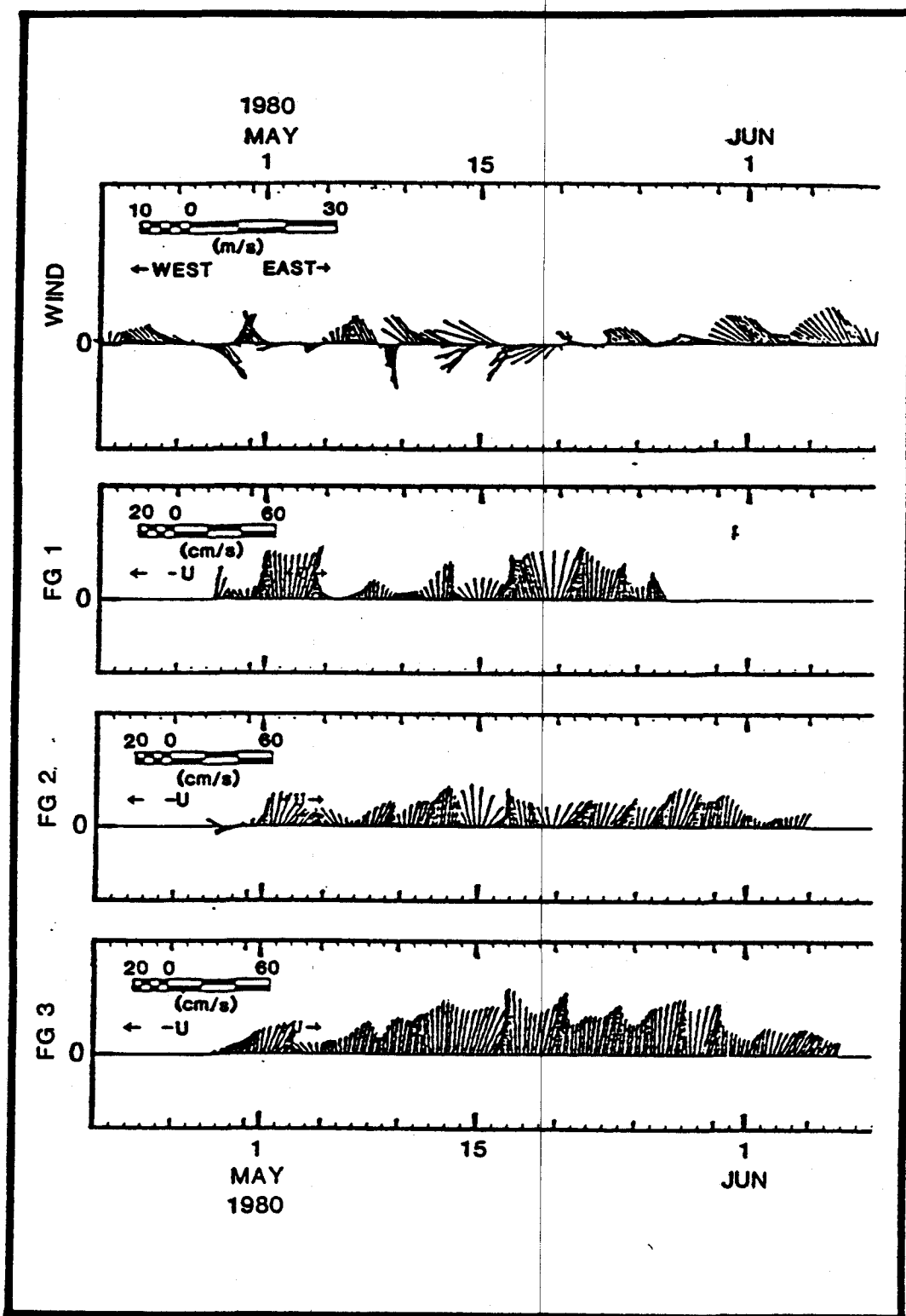


Figure 4.45. Comparative time-series plots of 40-h low-passed stick vectors for wind from NDBC Buoy 42008 ($28^{\circ}47.07'N$, $95^{\circ}18.71'W$) and currents from Flower Garden sites 1, 2, and 3 (instrument depths, about 50 m; water depths about 100 m), for May 1980. Vertically up is toward the east for the current stick vectors and towards the north for the wind stick vectors (From: Rezak et al., 1983).

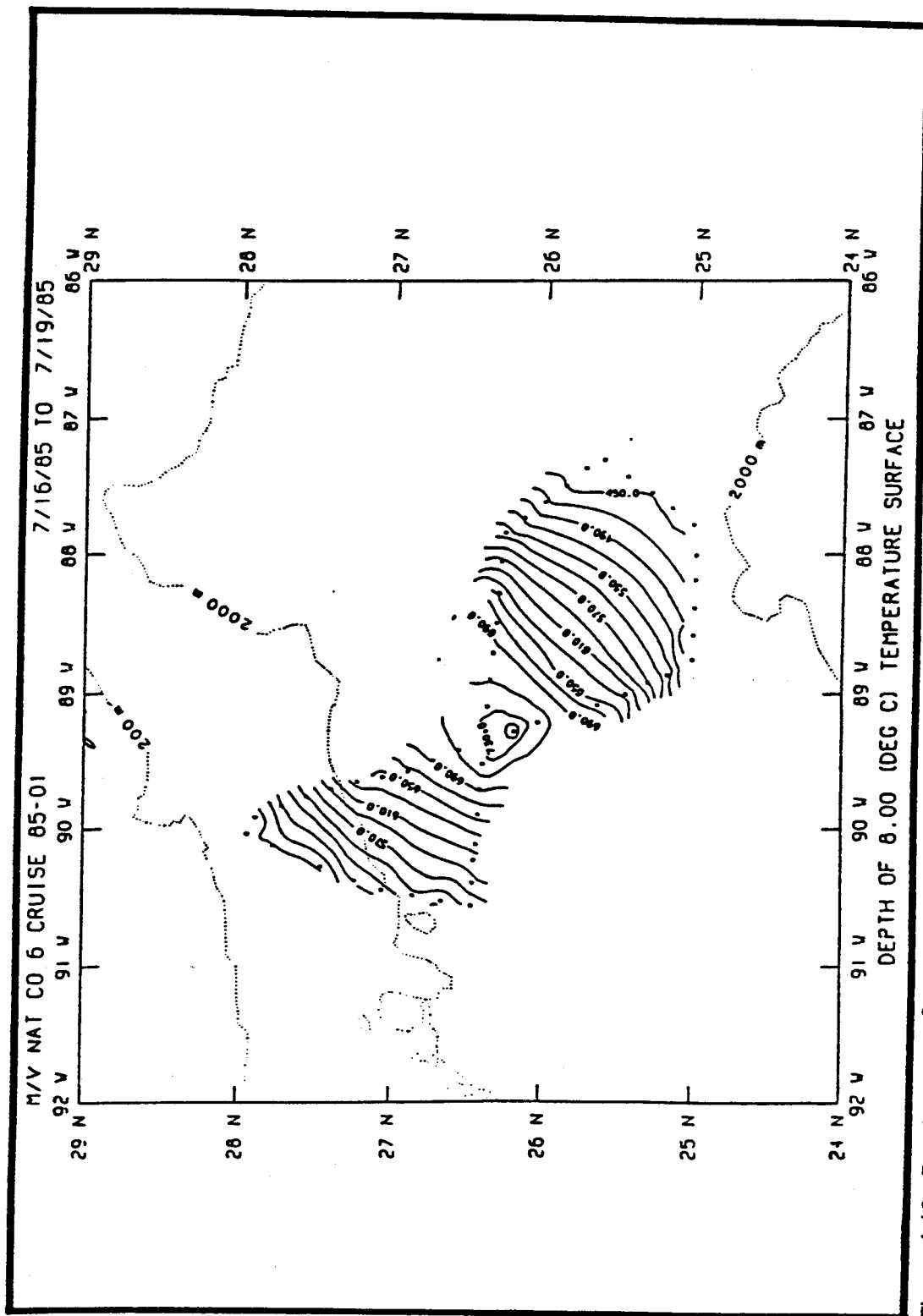


Figure 4.46. Depth of the 8°C temperature surface from XBT data in mid-July 1985 in an anticyclonic, warm-core eddy shed by the Loop Current in late June 1985 (From: Science Applications International Corporation, 1988).

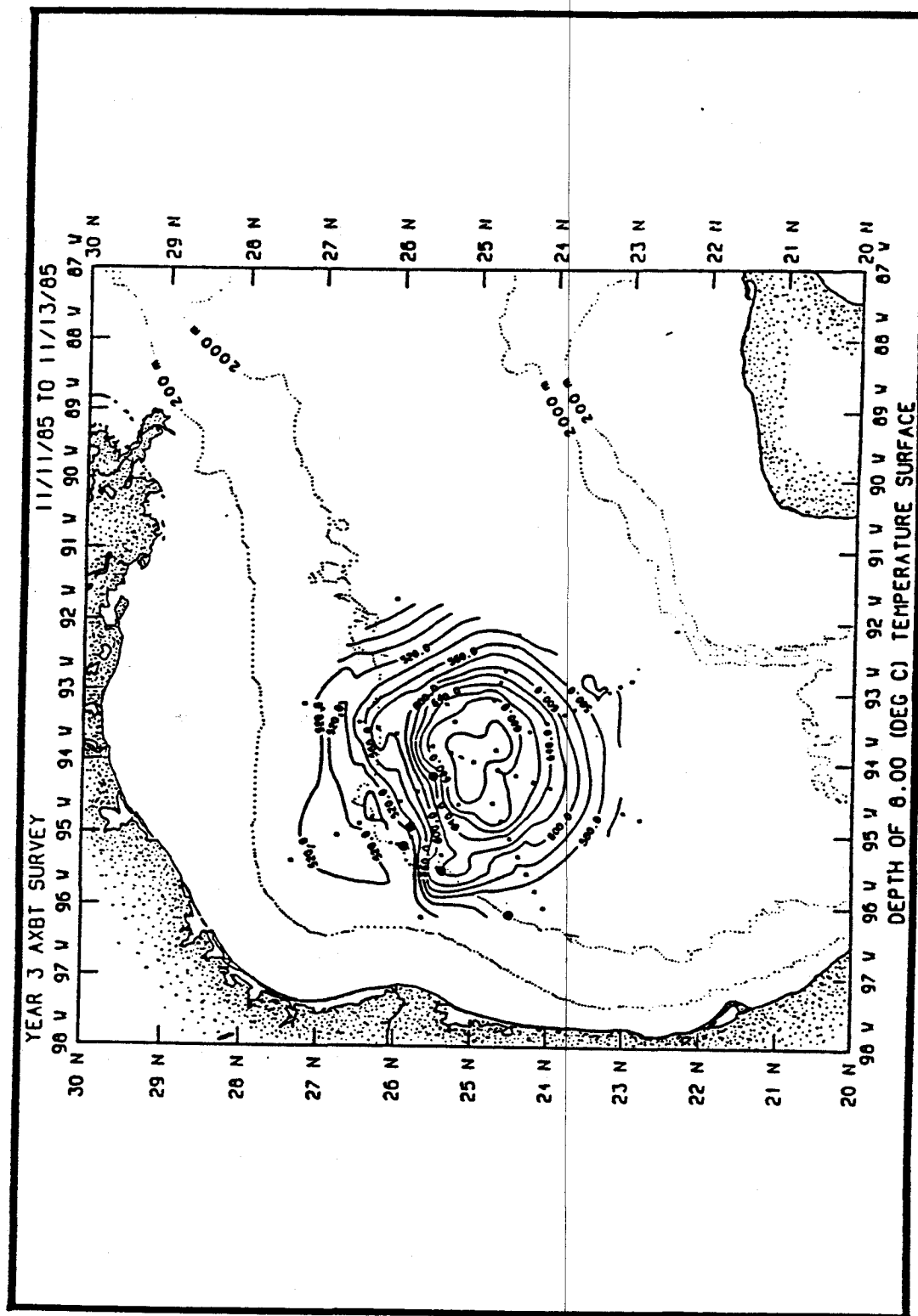


Figure 4.47. Depth of the 8°C temperature surface from AXBT data in mid-November 1985 in the same eddy as in Figure 4.46 (From: Science Applications International Corporation, 1988).

(Figure 4.48), the eddy moved to a position against the western continental slope, and its shape became quite elliptical.

The Texas-Louisiana slope is broad, and the available evidence does not resolve the question of whether an anticyclone's circulation can penetrate inshore of the 500-m isobath and onto the shelf, or instead, influences the shallow regions indirectly through flows associated with filaments, dispersion products, entrainment around the periphery, cross-slope flow between the anticyclone and leading and trailing cyclones, and other secondary mechanisms. Just south of about 26°N, the Rio Grande slope narrows to about half the width of the Texas-Louisiana slope. Here, as Figure 4.48 and the results of Elliott (1982), Merrell and Vazquez (1983), Brooks (1984) demonstrate, the perimeter of an anticyclone can press shoreward to the outer edge of the Mexican shelf, but it is not known if the circulation in the upper layer actually laps onto the shelf.

In the Western Gulf of Mexico, cyclonic eddies are frequently observed in association with Loop Current anticyclones (Elliott, 1979; Merrell and Morrison, 1981; Brooks and Legeckis, 1982; Merrell and Vazquez, 1983; Brooks, 1984). The formation of the cyclonic partner was observed for the first time during the Year 3 phase of the MMS Gulf of Mexico Physical Oceanography Program. Figures 4.47 and 4.48 show that between November 1985 and late January 1986, the small cyclonic feature on the northwest side of a Loop Current anticyclone greatly intensified as the anticyclone migrated onto the continental slope of the western margin. The cyclone's diameter increased to about 200 km and it lay entirely over the continental slope of southwest Texas between the 200- and 2000-m isobaths. The northeastward flow between the pair of vortices drew cool, low-salinity water off the Texas-Mexico continental shelf. Both this example and the studies cited above demonstrate that the vortex pair configuration drives a strong exchange between the western shelf and open Gulf waters. In Figures 4.47 and 4.48, one can also see a partially resolved cyclonic feature on the northeast side of the large anticyclone. Southward flow off the Texas-Louisiana shelf is suggested, but, since the slope is twice as wide as its western counterpart, the influence of this mechanism on the northern shelf may be much weaker.

To summarize, Loop Current eddies interact strongly with the base of the continental slope in the western Gulf of Mexico; they generate cyclonic eddies during topographic interaction; cyclonic eddies act in concert with Loop Current eddies to draw water on and off the shelf; other interactions between eddies and the continental shelf and upper slope are suggested by the existing data base but more information about the mechanisms through which eddies can influence the shelf and upper slope is needed.

4.8.6 Tidal Currents and Inertial Oscillations

These two classes of currents, although dynamically quite different, are grouped together because the period of inertial oscillation at the latitudes of the Texas-Louisiana shelf is very close to the periods of the diurnal tides, and it is difficult to separate

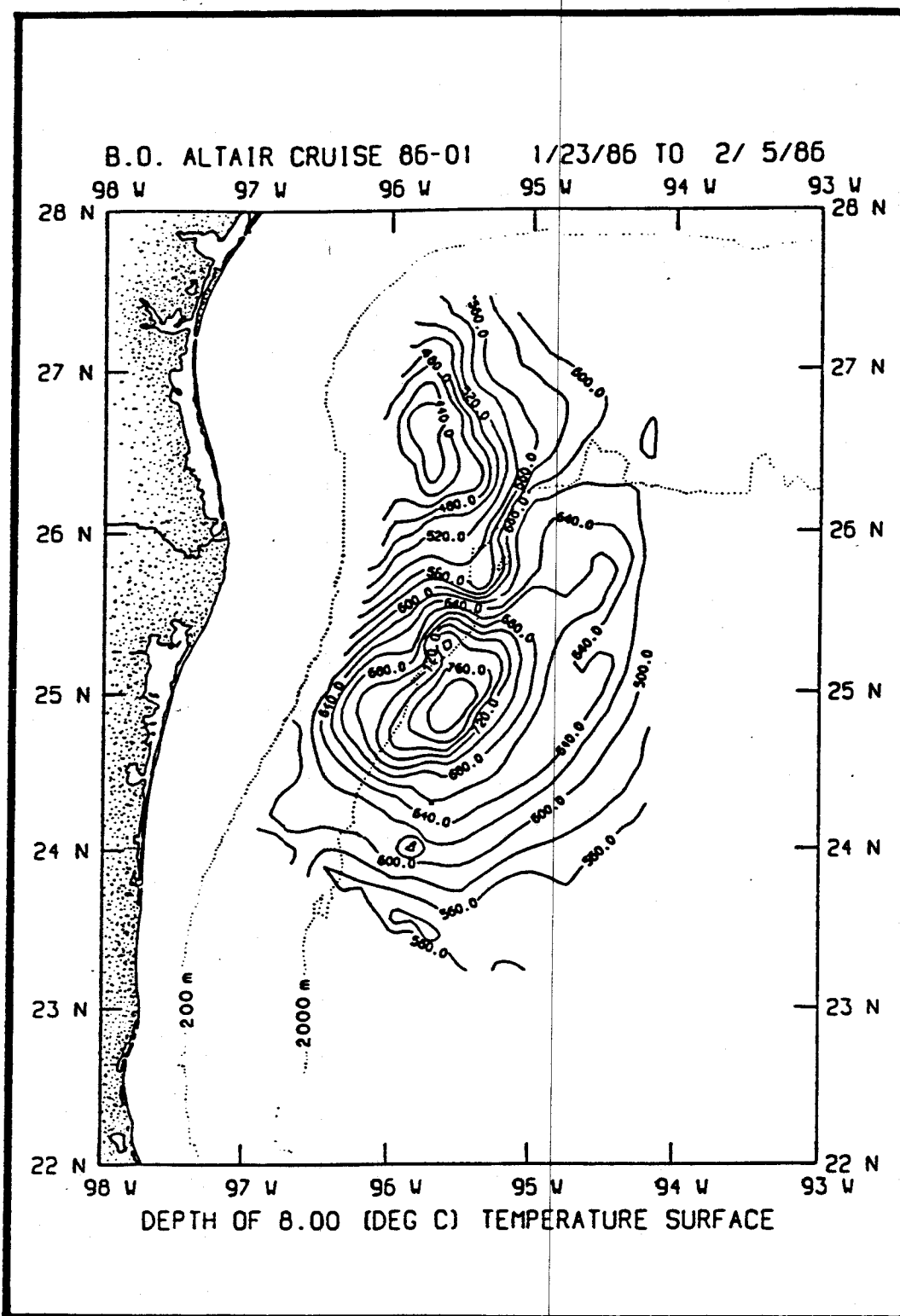


Figure 4.48. Depth of the 8°C temperature surface from XBT data in late January 1986 in the same eddy as in Figure 4.46 (From: Science Applications International Corporation, 1988).

them. Harmonic tidal analyses of current records from locations around the shelf routinely yield amplitudes of 1 to 4 cm/s for the K_1 , O_1 and M_2 constituents (Murray, 1976; Frey et al., 1981; Kelly et al., 1983a,b, 1984a,b, 1985a,b; Rezak et al., 1985). Tidal currents near the surface are often larger than those near the bottom, suggesting a baroclinic tidal component (Rezak et al., 1985). The eccentricity orientation and sense of rotation of tidal current ellipses varies by location, degree of stratification, season and depth. Although tidal currents are weak, they are relatively important in the bottom layer of the inner shelf when vertical stratification is strong and/or wind forcing is weak. At such times, tidal currents play an important role in mixing and advection (Kelly et al., 1983b).

Bursts of inertial oscillations lasting from 3 to 10 days or even longer appear in current meter records at most locations on the shelf (e.g., Kelly et al., 1983a,b; Brooks et al., 1984; Rezak et al., 1985). The maximum amplitude of inertial currents is on the order of 20 to 30 cm/s near the surface and 10 to 15 cm/s near the bottom. The oscillations occur most frequently in the late spring and summer months when vertical stratification is strong. An example of inertial currents can be seen in Figure 4.33. Daddio et al. (1978) analyzed two month-long current records, one in February and one in May, from the inner shelf west of the Mississippi River Delta. They demonstrated that the vigorous clockwise motions observed in the near-surface mixed layer were locally wind-induced inertial motions and not diurnal tides. The clockwise current oscillations with diurnal-inertial frequency were associated with strong frontal passages and uncorrelated with the temporal variations of tidal energy. A simple wind-driven model of the mixed layer satisfactorily reproduced the observed oscillations.

4.8.7 Shelf Waves

Low-frequency wave motions, commonly called continental shelf waves or topographic Rossby waves, can exist as a consequence of vorticity dynamics and can propagate current patterns and fluctuations along steep topographic gradients such as the edge of the continental shelf. These waves depend on the topographic gradients, not gravity, for their existence, so their most visible manifestations are horizontal current patterns. They propagate in a direction such that shallower water is on the right (northern hemisphere), have periods of a few days to weeks, wave lengths of hundreds of kilometers, and amplitudes of 5 to 30 cm/s or more. Shelf waves are often excited by atmospheric forcing on synoptic weather-related time scales, which in this shelf region are several days to a week (Table 4.1). Because of the long wavelengths of shelf waves and the distributed effects of wind forcing, it is difficult to separate the free, propagating waves from the locally forced response at a given location. Evidence of phase propagation is often inconclusive or confusing. On the Texas shelf, the phases tend to be further randomized by the irregular topography of the shelf edge, especially near rough areas such as the Flower Garden Banks.

Observations of shelf waves on the Texas-Louisiana shelf are limited. The strong coherence at 3- and 10-day periods between the bottom currents at Bryan Mound and West Hackberry sites along the coast

(Figure 4.35) may be related to shelf waves. Rezak et al. (1985) present evidence from current records at the Flower Garden Banks that suggest shelf waves with periods of 33 h, 2 days and 4 days. The 2-day wave has a wavelength of about 80 km and amplitudes as large as 15 cm/s but normally on the order of 4 cm/s. They also note that analyses of satellite thermal images show wavelike perturbations (onshore-offshore) in sea-surface isotherms at the shelf edge that have a wavelength of about 83 km, which is consistent with the wavelength they estimated for a shelf wave with a 2-day period. Brooks et al. (1984), by means of spectral coherence studies of currents at locations near the shelf edge, also find evidence of westward propagating shelf waves with a period of 2 to 3 days.

4.9 DATA GAPS AND INFORMATION NEEDS

The variability about the mean pattern of surface circulation developed by Cochrane and Kelly (1986b) is largely unknown except in the regions of the DOE/SPR studies. An instantaneous or synoptic picture of the surface currents on the shelf might exhibit several cells of circulation in the central region, rather than one large cyclonic gyre, and complex patterns near the west and east ends of the shelf. Multiple cells of circulation would imply multiple regions in which there is flow across the width of the shelf.

The spatial pattern of mean bottom circulation over the shelf, and its variability, needs to be described and related to the surface circulation and hydrographic variability.

Information about the exchange processes between the shelf and the oceanic Gulf regions is very limited. The influence of Loop Current eddies, the export of water masses created by cold-air outbreaks, the flushing of fresh water discharge on to the shelf, upwelling of deep ocean water masses onto the shelf, and transport of sediment off the shelf are important yet relatively unexplored topics for this shelf.

Our understanding of the effects and fate of the fresh water from the Mississippi-Atchafalaya River System is based largely on the salinity data from the GUS III cruises in the years 1963 to 1965, when river discharge was lower than average. The distribution of salinity on the shelf should be mapped with greater spatial resolution, both horizontally and vertically, during a year with above normal volumes of river discharge.

The structure and the dynamic behavior of the intense frontal zones on this shelf should be studied through a combination of observation and modeling, with a view towards defining the effect that stratification, both horizontal and vertical, has on the mixing, distribution and transport of nutrients, plankton and pollutants.

The rate at which the concentration of dissolved oxygen changes locally in bottom waters when there is vertical stratification is unknown. All terms in the diffusion equation for dissolved oxygen need to be studied: local rate-of-change, advection, production, consumption, and turbulent mixing, both vertical and horizontal.

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As the Nation's principal conservation agency, the Department of the Interior has responsibility for most of our nationally owned public lands and natural resources. This includes fostering the wisest use of our land and water resources, protecting our fish and wildlife, preserving the environmental and cultural values of our national parks and historical places, and providing for the enjoyment of life through outdoor recreation. The Department assesses our energy and mineral resources and works to assure that their development is in the best interest of all our people. The Department also has a major responsibility for American Indian reservation communities and for people who live in Island Territories under U.S. Administration.

