SEDIMENTATION IN FLUVIAL-DELTAIC WETLANDS AND ESTUARINE AREAS, TEXAS GULF COAST

Literature Synthesis

by

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INTRODUCTION

Deltaic and associated alluvial areas at the mouths of rivers that discharge into the bay–estuary–lagoon system along the Texas coast are the sites of extensive salt-, brackish-, and fresh-water marshes that are essential components of these biologically productive estuarine systems. These bayhead depositional systems are constructed primarily by fluvial sediments, sediments transported and deposited by the major rivers that enter estuarine waters. The loss of over 10,000 acres of wetlands in alluvial and deltaic areas of the Neches (White and others, 1987) and San Jacinto Rivers (White and others, 1985) has emphasized the need to examine in more detail the processes that establish and maintain, as well as degrade, these important natural resources along the Texas coast.

Background and Scope of Study

This report, which is a synthesis of published and unpublished data that focuses on fluvial-deltaic and estuarine sedimentation and associated interactive processes, is part of a study funded by the Texas Parks and Wildlife Department and Texas Water Development Board with funds allocated by the Texas Legislature for comprehensive studies of the effects of freshwater inflows on the bays and estuaries of Texas.*

*In response to House Bill 2 (1985) and Senate Bill 683 (1987), as enacted by the Texas legislature, the Texas Parks and Wildlife Department and the Texas Water Development Board must maintain a continuous data collection and analytical study program on the effects of and needs for freshwater inflow to the State's bays and estuaries. As part of the mandated study program, this research project was funded through the Board's Water Research and Planning Fund, authorized under Texas Water Code Sections 15.402 and 16.058 (e), and administered by the Department under interagency cooperative contracts No. IAC (86-89)0821 and IAC(88-89)145.

Most of the Texas freshwater inflow studies, past and ongoing, have focused on inundation, cycling and exchange of nutrients, salinity patterns, and fisheries production (TDWR, 1982). A significant part of the past research effort has dealt with the need to inundate deltaic wetlands (through freshwater inflows) in order to export nutrients into the estuarine system. Although habitat maintenance was one of the objectives of the investigations, little emphasis was placed on the geological processes that play a critical role in the construction of the deltaic and alluvial systems on which the biologically productive wetlands develop.

Among the objectives of this study is to focus on the sedimentary and associated interactive processes that develop, maintain, and/or degrade the environments. Information is provided on the present and historical (including geologic) role of fluvial sediments—sediments carried by rivers—in developing and maintaining estuarine habitats, with emphasis on wetlands, marine grassflats, and benthic communities. Interactive processes that are presented include: subsidence (both natural and human-induced), sea-level rise, riverine discharge and associated sediment loads, fluvial-deltaic-wetland sedimentation, bay-estuary-lagoon sedimentation, and biodeposition.

TEXAS BAY–ESTUARY–LAGOON SYSTEMS

LeBlanc and Hodgson (1959) distinguish between estuaries and lagoons along the Texas coast by differences in origin and by differences in alignment with respect to the Gulf shoreline. Estuaries, which have formed as a result of valley entrenchment, are generally aligned perpendicular to the coast, while lagoons, which formed as a result of shoreline processes, are generally aligned parallel to the coast (Fig. 1). Texas coastal water bodies, although perhaps most properly termed coastal lagoons (Morton and McGowen, 1980), have been variously referred to as bays, estuaries, and lagoons in the literature. For simplification in this report, the terms "bays" and "estuaries" are used interchangeably.

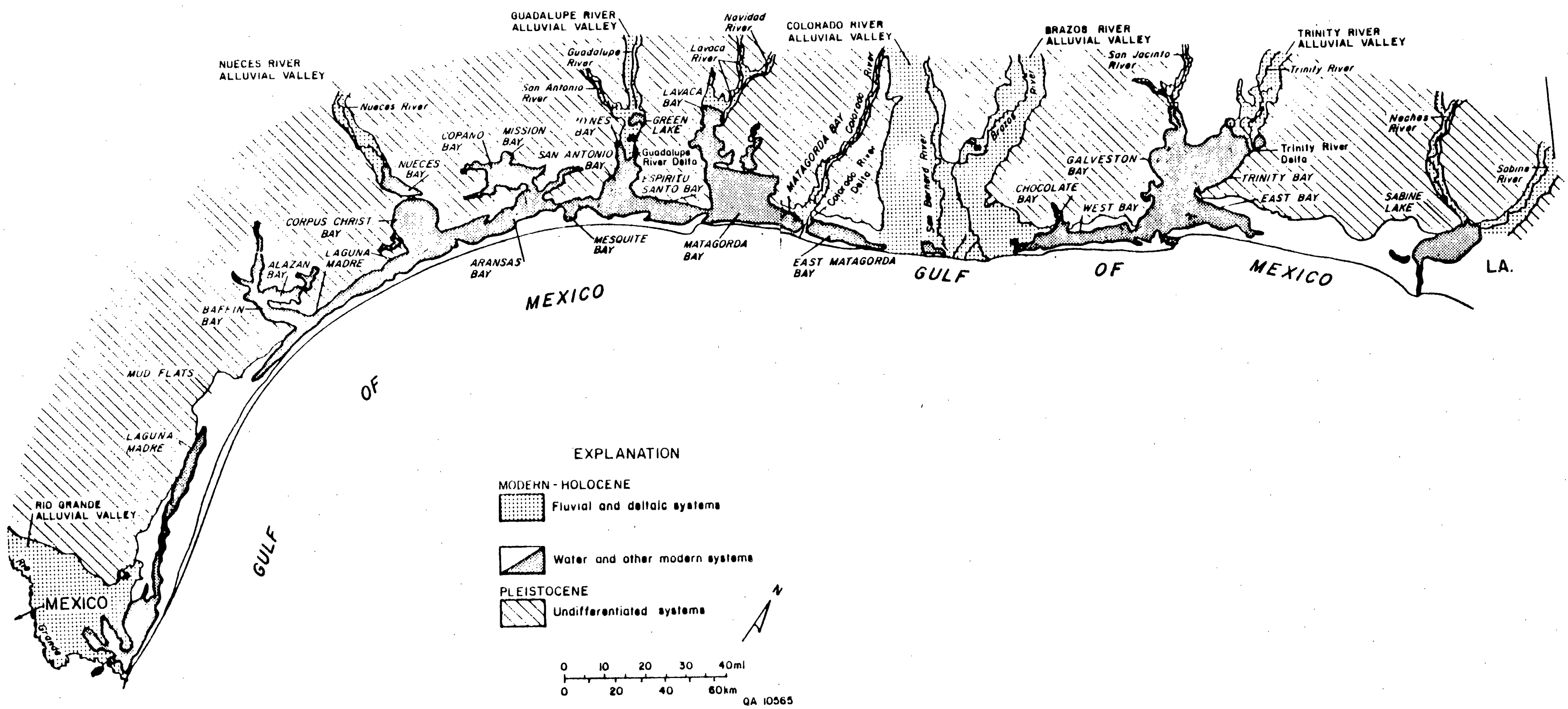


Figure 1. Bay-estuary-lagoon and major fluvial-deltaic systems along the Texas Gulf Coast. (Modified from LeBlanc and Hodgson, 1959.)

To properly understand the current processes that affect Texas estuarine environments, it is helpful to look briefly at past processes and events that have developed and helped shape the estuaries and associated deltaic areas along the Texas coast.

Origin of Texas Estuaries

Texas estuaries have evolved as deeply eroded valleys were flooded by a rising sea level (Price, 1933, 1947). Sea levels have fluctuated dramatically during the past million years as a result of alternating cooling and warming climatic cycles that have produced glacial and interglacial periods. During periods of glaciation, large amounts of water are locked up in continental ice sheets, resulting in dramatic drops in worldwide sea level (Fig. 2). Over the past 2.5 to 3 million yr, evidence shows that there have been at least eight major cycles in which sea level has fallen (producing a lowstand) and risen (highstand) (Beard and others, 1982). The last major cycle is pertinent to this discussion because it has had the most profound effect on our modern bay–estuary–lagoon system and Gulf shoreline.

During the most recent major period of glaciation and lowstand (ending about 18,000 yr ago), sea level along the Texas coast was approximately 120 m (400 ft) below today's level (LeBlanc and Hodgson, 1959; Curray, 1960); this placed the shoreline between 80 and 220 km (50 and 140 mi) offshore (LeBlanc and Hodgson, 1959). During the lowstand, the base levels of rivers along the Texas coast, as well as throughout the world, were lowered and extensive down-cutting and erosion formed deep valleys. The valleys cut by rivers along the Texas coast range from about 15 to 45 m (50 to 150 ft) deep (relative to today's sea level) along the valley axes near the gulfward margin of the bay shorelines (Shepard and Moore, 1955; Fisk, 1959; Behrens, 1963; McEwen, 1969; Byrne, 1975; Wright, 1980). As sea level rose, the valleys were flooded.

Approximately 4,500 yr before present (B.P.), the rise in sea level slowed as it approached today's level. During the rise in sea level, rivers meandered within their valleys, depositing large point-bar sand bodies and extensive overbank mud sheets (Fisher and others, 1972). The erosion of

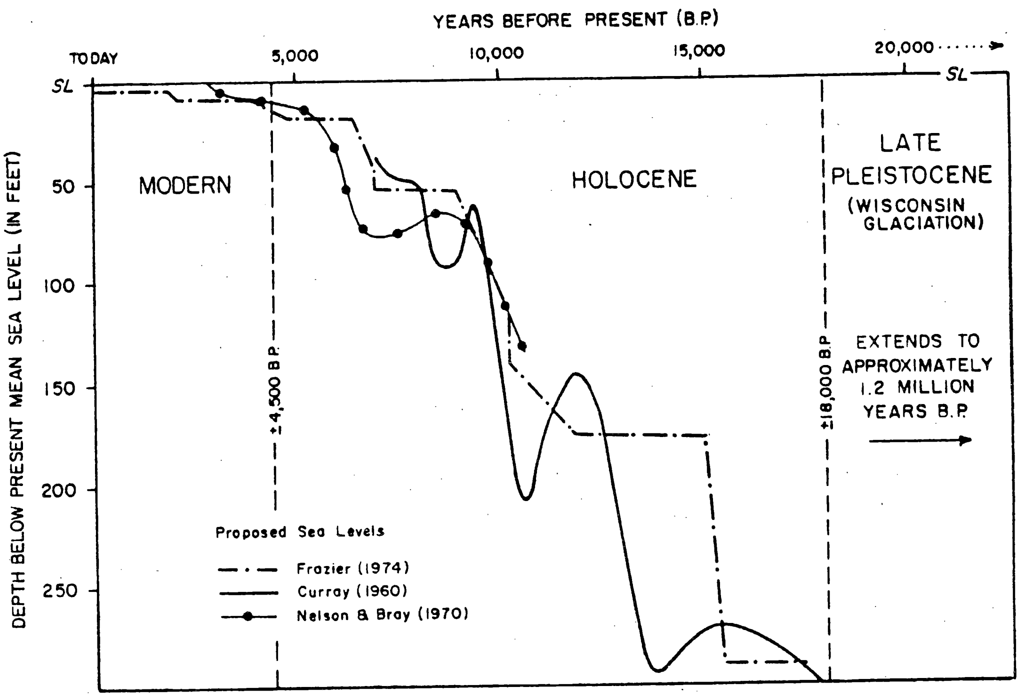


Figure 2. Sea-level changes during the past 18,000 yr, as interpreted by various authors. All curves show a general trend of rising sea level although the authors interpret minor sea-level fluctuations differently. A generalized curve would show sea level 4,500 yr ago to have been about 15 ft lower than at present. (From Fisher and others, 1973.)

the valleys and subsequent deposition was recorded by most streams, including the Trinity River (Fig. 3). Deposition of sediment in the entrenched valleys by rivers like the Trinity could not keep pace with sea-level rise, and much of the river valleys was drowned, producing estuaries like Trinity Bay. Trinity Bay and other Texas bays have since been enlarged by shoreline erosion, and the deeper parts of the submerged valleys have been filled slowly by bay sediment. The major sediment depositional centers in the estuaries are the bayhead deltas. These active depositional features have filled much of the lower stream valleys and have advanced (prograded) over bay muds at the heads of the bays. Along the gulfward reach of the estuaries, a series of barrier islands and peninsulas has formed, restricting the exchange of marine and estuarine waters to relatively narrow tidal inlets.

In contrast to rivers like the Trinity, three Texas rivers—Brazos, Colorado, and Rio Grande—filled their estuaries with sediments and constructed broad deltaic plains that protrude into the Gulf of Mexico (Price, 1947; LeBlanc and Hodgson, 1959). The fact that some rivers have filled their valleys and others have not is related to the nature of their drainage basins and sediment supply (Fisher and others, 1972). The depositional patterns developed by the Brazos and Colorado Rivers indicate that, in the past, they merged to form a single alluvial system that rapidly filled their valleys and estuaries (LeBlanc and Hodgson, 1959; McGowen and others, 1976a) (Fig. 1). The Colorado River has more recently abandoned its ancestral courses (which included Caney Creek) and now flows in a more narrow alluvial valley that intersects Matagorda Bay southwest of its previous course near the Brazos (LeBlanc and Hodgson, 1959; Wilkinson and Basse, 1978) (Fig. 4). The Brazos River discharges into the Gulf near Freeport and has built a small delta at its mouth. The evolution of the Texas coastal shoreline is depicted in Figure 5.

General Setting

The Texas Gulf shoreline is nearly 595 km (370 mi) long and consists of seven major estuarine systems (Fig. 1). The total open-water surface area of the estuaries at mean high water (MHW) is 620,634 hectares (1,532,430 acres) (Diener, 1975). Many of the bays are comparatively small,

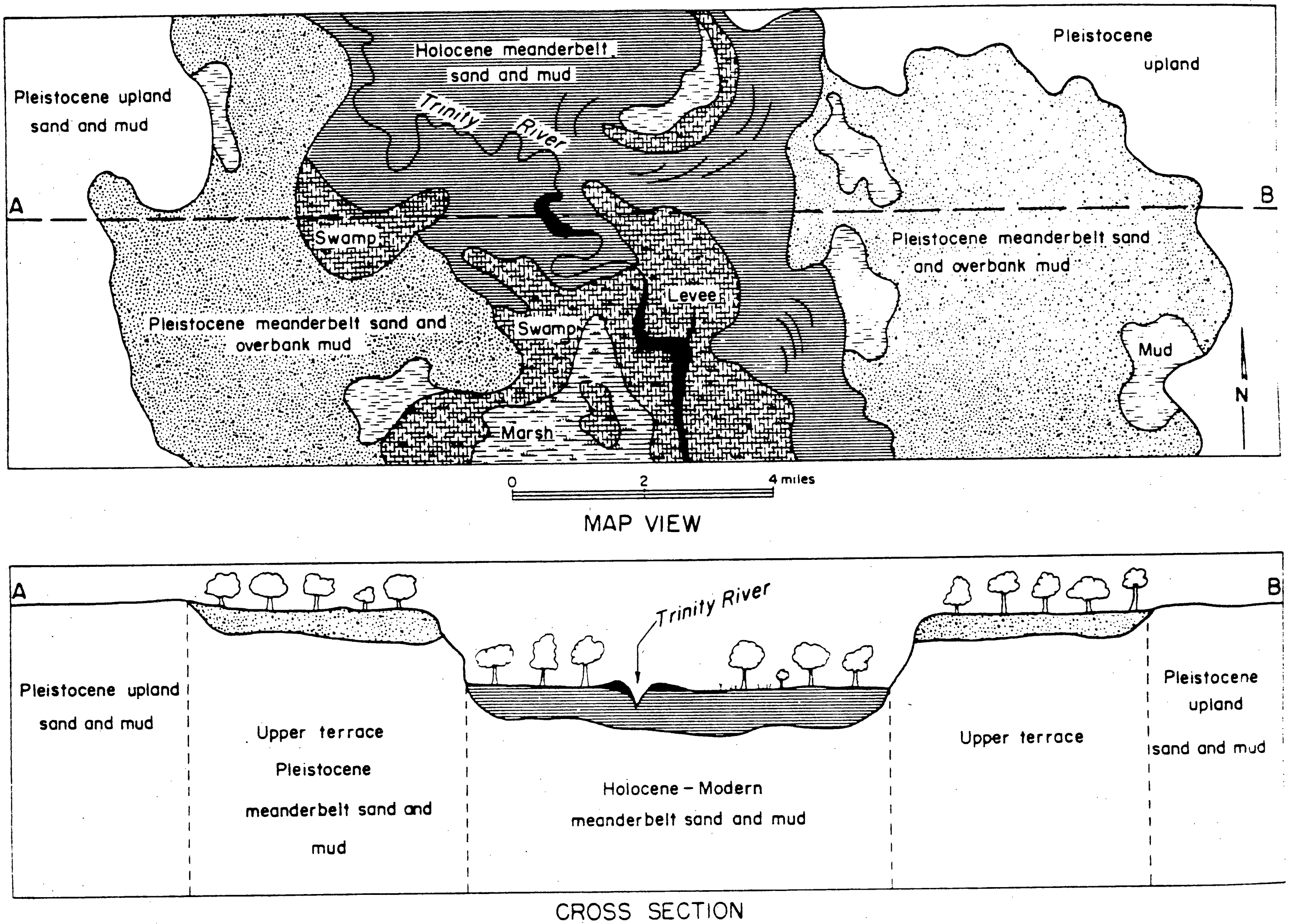


Figure 3. Fluvial environments and facies deposited along the incised lower Trinity River valley, Galveston–Houston area. (From Fisher and others, 1972.)

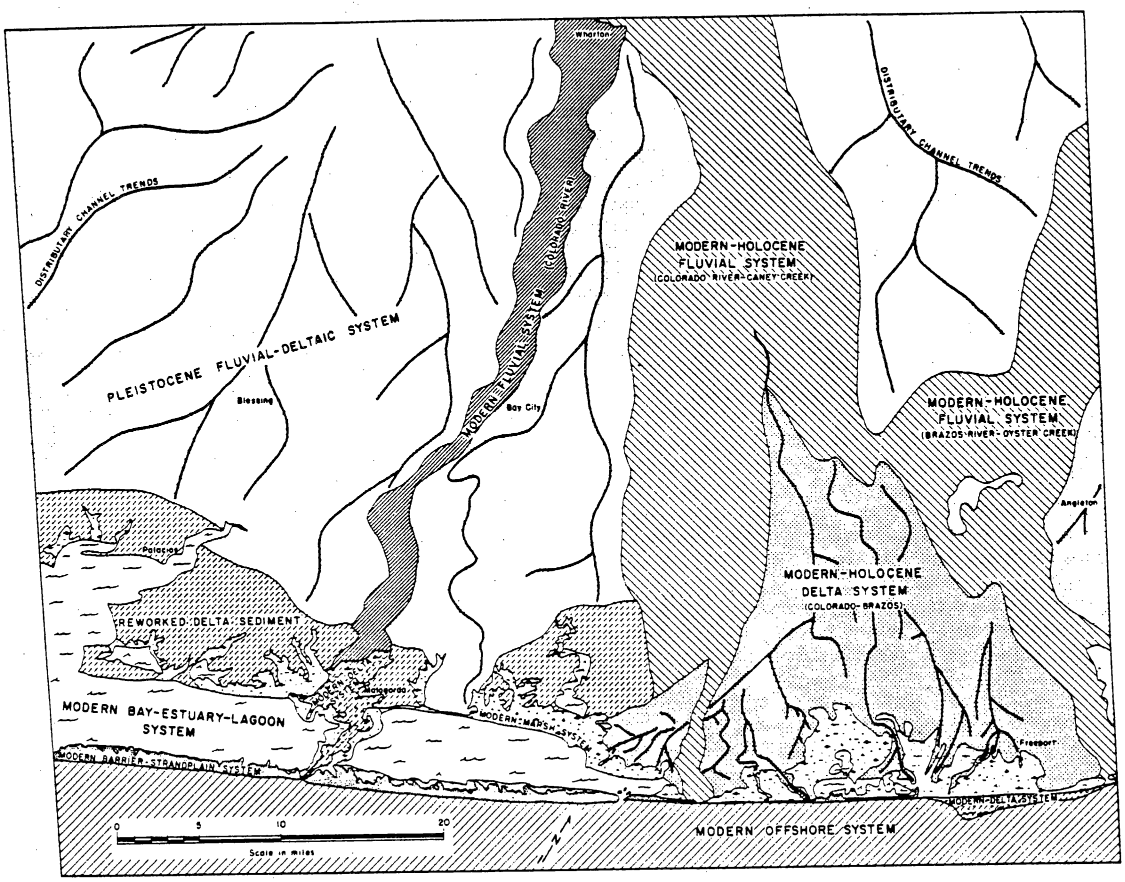


Figure 4. Fluvial and deltaic systems of the Colorado and Brazos Rivers in the Bay City–Freeport area. (From McGowen and others, 1976.)

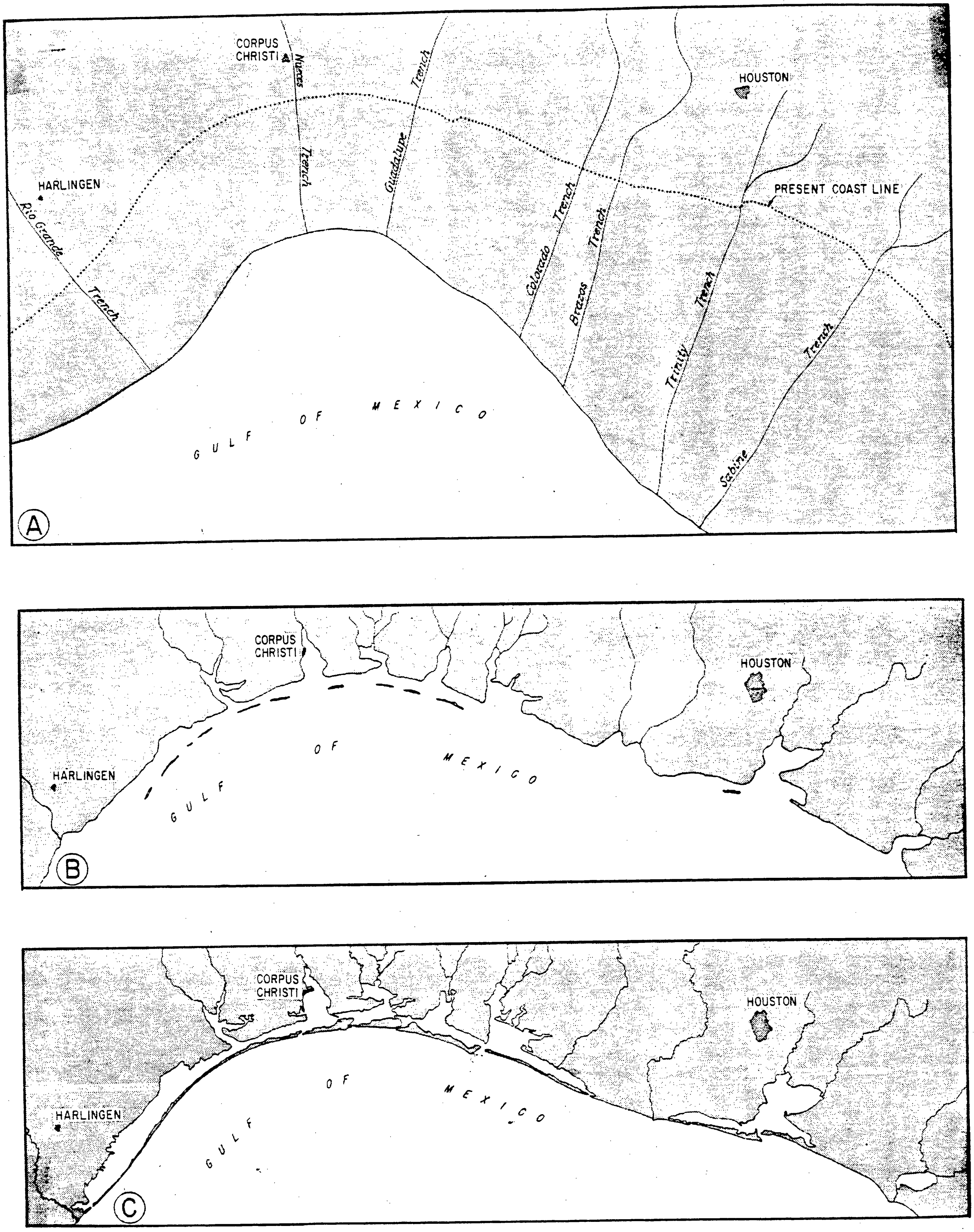


Figure 5. Development of the Texas shoreline through the late Pleistocene falling sea level stage (A), the early recent sea level stage (B), and recent standing sea level stage (C). (From LeBlanc and Hodgson, 1959.)

having surface areas of less than 809 hectares (2,000 acres), but some are very large, having surface areas of over 40,469 hectares (100,000 acres) at MWH (table 1). These systems are characterized by diverse climatic conditions and hydrologic features. Tables 2 through 7 provide a summary of some of the coastal processes and climatic conditions that affect these diverse bay–estuary–lagoon systems and their adjacent Gulf shorelines.

Climate

The bay–estuary–lagoon system along the Texas Coastal Zone is affected by a diverse climatic setting that systematically changes down the coast. Climate along the upper Texas coast in the Beaumont–Port Arthur and Galveston–Houston areas is humid (Thornthwaite, 1948) (tables 2 and 3). Average annual precipitation ranges from approximately 127 cm (50 in) in the Beaumont–Port Arthur area to 102 cm (40 in) in the Galveston–Houston area (Fig. 6). Between 1931 and 1960, the upper Texas coast had from 13 cm (5 in) to more than 30 cm (12 in) of excess moisture from precipitation after evaporation and plant transpiration (Fisher and others, 1972, 1973). Temperatures generally range from average winter lows of 7 to 9°C (near 45°F) to average summer highs in the low to mid-30's (°C) (90 to 95°F). Two principal wind regimes dominate the Texas Coastal Zone—persistent, southeasterly winds from March through November and short-lived but strong northerly winds from December through February (Fisher and others, 1972, 1973).

Climate along the middle Texas coast from the Bay City–Freeport area to the Corpus Christi area is subhumid to dry subhumid (Thornthwaite, 1948) (tables 4, 5, and 6). Average annual precipitation ranges from near 127 cm (50 in) the Bay City–Freeport area to 81 cm (32 in) in the Corpus Christi area (Fig. 6). Between 1931 and 1960, the middle Texas coast had from 10 cm (4 in) of excess moisture from precipitation after evaporation in the eastern part of the Bay City–Freeport area to a precipitation deficit of about 30 to 41 cm (12 to 16 in) in the Corpus Christi area (Brown and others, 1976; McGowen and others, 1976a, 1976b). Temperatures generally range from

Table 1. **Dimensions and average tidal range of bays–estuaries–lagoons. (Modified from Diener, 1975.)**

Location	Surface area		Depth at mean low water**		Average tidal range
	Mean low water	Mean high water*	Maximum	Average	
	Acres		Feet		
Sabine Lake					
Sabine Lake	43.960	44.830	24	5.1	0.2
Sabine Pass	1.360	1.360	40		
Galveston Bay					
East Bay	33,370	33.690	12	3.3	1.2
Trinity Bay	83,310	86,240	17	5.2	1.0
Galveston Bay (upper)	69.890	70,080	42	5.7	1.0
Galveston Bay (lower)	89.380	90,390	44	6.5	1.4
Lake Anahuac (Turtle Bay)	4.660	4.850	5	2.1	--
Scott-San Jacinto Bay	3.230	4.310	40	1.8	1.0
Clear Lake	1.260	1.280	14	2.7	0.9
Dickinson Bay	1.520	1.540	6	2.1	0.7
Moses Lake (Dollar Bay)	2.130	2.140	36	5.2	0.5
Offats Bayou	1.180	1.200	28	14.5	1.0
Jones Lake	1.040	1,050	2	1.6	1.0
West Bay	44.390	45.420	25	3.9	0.9
Chocolate Bay	4.890	4.920	12	2.6	0.6
Bastrop-Oyster Bay	9.690	10.410	20	3.2	0.7
Matagorda Bay					
East Matagorda Bay	37.810	39.080	5	3.4	0.4
Matagorda Bay	167.570	170.130	36	8.0	0.7
Oyster Lake	2.450	2 570	12	2.7	0.5
Tres Palacios Bay	9.440	9.860	12	4.1	0.6
Turtle Bay	1.280	1.760	5	2.5	0.6
Carancahua Bay	12.160	12.300	7	3.8	0.5
Salt. Redfish Lakes	920	950	4	1.2	0.5
Keller Bay	4.770	4.850	8	3.2	0.6
Lavaca Bay	39.970	40.080	36	4.2	0.7
Swan Lake	860	880	3	1.4	0.1
Lavaca River Estuary	740	760	13	8.0	0.2
Chocolate Bay	1.440	1.760	12	2.7	0.5
Powderhorn Lake	2.890	2.970	4	2.2	0.7
Cedar Lakes Complex	3.760	3.840	12	2.1	0.5

Table 1 (cont.)

Location	Surface area		Depth at mean low water**		Average tidal range
	Mean low water	Mean high water*	Maximum	Average	
	---Acres---		---Feet---		
San Antonio Bay					
Espiritu Santo Bay	38.940	40.630	14	5.9	0.3
San Antonio Bay	76.530	77,700	12	4.6	0.3
Guadalupe Bay	2.070	2,090	9	2.7	0.2
Mission Lake	1.820	2,400	--	--	--
Hynes Bay	6.580	6,610	3	2.4	0.2
Ayres Bay	2,220	2,550	12	3.2	0.3
Mesquite Bay	8.080	9.220	12	3.4	0.2
Copano Bay					
St. Charles Bay	8.410	8,730	6	3.6	0.2
Mission Bay	3,760	3,760	2	1.9	0.1
Copano Bay	41,740	42,930	9	3.7	0.3
Port Bay	1.650	2,000	9	2.2	0.2
Mission Lake	100	100	--	--	--
Aransas Bay	56.220	59,220	25	7.8	0.4
Corpus Christi					
Redfish Bay	9.630	13.420	17	2.0	0.4
Corpus Christi Bay	73.820	75.560	40	10.5	0.7
Nueces Bay	18.470	18,550	3	2.2	0.4
Oso Bay	5.070	5,070	15	1.6	0.9
Laguna Madre					
Upper Laguna Madre	47.240	68,360	12	2.8	0.7
Lower Laguna Madre	175.160	329.740	26	4.7	1.0
South Bay-La Badilla Grande Complex	4.380	7,300	36	1.5	1.5
Baffin Bay	31.870	32,610	12	7.7	0.5
Alazan Bay	13 860	14,750	4	2.9	0.5
Cayo del Infernillo	700	1.630	2	0.7	0.5
Laguna Salada	3.230	3.530	6	2.8	0.5
Cayo del Grullo	4.470	8,470	6	2.8	0.5

*Does not include peripheral marsh areas.

**Exclusive of navigation channels

Table 2. Generalized characteristics of active coastal processes and conditions in the Beaumont–Port Arthur area. (From White and others, 1987.)

Climatic zone:	Humid *(Thornthwaite, 1948)*
Average annual precipitation:	51.5 to 55.7 inches/yr (130.8 to 141.5 cm/yr) *(Fisher and others, 1973)*
Dominant wind directions:	Southeasterly, northerly *(Fisher and others, 1973)*
Astronomical tidal range: Gulf shoreline (Sabine Pass jetty) Diurnal range: Bay shoreline (average, Sabine Lake):	 2.5 ft (0.8 m) *(U.S. Department of Commerce, 1978)* 0.2 ft (6 cm) *(Diener, 1975)*
Tidal current velocities: Sabine Pass Average maximum flood: Average maximum ebb:	 2.7 ft/s (0.8 m/s) *(U.S. Department of Commerce, 1983)* 2.9 ft/s (0.9 m/s) *(U.S. Department of Commerce, 1983)*
Wave height (Gulf): (Caplan, Texas) Onshore wave height:	 Between 2.5 and 3.5 ft (0.8 and 1.1 m) about 65% of the time *(U.S. Army Corps of Engineers, 1956)*
Direction of net longshore sediment transport:	Southwesterly *(Fisher and others, 1973)*
Maximum recorded hurricane surge height on open coast: At Sabine Pass: Near High Island:	 6.7 ft (2.0 m) above MSL *(U.S. Department of Commerce unpublished data)* 4.2 ft (1.3 m) MSL *(Bodine, 1969)*
Hurricane probability:	12% in any one year *(Simpson and Lawrence, 1971)*
Gulf shoreline change, Sabine Pass to Bolivar Roads from 1882 to 1974:	Net rates are minor or moderate except for extreme net accretion of 26 and 28 ft/yr (7.9 to 8.5 m/yr) at points adjacent to the jetties at Sabine Pass and Galveston Harbor. Net erosion occurred at most other coastal points and ranged from 1 to 17.4 ft/yr (0.3 to 5.3 m/yr) and averaged 6.4 ft/yr (2.0 m/yr) *(Morton, 1975)*.
Subsidence: Sabine Pass: Estimated rate based on tide-gauge records: Magnitude related to withdrawal of oil, gas, associated ground water, and solution mining of sulfur:	 0.5 inch/yr (1.25 cm/yr) during 1960–1969 *(Swanson and Thurlow, 1973)* Generally less than 0.5 ft (0.15 m) but locally exceeding 1 ft (0.3 m) during 1918–1977 *(Ratzlaff, 1980)*

Table 3. Generalized characteristics of active coastal processes and conditions in the Galveston–Houston area. (From White and others, 1985.)

Climatic zone:	Humid *(Thornthwaite, 1948)*
Average annual precipitation:	41.8 to 51.5 inches/yr (106.2 to 130.8 cm/yr) *(Fisher and others, 1972)*
Dominant wind directions:	Southeasterly, northerly *(Fisher and others, 1972)*
Average wind speed (in 1978 at Texas City):	6.8 knots (12.6 km/hr) *(Shew and others, 1981)*
Astronomical tidal range: Gulf shoreline (Galveston Pleasure Pier) Mean diurnal: Bay shoreline (mean):	 2.1 ft (0.6 m) *(U.S. Department of Commerce, 1978)* 0.5 to 1.4 ft (0.2 to 0.4 m) *(Diener, 1975)*
Tidal current velocities: Bolivar Roads Average maximum flood: Average maximum ebb:	 3.3 knots (1.7 m/sec) *(Bernard and others, 1959)* 4.3 knots (2.2 m/sec) *(Bernard and others, 1959)*
Wave height (Gulf): (Caplan, Texas) Onshore wave height:	 Between 2.5 and 3.5 ft (0.8 and 1.1 m) about 65% of the time, *(U.S. Army Corps of Engineers, 1956a)*
Direction of net longshore sediment transport:	Southwesterly *(Fisher and others, 1972)*
Maximum hurricane surge height on open coast:	12.7 ft (3.9 m) above MSL *(Bodine, 1969)*
Hurricane frequency:	12% in any one year *(Simpson and Lawrence, 1971)*
Gulf shoreline change, Bolivar Roads to San Luis Pass from 1850-52 to 1973-74:	Total gain from accretion of 1,074 acres and loss from erosion of 1,183 acres; net loss of 109 acres *(Morton, 1977)*
Subsidence: Pasadena - Houston Ship Channel area:	 8.5 to 9 ft (2.6 to 2.7 m) during 1906-1973 *(Ratzlaff, 1980)*
Faulting: Houston metropolitan area:	 Offset by at least 160 faults *(Verbeek and Clanton, 1981)*

Table 4. Generalized characteristics of active coastal processes and conditions in the Bay City–Freeport area. (From White and others, 1988.)

Climatic zone:	Subhumid *(Thornthwaite, 1948)*
Mean annual precipitation:	40.6 to 49.2 inches (103.1 to 124.9 cm) *(McGowen and others, 1976)*
Dominant wind directions:	Southeasterly, northerly *(McGowen and others, 1976)*
Astronomical tidal range:	
Gulf shoreline (Freeport Harbor)	
Diurnal range:	1.8 ft (0.5 m) *(U.S. Department of Commerce, 1978)*
Mean:	.09 ft (0.3 m) *(U.S. Department of Commerce, 1978)*
Bay shoreline (Matagorda Bay):	0.5 to 0.7 ft (0.2 m) *(McGowen and Brewton, 1975)*
Direction of net longshore sediment transport:	Southwesterly *(McGowen and others, 1976)*
Estimated maximum hurricane surge height at Freeport:	9.5 ft (2.9 m) above MSL *(Bodine, 1969)*
Hurricane frequency:	Probability of occurrence along 50-mi (80.5-km) segment of coast in Bay City–Freeport area: 14% in any one year *(Simpson and Lawrence, 1971)*
Net rate of Gulf shoreline erosion over period of about 120 yr:	
Matagorda Peninsula:	About 2 to 3 ft/yr (0.6 to 0.9 m/yr), on average, although exceeding 10 ft/yr (3 m/yr) just south of Brown Cedar Cut *(Morton and others, 1976)*
New Brazos River delta to Brown Cedar Cut:	12.7 ft/yr (3.9 m/yr), on average *(Morton and Pieper, 1975a)*
Subsidence:	
Freeport:	
Estimated rate based on tide-gauge records:	0.44 inch/yr (1.12 cm/yr) during 1959–1971 *(Swanson and Thurlow, 1973)*
Magnitude related to withdrawal of ground water:	Generally less than 0.5 ft (0.15 m) but locally exceeding 2 ft (0.6 m) during 1906–1973 *(Ratzlaff, 1980)*

Table 5. Generalized characteristics of active coastal processes and conditions in the Port Lavaca area. (From White and others, 1989.)

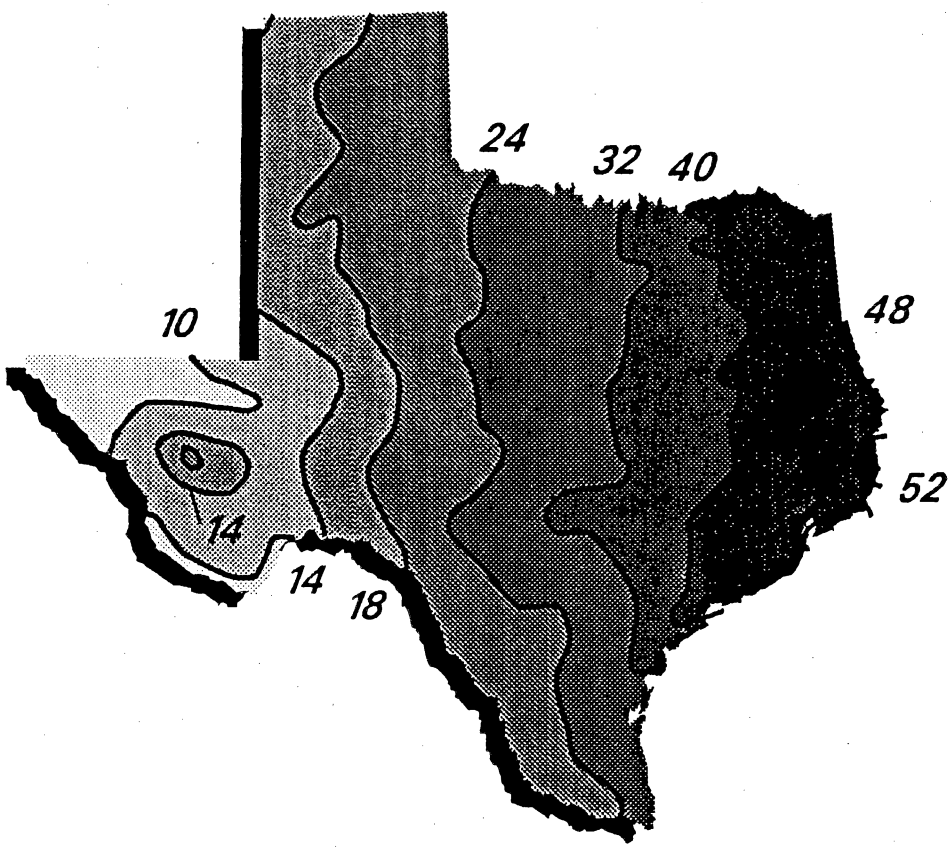
Climatic zone:	Subhumid *(Thornthwaite, 1948)*
Mean annual precipitation:	32 to 39 inches (81.3 to 99.1 cm) *(McGowen and others, 1976)*
Dominant wind direction:	Southeast, north *(McGowen and others, 1976)*
Astronomical tidal range: Gulf shoreline (Pass Cavallo) Diurnal range: Mean: Bay shoreline (Port Lavaca) Diurnal range: Mean:	 1.4 ft (0.4 m) *(U.S. Department of Commerce, 1978)* 0.7 ft (0.2 m) *(U.S. Department of Commerce, 1978)* 0.7 ft (0.2 m) *(U.S. Department of Commerce, 1978)* 0.3 ft (0.1 m) *(U.S. Department of Commerce, 1978)*
Direction of net longshore sediment transport (Gulf shoreline):	Southwesterly *(McGowen and others, 1976)*
Estimated peak hurricane surge height on open coast near Port O'Connor:	12.3 ft (3.7 m) m.s.l. *(Bodine, 1969)*
Hurricane frequency:	Probability of occurrence along 50-mi (80.5-km) segment of coast in Port Lavaca area is 9 percent in any one year *(Simpson and Lawrence, 1971)*
Net rate of Gulf shoreline accretion or erosion over period of 117 yr: Matagorda Island:	 Less than 1 ft (0.3 m) of average annual accretion in southern half of island; accretion rates in the northern half range from 1.1 to 9.1 ft (0.3 to 2.8 m) per year. Erosion rates near Pass Cavallo range from 5.1 to 17.3 ft (1.6 to 5.3 m) per year *(Morton and Pieper, 1976)*
San José Island:	Less than 1.5 ft (0.5 m) of average annual erosion and accretion *(Morton and Pieper, 1976)*

Table 6. Generalized characteristics of active coastal processes and conditions in the Corpus Christi area. (Modified from White and Galloway, 1977.)

Climatic zone:	Dry subhumid (*Thornthwaite, 1948*)
Mean annual precipitation:	30 to 32 inches/yr (76 to 81 cm/yr) (*Carr, 1967*)
Dominant wind directions:	Southeasterly; north-northeasterly (*Lohse, 1955*)
Average wind speed (in 1980):	12.8 mi/hr (20.6 km/hr) (*U. S. Dept. of Commerce, 1980a*)
Direction of net sand transport by winds:	Northwesterly (*Hunter and others, 1972*)
Astronomical tidal range:	
Gulf shoreline (Port Aransas)	
Mean diurnal	1.5 ft (0.48 m) (*Hayes, 1965*)
Maximum diurnal	2.5 ft (0.76 m) (*Collier and Hedgpeth, 1950*)
Bay shoreline, mean	Approx. 0.5 ft (0.15 m) (1 ft [0.3 m] lower than Gulf) (*Watson and Behrens, 1976*)
Tidal current velocities:	
Aransas Pass	
Average maximum flood	2.0 ft per second (fps) (0.6 m/s)
Average maximum ebb	1.5 fps (0.5 m/s) (*U. S. Dept. of Commerce, 1980b*)
Fish Pass	
Average maximum	3 fps (0.9 m/s)
Usual value	Below 2 fps (0.6 m/s) (*Defehr and Sorensen, 1973; Watson and Behrens, 1976*)
Wave height (Gulf):	
Usual height	Below 4 ft (1.2 m) (*Davis and Fox, 1972*)
Mean height	2.6 ft (0.8 m) (*Watson and Behrens, 1976*)
Longshore current velocities (Gulf):	
Range	0 to 3.9 fps (0 to 1.2 m/s) (*Davis and Fox, 1972*)
Average	0.38 fps (0.1 m/s) (fall) and 0.7 fps (0.2 m/s) (winter) (*Davis and Fox, 1972*)
Direction of net longshore sediment transport:	Southwesterly (*Lohse, 1955; Behrens and Watson, 1974*)
Average rate of Gulf shoreline erosion over period of about 100 years (Mustang Island):	2.0 ft/yr (0.6 m/yr) (*Morton and Pieper, 1977*)
Maximum hurricane surge height recorded at Aransas Pass (1919 to 1977):	11.5 ft (3.5 m) (1919) (*Price, 1956*)
Hurricane frequency:	
Probability of occurrence along 50-mile (80.5-km) segment of coast in Corpus Christi area	7% in any one year (*Simpson and Lawrence, 1971*)

Table 7. Generalized characteristics of active coastal processes and conditions in the Brownsville–Harlingen area. (From White and others, 1986.)

Climatic zone:	Semiarid (*Thornthwaite, 1948*)
Mean annual precipitation:	25 to 26 inches (63 to 66 cm) (*U.S. Department of Commerce, 1982*)
Dominant wind directions:	Southeasterly, north (*Brown and others, 1980*)
Average annual wind speed (Brownsville):	Prevailing south to southeasterly - 11.8 mi/hr (19 km/hr); north winds up to 26 mi/hr (40 km/hr) (*Espey, Huston and Associates, Inc., 1981*)
Direction of net sand transport by winds:	Northwestward (*Brown and others, 1980*)
Astronomical tidal range:	
Gulf shoreline	
Mean diurnal:	1.4 ft (0.4 m) (*U.S. Department of Commerce, 1978*)
Lower Laguna Madre shoreline (mean):	1 ft (0.3 m) (*Diener, 1975*)
Tidal current velocities:	
Brazos Santiago Pass	
Average maximum flood:	0.94 knots (0.5 m/s) (*Espey, Huston and Associates, Inc., 1981*)
Average maximum ebb:	0.73 knots (0.4 m/s) (*Espey, Huston and Associates, Inc., 1981*)
Wave height (Gulf):	
Usual height:	2.5 to 3.5 ft (0.8 to 1.1 m) (*U.S. Army Corps of Engineers, 1956b*)
Longshore current velocities (Gulf):	Up to 3 knots (5.6 km/hr) (*Lohse, 1952*)
Direction of net longshore sediment transport:	Northward (*Brown and others, 1980*)
Net rate of Gulf shoreline erosion over period of about 120 years (South Padre Island):	<1 to 13 ft/yr (.02 to 4.0 m/yr) (*Morton and Pieper, 1975b*)
Estimated peak hurricane surge height recorded at Port Isabel:	11 ft (3.4 m) above MSL (*Bodine, 1969*)
Hurricane frequency:	
Probability of occurrence along 50-mi (80.5-km) segment of coast in Brownsville-Harlingen area:	8% in any one year (*Simpson and Lawrence, 1971*)



EXPLANATION

40 Average annual precipitation, 1951-1980 (inches)

Figure 6. Average annual precipitation in Texas. (Modified from Riggio and others, 1987.)

winter minimum lows of 8 to 9°C (46 to 48°F) to average maximum summer highs in the low to mid-30's (°C) (90 to 95°F)

Along the lower Texas coast in the Kingsville and Brownsville–Harlingen areas, climate is semiarid (Thornthwaite, 1948) (table 7). The annual rainfall ranges from 66 cm (26 in) along the southern coastline in the Brownsville–Harlingen area to 87.6 cm (34.5 in) along the northern coastline of the Kingsville area (Brown and others, 1977, 1980). The Brownsville area experienced a precipitation deficit of 58 to 79 cm (23 to 31 in) of moisture between 1931 and 1960 (Brown and others, 1980). Between 1931 and 1960 the average annual mean free-air temperature in the area was about 23 to 23.1°C (73 to 73.5°F) (Brown and others, 1977, 1980).

Salinity

Water salinities vary considerably both between bay–estuary–lagoon systems and within each system, in part because of the regional variations both in fresh-water inflows from rivers and streams and in salt water interchange from tidal passes. Compounding the complexity in each system are seasonal and cyclic climatic variations that produce substantially higher than normal salinities during dry periods and lower than normal salinities during wet periods.

Average salinities in Texas estuaries range from a low of about 2 parts per thousand (ppt) in the Sabine–Neches estuary (Armstrong, 1982) to over 54 ppt in the upper part of Cayo del Grullo in the Baffin Bay system (Brown and others, 1977). On the upper coast, salinities in Sabine Lake generally range from less than 10 ppt in the upper part of the lake to between 10 and 20 ppt in the tidally influenced lower part (Fisher and others, 1972). Salinity decreases with increasing distance from Sabine Pass, such that salinity is slightly lower in the central part of Sabine Lake than in the lower open-bay area.

Of the bays in the Galveston Bay system, salinities are generally highest in West Bay, followed, in order of decreasing average salinity, by Galveston, East, and Trinity Bays. Average salinities in West Bay are generally more than 15 ppt and range into the 30's, which is in marked contrast to Trinity Bay, where average salinities range from less than 5 to about 10 ppt (White and others, 1985). Salinities in Trinity Bay can drop to 0 ppt or exceed 25 ppt.

Salinity data for the Matagorda Bay system, including Matagorda, East Matagorda, Tres Palacios, Carancahua, and Lavaca Bays, can be found in Ward and Armstrong (1980) and Jones and others (1986). Data for San Antonio Bay are reported in Matthews and others (1974). Salinities in Lavaca and Matagorda Bays generally increase toward Pass Cavallo. Average salinities in Matagorda Bay are generally above 20 ppt and range into the 30's. In East Matagorda Bay salinities range from 15 ppt near the Colorado River delta to 17.4 in the northeastern part of the bay. Matthews and others (1974) found that salinities in upper San Antonio Bay ranged from approximately 0.5 to 9.0 ppt and in the most gulfward region from approximately 6.0 to 26.0 ppt.

In the Corpus Christi area, salinities are generally highest in upper Laguna Madre, followed, in order of decreasing average salinity, by Corpus Christi, Redfish, Aransas, Nueces, and Copano Bays (Holland and others, 1975; Brown and others, 1976; Hildebrand and King, 1978). Average salinities in upper Laguna Madre are generally above 30 ppt and range into the 40's and occasionally higher, which is in marked contrast to Copano Bay, where average salinities range from about 10 to 15 ppt, increasing toward the mouth of the bay. Average monthly median salinities in the upper part of Copano Bay fluctuate around 14.4 ppt, and in the upper part of Nueces Bay fluctuate around 21.7 ppt (Texas Department of Water Resources, 1981a). Monthly mean salinities in mid-Corpus Christi Bay vary, as demonstrated by measurements in 1973 that show a high of 35 ppt in February and a low of 15.6 ppt in October (Holland and others, 1975).

Average salinities for the bay–estuary–lagoon system in the Kingsville area, including Baffin, Alazan, Cayo del Grullo, and that part of upper Laguna Madre near Baffin Bay, are probably the highest on the Texas coast. Calculated average surface salinities of Baffin Bay and associated water bodies range from a low of just less than 50 ppt in Laguna de los Olmos to over 54 ppt in Cayo del Grullo (Brown and others, 1977). Average surface salinities for eight stations in Baffin Bay range from approximately 40 ppt in May 1966 to 70 ppt in December 1964 (Behrens, 1966).

In the Brownsville area, salinities generally increase from the southern end of lower Laguna Madre at Port Isabel to north of Port Mansfield (Brown and others, 1980; Espey, Huston and Associates, Inc., 1981). Salinities in the Port Isabel area range from 23 to 36 ppt and are influenced by the exchange of Gulf water through Brazos Santiago Pass. Salinity at the northern end of lower Laguna Madre ranges from 20 to 40 ppt and averages about 38 ppt.

Bathymetry

Bathymetric data (table 1) are taken from Diener (1975) and represent averages of the most recent soundings at mean low water (MLW) exclusive of navigation channels. Average depths range from 0.2 m (0.7 ft) in Cayo del Infernillo of the Baffin Bay system to 4.4 m (14.5 ft) in Offats Bayou of the Galveston Bay system. Average depths of the larger bay–estuary–lagoon systems range from 0.9 m (2.8 ft) in upper Laguna Madre to 3.2 m (10.5 ft) in Corpus Christi Bay. Many of the bays are shallow, with average depths of less than 1.2 m (4 ft). Maximum depths occur in the navigation channels and near the tidal passes.

Tides

Astronomical tidal variations in Texas estuaries are small compared with estuaries of the Atlantic and Pacific Coasts. In the Gulf of Mexico the principal variations in the tide are due to changing declination of the moon (U. S. Department of Commerce, 1978). Tidal range in the northwest Gulf of Mexico during maximum declination of the moon is about 0.8 m (2.6 ft) and at minimum declination about 0.2 m (0.7 ft)(Ward and others, 1980). Meteorological events are more important than astronomical tides in affecting estuaries, as they alternately expose and flood the greatest area of tidal flat and marsh (Collier and Hedgpeth, 1950).

The most noticeable fluctuations in bay levels are caused by direction and force of the wind or wind tides. The amount of open-bay fetch and direction of wind tides control the effectiveness of wind-tidal activity (Brown and others, 1976). For example, broad fetch, as in Trinity Bay and the western arm of Matagorda Bay, and persistent southeast winds aligned with the axis of the bay, result in high wind tides that may build tide heights 0.6 to 0.9 m (2 to 3 ft) above normal (Holliday, 1973). Frontal passage, such as during a norther, can also drastically affect the wind tides and estuarine water levels.

Relative Sea-Level Rise

The bay–estuary–lagoon system is affected by many interactive processes. One of the most important at work along the Texas coast today is relative sea-level rise. Stated very simply, for example, if coastal wetlands do not receive and trap sufficient sediments (organic or inorganic) so that the aggradation rate (vertical accretion) is equal to or greater than the rate of relative sea-level rise, the wetlands will ultimately be lost and replaced by open water. If bay–estuary–lagoon sedimentation rates do not keep pace with rates of relative sea-level rise, water depths will increase through time; if the sedimentation rates exceed relative sea-level rise, water depths will decrease and the bay or estuary may eventually fill with sediment.

Relative sea-level rise as used here refers to a rise in sea level with respect to the surface of the land, whether it is caused by actual sea-level rise or land-surface subsidence; the current general trend along the Texas coast involves both of these processes working together.

Sea-level fluctuations occur for a variety of reasons and on broad spatial and temporal scales. Nummedal (1983) reviewed sea-level fluctuations and how they are affecting the coast of Louisiana, and Morton and Price (1987) present information on Late Quaternary sea-level fluctuations and their relation to the shallow-water depositional complexes on the Texas coastal plain and shelf. Nummedal (1983) notes that sea-level changes can be categorized into two fundamental groups that operate on global and local scales. Among the global factors are (1) the volume of the ocean basins controlled by sea-floor-spreading rates, sedimentation, and opening and closing of marginal seas, and (2) the volume of oceanic water that has changed in response to glaciation and, possibly, water temperature. Local factors include subsidence of continental margins, movement of the land surface along faults, compaction due to dewatering of sediments, and many atmospheric factors.

Various methods have been used to measure changes in mean sea level; a primary method during historic time is to compare records from tide gauges and examine the trends over as long a period as the records allow (Marmer, 1954; Swanson and Thurlow, 1973). Using this method along the Gulf Coast, Swanson and Thurlow (1973) concluded that subsidence is an important factor with regard to mean sea-level rise.

The discussion to follow will touch briefly on eustatic (global) sea-level rise but will focus principally on compactional subsidence, natural and human-induced, because these processes appear to be the most significant with respect to the Texas coast.

Eustatic Sea-Level Rise

It is generally accepted that sea level is rising on a worldwide (eustatic) basis (Hicks, 1978; Gornitz and others, 1982), apparently in response to a global warming trend resulting from increases in atmospheric CO_2 and the resulting "greenhouse" effect (Hansen and others, 1981), which can add volume to the oceans through glacial melt water (Etkins and Epstein, 1982; Meier, 1984) and perhaps thermal expansion (Gornitz and others, 1982). The worldwide rate of sea-level rise, based on tide-gauge records over the past century, is about 1.2 mm per year (Gornitz and others, 1982). For the Gulf of Mexico and Caribbean region, the trend is approximately 2.4 mm/yr (Gornitz and Lebedeff, 1987). For the Texas coast, the rate of global sea-level rise is not nearly as significant as relative sea-level rise due to compactional subsidence (Swanson and Thurlow, 1973).

Subsidence

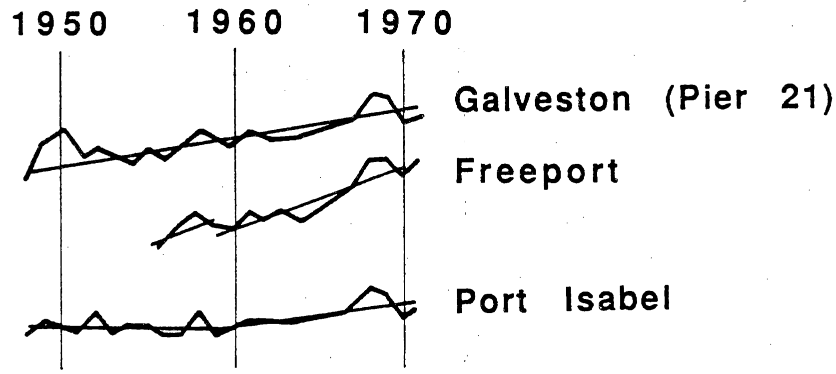
There are many causes of subsidence (Nummedal, 1983) including regional downwarping or tilting of the earth's crust due to loading, which is significant over a geologic time frame along the Texas coast (Winker, 1979), but is not as significant when viewed over a historic timeframe. Holdahl and Morrison (1974) reported slight subsidence along the Gulf Coast region ranging between 0.0 and 1.5 mm/yr, in addition to anomalous subsidence in the Houston and Corpus Christi areas (discussed in succeeding sections). The most significant subsidence along the Texas coast appears to be due to compactional subsidence, especially as affected by subsurface fluid withdrawal (water, oil, gas, and, locally, sulfur).

Using tide-gauge records along the Texas and Louisiana coasts, and comparing them with records from the more stable (tectonically or geologically) Florida coast, Swanson and Thurlow (1973) concluded that subsidence is occurring along the Texas coast at rates of from 0.5 to 1.2 cm/yr. Furthermore, they found that rates for the period from 1959 to 1971 are higher than rates before 1959 (1948–1959) (Fig. 7A). Highest rates along the Texas coast are at Sabine Pass, Freeport, and Port Aransas (1.12 to 1.28 cm/yr) and the lowest rate is at Port Isabel (0.49 cm/yr). Subsidence in the Freeport area was believed to be due to the delta environment of the Brazos River, and subsidence in the area of Pier 21 at the mouth of West Bay near Galveston to faulting and withdrawal of oil and gas (Swanson and Thurlow, 1973). Swanson and Thurlow (1973) noted that their data, which show an overall tendency toward subsidence along the entire coast, supported Shepard and Moore (1960), who suggested that much of the Texas coast could be subsiding due to sediment overburden and compaction of underlying fine sediment.

Turner (1987) evaluated Galveston's long-term tide record (1909–1982) and confirmed as had been noted by Penland and others (1988) that there has been an acceleration in the rate of relative sea-level rise. The rate from 1942 to 1962 is 0.32 cm/yr, and from 1962 to 1982, it is 1.15 cm/yr (Fig. 7B). However, Turner (1987) suggests that these variations (based on an 18.6-yr lunar epoch) are short-term fluctuations that are centered around the more constant long-term mean of 0.62 cm/yr (Fig. 7B). He suggests that, disregarding possible future accelerations due to the greenhouse effect, the rate of sea-level rise should decline during this decade if the historical trends continue.

Compactional subsidence occurs as sediments are consolidated, generally as a result of compressive forces from overlying material (sediments and water) and the dewatering of the

A)



Location	Time period	Rate (mm/yr)
Sabine Pass	1960-1969	12.5
Galveston (Pier 21)	1959-1971	6.0
Galveston (PP)	1959-1970	8.6
Freeport	1959-1971	11.2
Port Aransas	1959-1969	12.8
Port Isabel	1959-1971	4.9

B)

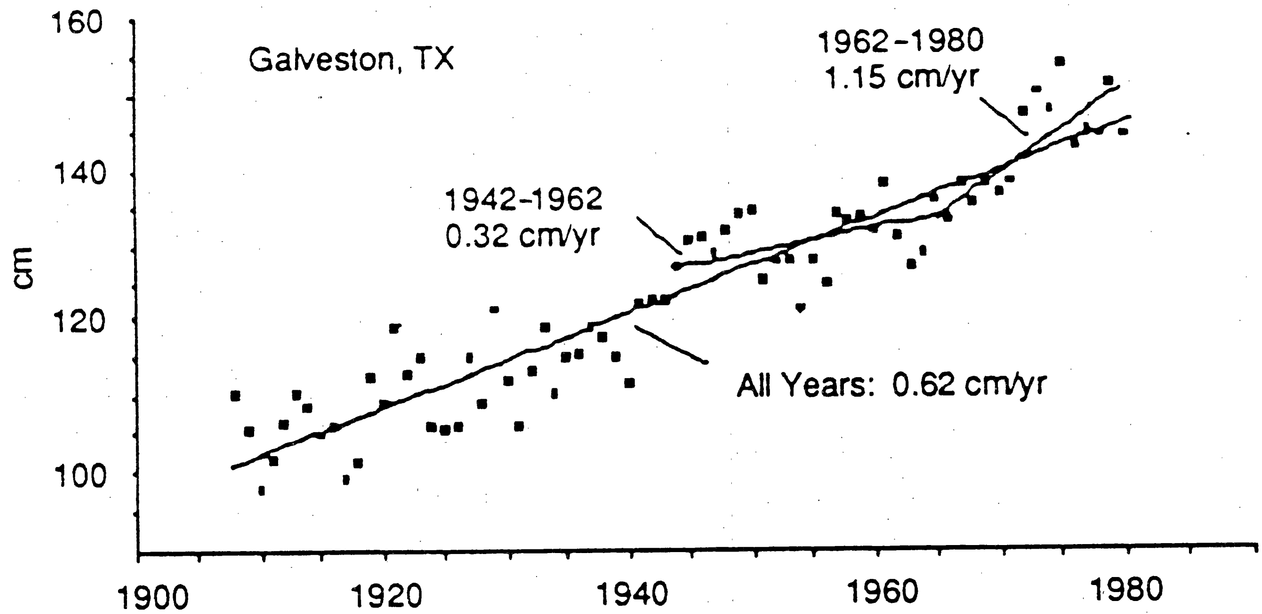


Figure 7. Subsidence and water-level changes at selected sites along the Texas coast (A), and water-level changes at Galveston (B), based on tide-gauge records. (A) (From Swanson and Thurlow, 1973; PP = Pleasure Pier). (B) Annual changes in water level at Galveston, with variations in rates of rise indicated for different periods. (From Turner, 1987).

compacting sediments either naturally or as influenced by withdrawal of fluids. In a delta plain, natural compaction and resulting subsidence are highest during the initial period after deposition and abandonment (first few hundred years), and diminish with age as the rate of sediment dewatering declines (Penland and others, 1988). Penland and others (1988) estimated the rate of subsidence in younger sediments (0-500 yr Before Present, B.P.) to be 0.62 cm/yr (0.24 in/yr), and in older sediments (500-3,000 yr B.P.) to be 0.18 cm/yr (0.07 in/yr).

The highest rates of subsidence along the Texas coast have been caused by withdrawal of underground fluids, principally water (Winslow and Doyel, 1954; Gabrysch, 1969; Brown and others, 1974; Gabrysch and Bonnet, 1975; Kreitler, 1977; Ratzlaff, 1980; Gabrysch, 1984). Production of oil and gas can also cause subsidence (Pratt and Johnson, 1926; Gustavson and Kreitler, 1976; Kreitler, 1977; Verbeek and Clanton, 1981). Extreme local subsidence has occurred in relation to sulfur mining around salt domes along the Texas Coast (Ratzlaff, 1980; Mullican, 1988).

Brown and others (1974) reported that along the Texas Coastal Zone, the rates of subsidence, both in terms of area impacted and drops in surface elevation, have progressively increased since 1940 (Fig. 8).

Subsidence in the Houston–Galveston Area

The most extensive subsidence, both in terms of vertical and areal magnitude, due to withdrawal of fluids is in the Houston–Galveston area (Ratzlaff, 1980), where more than 2.7 m (9 ft) and possibly as much as 3.0 m (10 ft) of subsidence has occurred in the vicinity of the Houston Ship Channel between 1906 and 1978 (almost 2.7 m [9 ft] of which occurred during 1943 to 1978) (Gabrysch, 1984). Maximum subsidence is in the center of a subsidence "bowl" that encompasses an area from near Freeport (where another smaller "bowl" is centered) to an area north of the Trinity River delta (Fig. 9). Average maximum rates of subsidence at the center of the "bowl" have been as high as 122 mm/yr (0.4 ft/yr) for the period 1964 to 1973 (Gabrysch and Bonnet, 1975).

According to Gabrysch and Bonnet (1975), subsidence due to withdrawal of ground water from an artesian aquifer results from a decrease of hydraulic pressure and attendant movement of water

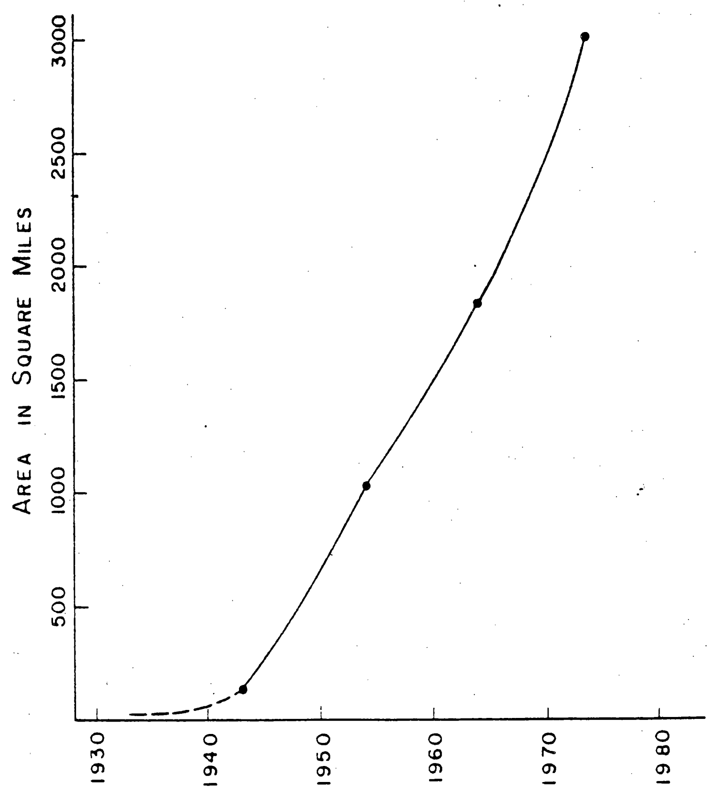


Figure 8. Cumulative area in the Texas Coastal Zone affected by land-surface subsidence in excess of 30 cm (1 ft) between 1943 and 1973. (From Brown and others, 1974.)

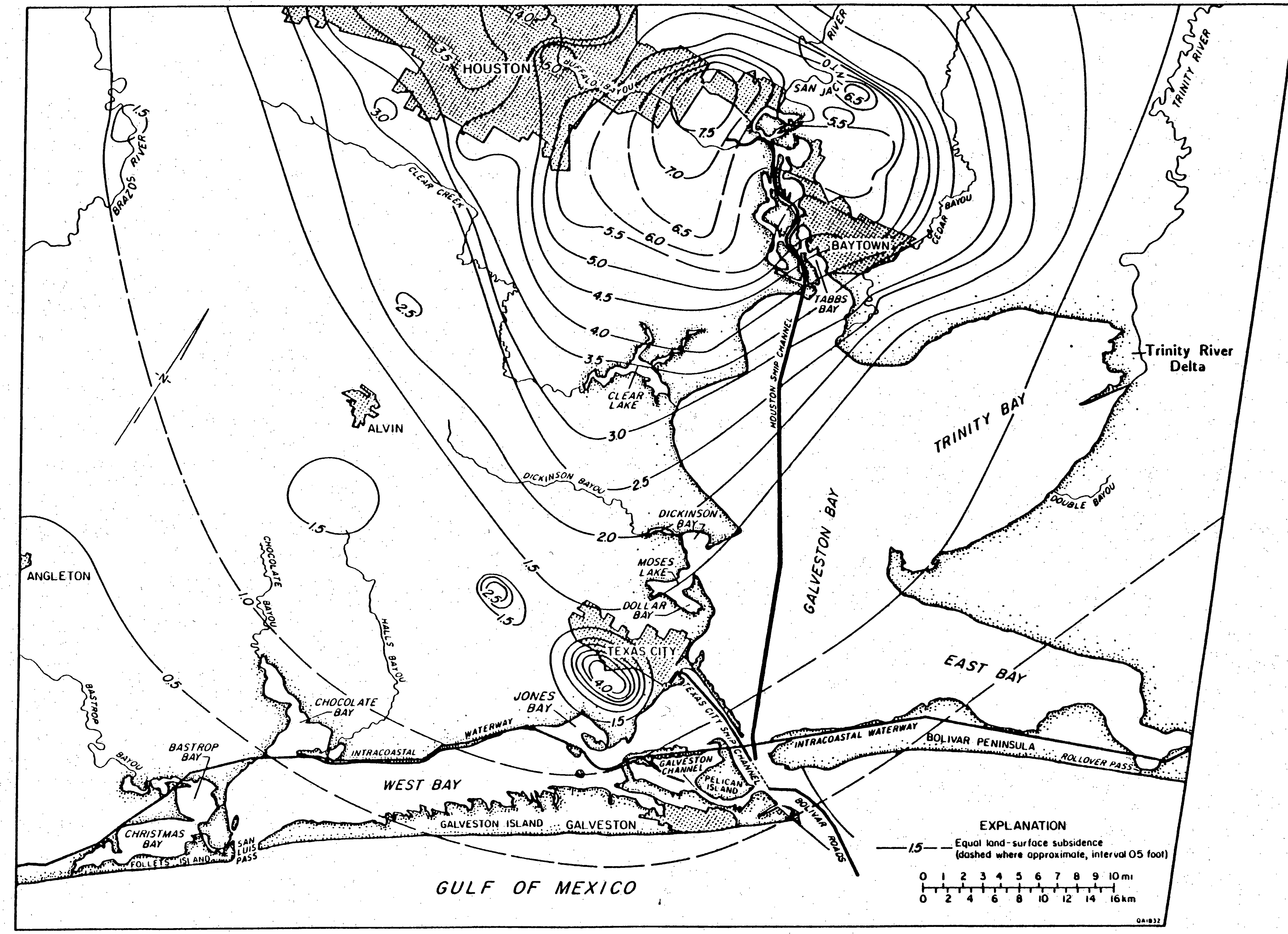


Figure 9. Land-surface subsidence in the Houston–Galveston area, 1943-1973. (From Gabrysch and Bonnet, 1975.)

from clays to adjacent sands leading to compaction of the clays. Most of the compaction is permanent because of the inelastic nature of the clay; thus, even with total recovery of artesian pressure, less than 10 percent rebound can be expected (Gabrysch and Bonnet, 1975).

Methods used by the USGS for measuring subsidence include conventional leveling, extensometers, and tide gauges (Gabrysch, 1984). Conventional leveling is the most frequently used method, and involves comparing the elevations of benchmarks that have been measured at different times using precise leveling techniques. Borehole extensometers have been used at specific locations to determine small changes in elevations; extensometers can provide very precise, continuous records with information on the compacting interval, but they are costly to install and have small areal application (Gabrysch, 1984). Subsidence can be determined by comparing tide-gauge records from two different stations, but this method is less precise than leveling and extensometers. Gabrysch (1984) reported that evaluation of tide-gauge data from five stations in Galveston Bay and Buffalo Bayou indicated that elevation change of less than 150 mm (0.5 ft) and perhaps as little as 30 mm (0.1 ft) could be detected (Fig. 10).

Other parts of the Texas coastal region, in addition to the Houston–Galveston and Freeport areas, where subsidence has occurred include (1) Beaumont–Port Arthur, where an area over Spindletop Dome has subsided about 1.5 m (5 ft) between 1925 and 1977, and an area near Port Acres has subsided approximately 0.9 m (3 ft) between 1959 to 1977, primarily due to withdrawal of oil and gas and associated ground water; (2) Jackson and Matagorda Counties inland from Matagorda Bay, where subsidence of more than 0.46 m (1.5 ft) occurred during 1943 to 1973 as a result of ground-water withdrawals; and (3) the western part of Corpus Christi, where more than 1.5 m (5 ft) of subsidence occurred during 1942 to 1975 due to withdrawals of oil, gas, and associated ground water (Ratzlaff, 1980).

Faulting and Subsidence

In some areas along the Texas coast, subsidence may be accompanied by active surface faults. A good example is a fault in the Saxet oil and gas field west of Corpus Christi (Price, 1933). The

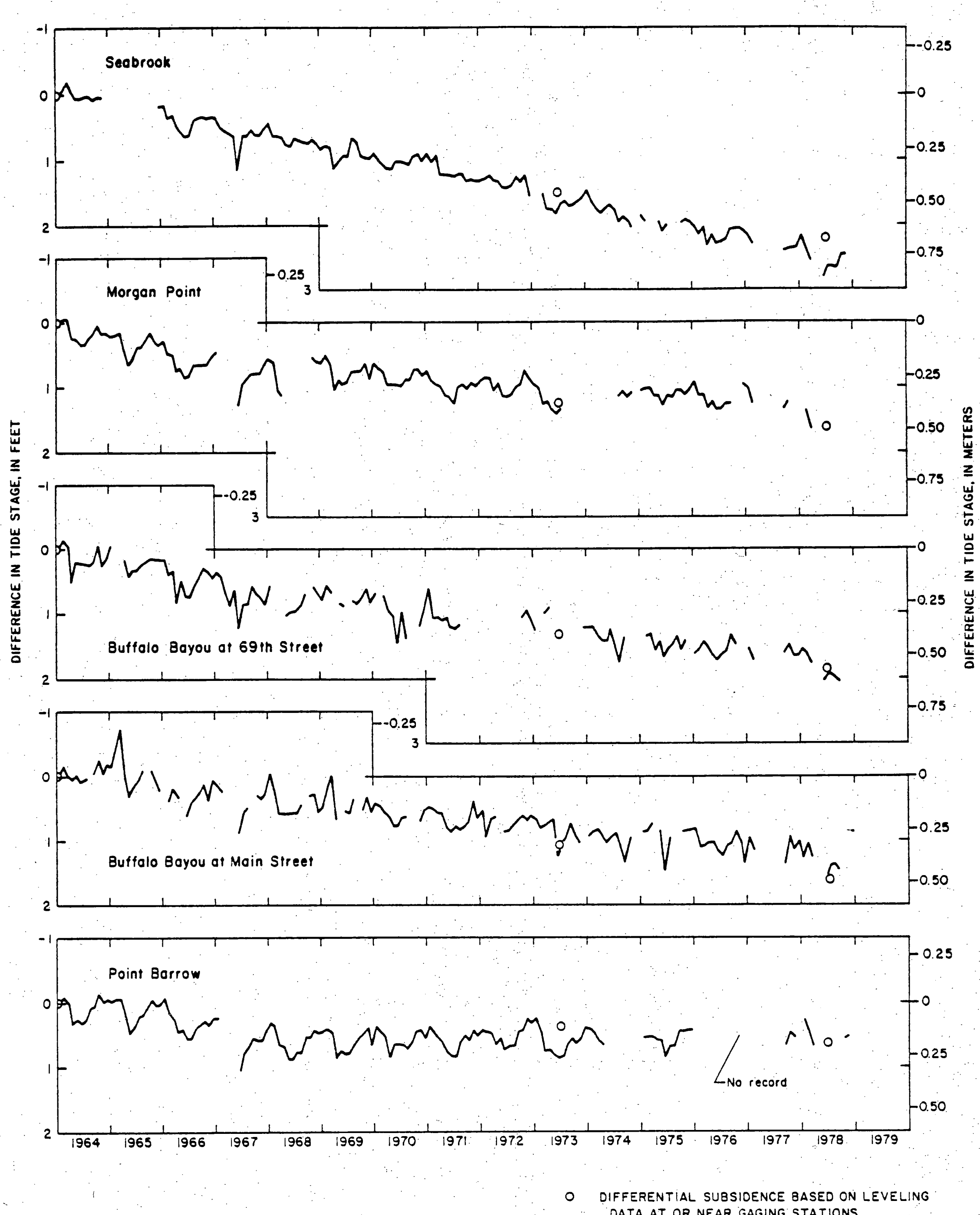


Figure 10. Subsidence as reflected by differences in mean monthly tide stage relative to the gauge record at a designated base station (Galveston Railroad Causeway in the above case). Change in the difference between mean water level as measured by tide gauges at stations a few miles apart reflects the change in elevation between the stations. (From Gabrysch, 1984.)

fault has produced a 2 m (6 ft) scarp at the surface (Gustavson and Kreitler, 1976). Profiles constructed from releveling lines across benchmarks show rapid increases in subsidence at the fault (Fig. 11). Subsidence rates during the period from 1950 to 1959 were 70 mm/yr (0.22 ft/yr), which was an increase of almost twice the rate of 40 mm/yr (0.14 ft/yr) during the period of 1942 to 1950 (Fig. 11). Gustavson and Kreitler (1976) theorized that an increase in gas production from 1950 to 1959 may have been responsible for compaction of shallow reservoir sands on the downthrown side of the fault leading to differential subsidence and accelerated fault movement. Evidence of the fault can be seen where it crosses highways and other structures. Lower elevations in the subsidence bowl inhibit drainage of surface water locally and promote ponding of water.

Although the fault in the Saxet field in the Corpus Christi area is a good example of an active surface fault, the major zone of surface faulting along the Texas coast is in the Houston–Galveston area, where 150 linear km (95 linear mi) of faulting has been reported (Reid, 1973; Brown and others, 1974). Surface faults correlate with, and appear to be extensions of, subsurface faults in many areas (Weaver and Sheets, 1962; Van Siclen, 1967; Kreitler, 1977; Verbeek and Clanton, 1981). Most of the surface faulting in the Houston metropolitan area has apparently taken place during the last few decades (Verbeek and Clanton, 1981), largely due to fluid withdrawal (water, oil, and gas), which has reinitiated and accelerated fault activity (Reid, 1973; Kreitler, 1977; Verbeek and Clanton, 1981).

The range in measurable vertical displacement of surface traces of faults is from 0 to 3.9 m (12 ft) (Reid, 1973). Rates of fault movement commonly range between 5 mm/yr and 20 mm/yr (0.2 in/yr and 0.8 in/yr) (Verbeek and Clanton, 1981), but many exceed 40 mm/yr (1.6 in/yr) (Van Siclen, 1967; Reid, 1973; Everett and Reid, 1981). Movement along surface faults apparently occurs episodically (Reid, 1973). Highways, railroads, industrial complexes, airports, homes, and other structures placed on active faults in the Houston area have undergone millions of dollars worth of damage annually (Clanton and Verbeek, 1981).

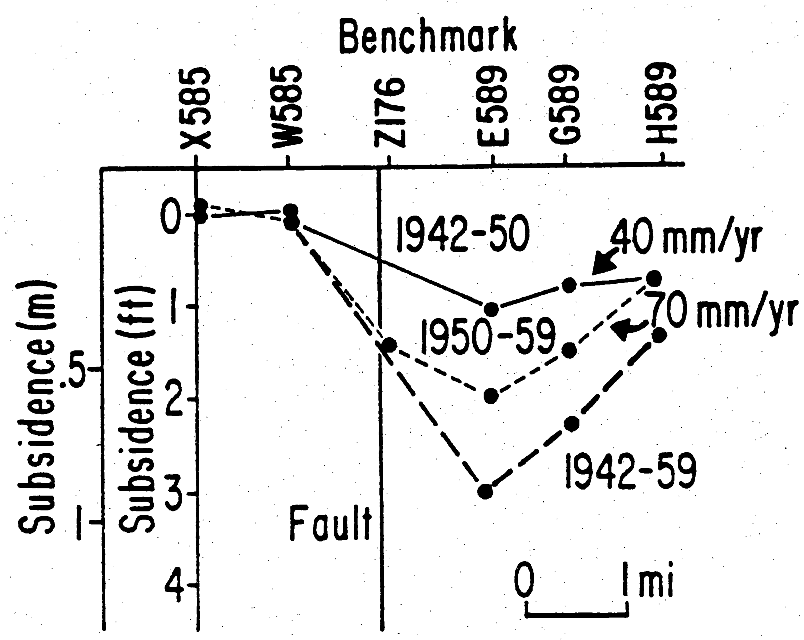


Figure 11. Subsidence over an oil and gas field. (From Kreitler, 1977.)

Effects of Subsidence and Faulting on Texas Coastal Wetlands

Subsidence in the Houston–Galveston area has had a significant effect on wetlands in the area (Johnston and Ader, 1983; White and others, 1985). One of the most dramatic examples of wetland losses due to subsidence is along the San Jacinto River. More than 560 hectares (1,389 acres) of fluvial woodlands, swamps, and marshes were displaced by open water between 1956 and 1979 (White and others, 1985). The lower reach of the San Jacinto River, near its confluence with Buffalo Bayou and the Houston Ship Channel, is in an area of subsidence (Fig. 9) caused mostly by ground-water withdrawal (Gabrysch and Bonnet, 1975).

The change in wetlands along the lower San Jacinto River valley (discussed more thoroughly in a later section) is pronounced because of the proximity of the valley to the center of maximum subsidence. However, wetlands associated with other streams and valleys located around the Trinity Bay and Galveston Bay systems are also changing as a result of human-induced subsidence and accompanying relative sea-level rise. Replacement of marshes by open water is occurring along the bay margins as well (Fig. 12). Changes in the distribution of wetlands as a result of natural compactional subsidence have also been reported along the Texas coast (Donaldson and others, 1970; McGowen and Brewton, 1975; White and others, 1988).

Faults have affected marshes from the Freeport area to Sabine Lake (White and others, 1985, 1987, 1988). As vertical displacement occurs along a fault that intersects a marsh, more frequent and eventually permanent inundation of the wetland surface on the downthrown side of the fault can lead to replacement of marsh vegetation by open water if marsh sedimentation rates do not keep pace with submergence rates (Fig. 13). This has occurred at several locations along the upper Texas coast, as exemplified by a marsh system on the bay side of Bolivar Peninsula, where approximately 500 hectares (1,230 acres) of salt-water marsh has been replaced primarily by "barren" shallow subaqueous flats and open water (Fig. 14). In this area, at least two surface faults intersect marsh substrates. Benchmark releveling profiles along State Highway 87 indicate that the faults are active; a marked increase in subsidence occurs on the downthrown side (Fig. 15). More than 25 faults that cross wetlands along the upper coast (Freeport area to Sabine Pass) have been identified

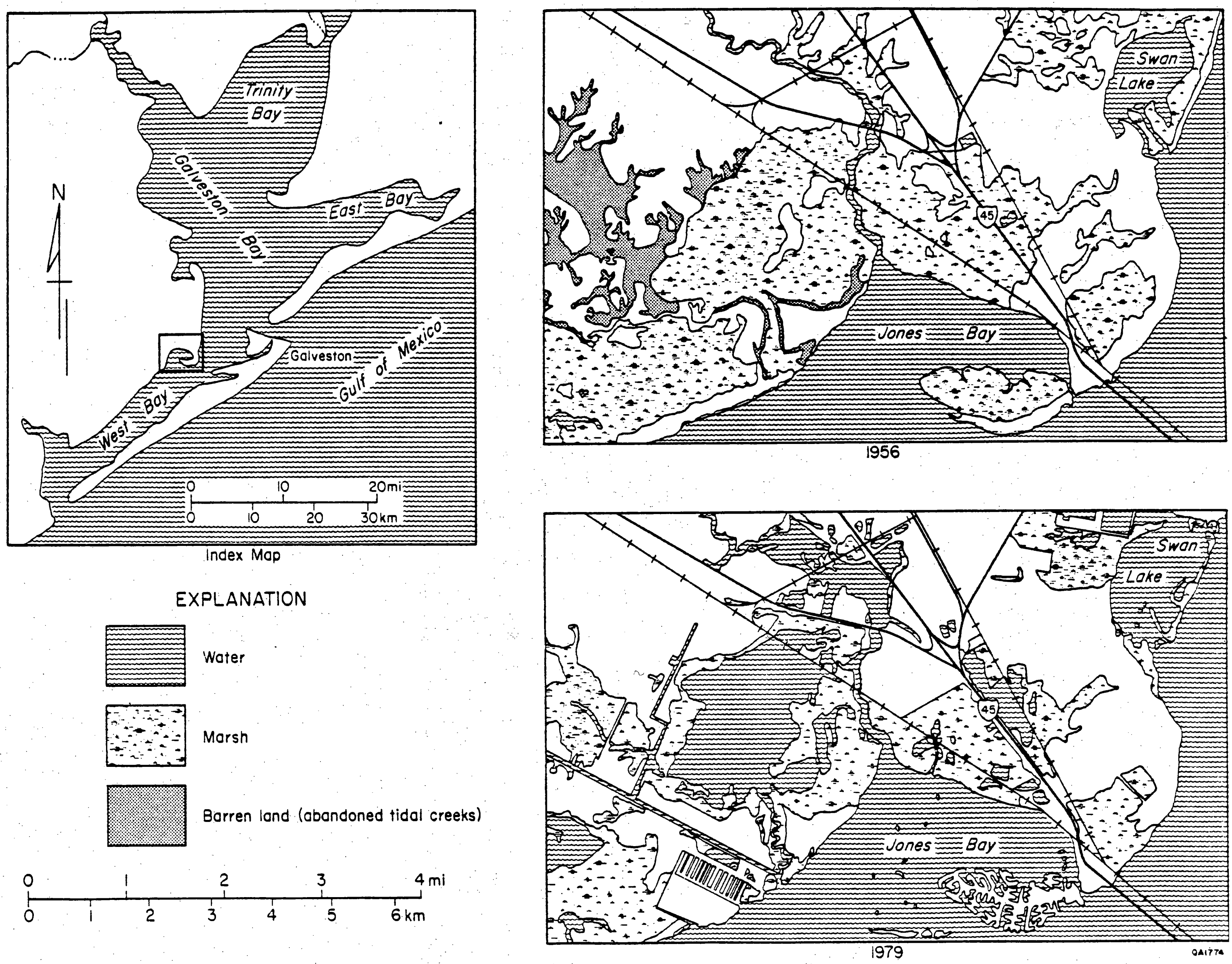


Figure 12. Changes in the distribution of wetlands near Jones Bay and Swan Lake near Galveston. Note increase in open water in 1979, apparently as a result of subsidence. (From White and others, 1985.)

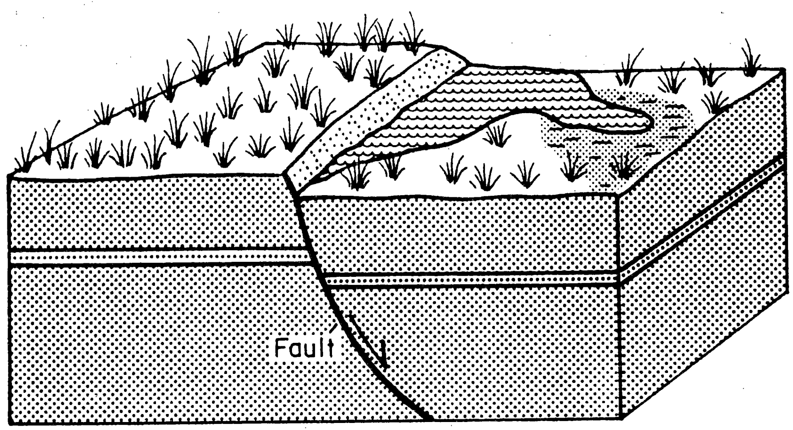
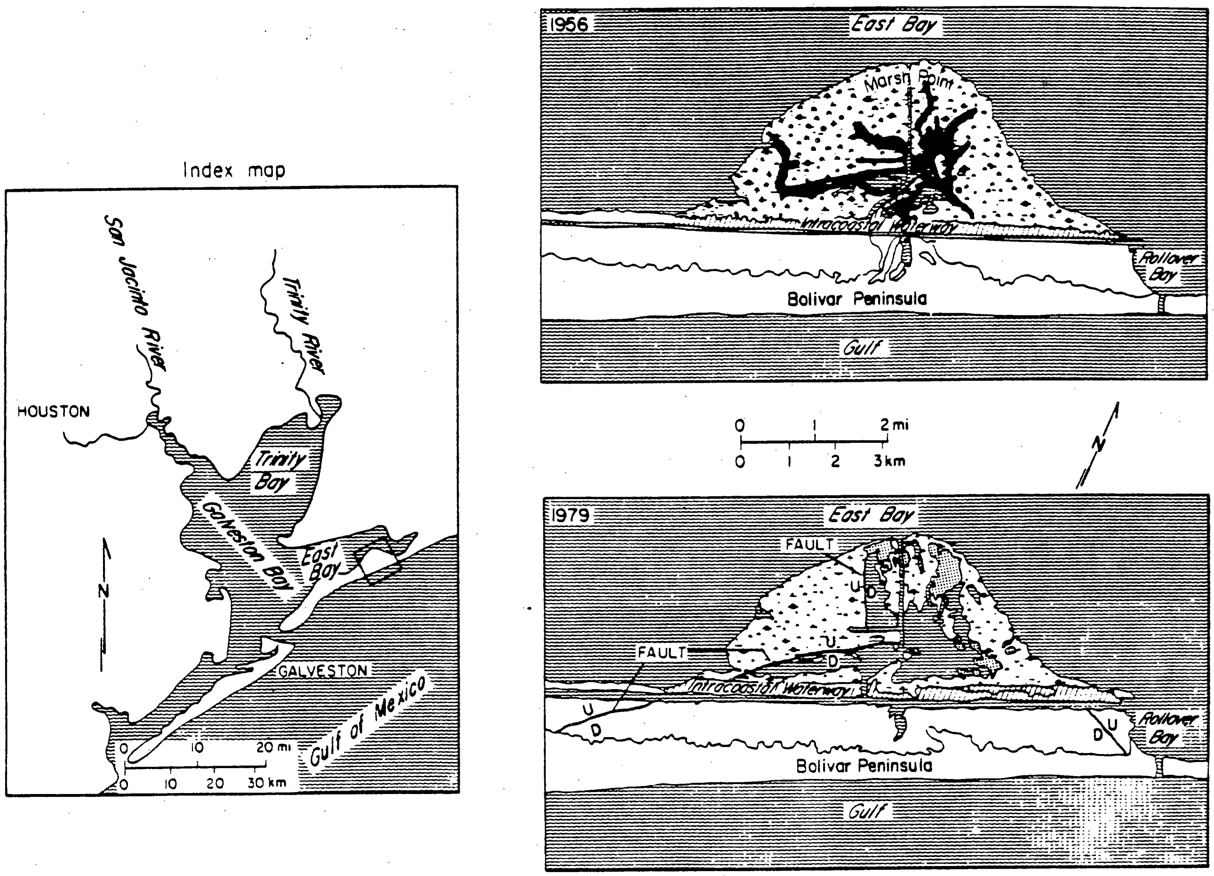


Figure 13. Block diagram of changes in wetlands that can occur along an active surface fault. There is generally an increase in low marshes, shallow subaqueous flats, and open water on the downthrown side of the fault relative to the upthrown side.



Wetland Map Unit	1956		1979		Net Change	
	Acres	Hectares	Acres	Hectares	Acres	Hectares
Water	140	57	1130	458	+990	+401
Marsh	3480	1409	2560	1037		
Marsh-covered tidal channels	640	259				
TOTAL MARSH	4120	1668	2560	1037	-1560	-631
Barren tidal flats			270	109	+270	+109
Dredged spoil/uplands	520	211	720	292	+200	+81
TOTAL	4780	1936	4680	1896	-100*	-40*

*Most of the loss in total area is probably due to erosion of marsh along the margins of East Bay. Historical monitoring indicates that this shoreline is undergoing erosion at rates of up to 3ft (1 m) per year (Paine and Morton, 1986)

(Areas were calculated using a square-count method; smallest squares used were equivalent to 6.4 acres, or 2.6 hectares).

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Figure 14. Changes in the distribution of wetlands affected by surface faults. Increases in the areal extent of water in 1979 relative to 1956 are apparently related to localized subsidence and active faults (D = downthrown side of fault, U = upthrown side). (From White and others, 1985.)

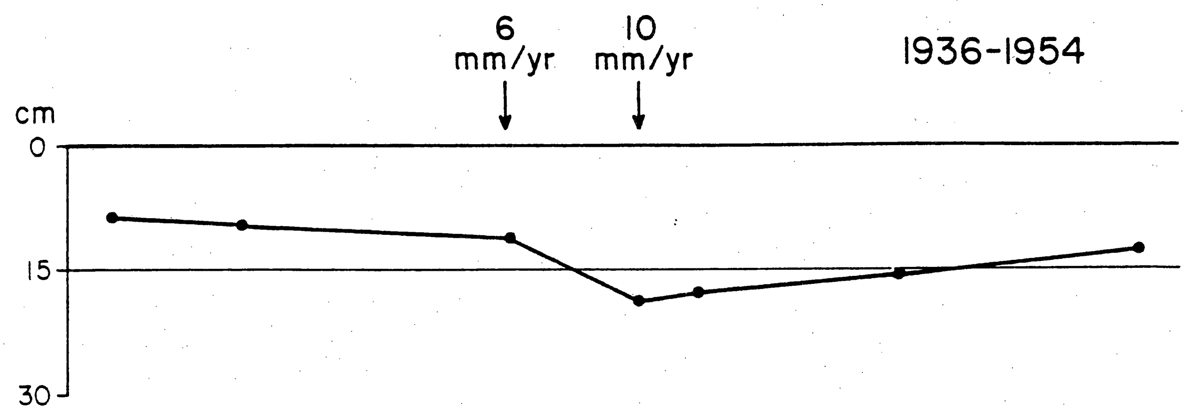


Figure 15. Land subsidence profile based on benchmark-releveling data along State Highway 87 on Bolivar Peninsula. The increase in subsidence along the profile indicates that it crosses an active fault, probably an extension of the fault with the NE-SW strike in figure 14. (Profile from C. W. Kreitler, unpublished data.)

on aerial photographs. One fault has affected wetlands that have developed on modern fluvial-deltaic deposits along the lower Neches River valley at the head of Sabine Lake (White and others, 1987); this area is discussed in more detail in a later section of this report.

Characteristics of Major River Systems Discharging into Coastal Basins

Major coastal rivers and statewide drainage basins are shown in Figures 1 and 16, respectively. Slopes of the Cenozoic Gulf Coastal Plain are relatively steep across the San Marcos Arch but more gentle across the Houston and Rio Grande Embayments (Fig. 17). The gradients of rivers reflect the slopes of these different tectonic provinces (Morton and Donaldson, 1978). Rivers such as the Nueces and Guadalupe on the central Texas coast were affected by uplift along the San Marcos Arch and have steeper gradients than rivers crossing the Houston Embayment (Sabine, Neches, Trinity, and Brazos Rivers) and Rio Grande Embayment (Rio Grande River) (Fig. 18).

Winker (1979) reported that modern rivers that cross the Texas coastal plain can be characterized in terms of drainage basin, discharge, and sediment load; these parameters allow calculation of ratios that further define the nature of the river systems (Fig. 19). As indicated by Winker, the dominance of the Rio Grande, Colorado, and Brazos over other Texas rivers is clearly reflected in drainage basin area, average annual discharge, and sediment load; the relatively systematic decline in runoff depths toward the southwest reflects the climatic gradient (Fig. 6). The fact that average denudation rates show a more complex pattern than runoff depths may be the result of human modification of the drainage basins (Winker, 1979; the data he used was pre-1960).

As stated by Milliman and Meade (1983), two basic methods are used to estimate the amount of sediments transported by rivers to the oceans (and estuaries): one method estimates the denudation of the land (as illustrated in the bottom graph in Fig. 19), and the other estimates the mass carried by the rivers. Of these two, the denudation method yields a much larger estimate of sediment load because it includes a large amount of sediment that never reaches the oceans (Milliman and Meade, 1983). Factors controlling denudation, a term commonly used as a synonym

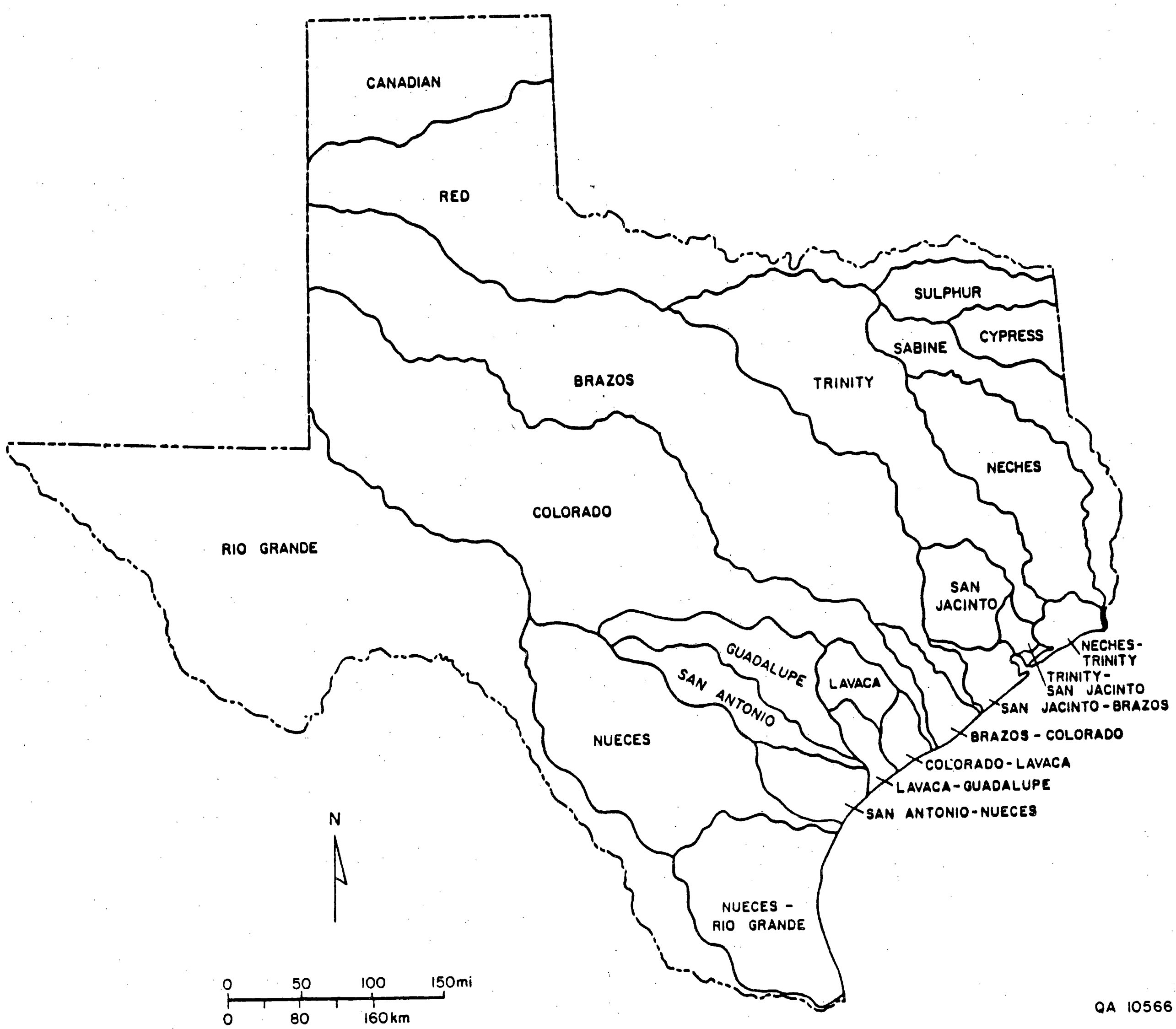


Figure 16. River and coastal basins in Texas. (From Texas Water Development Board.)

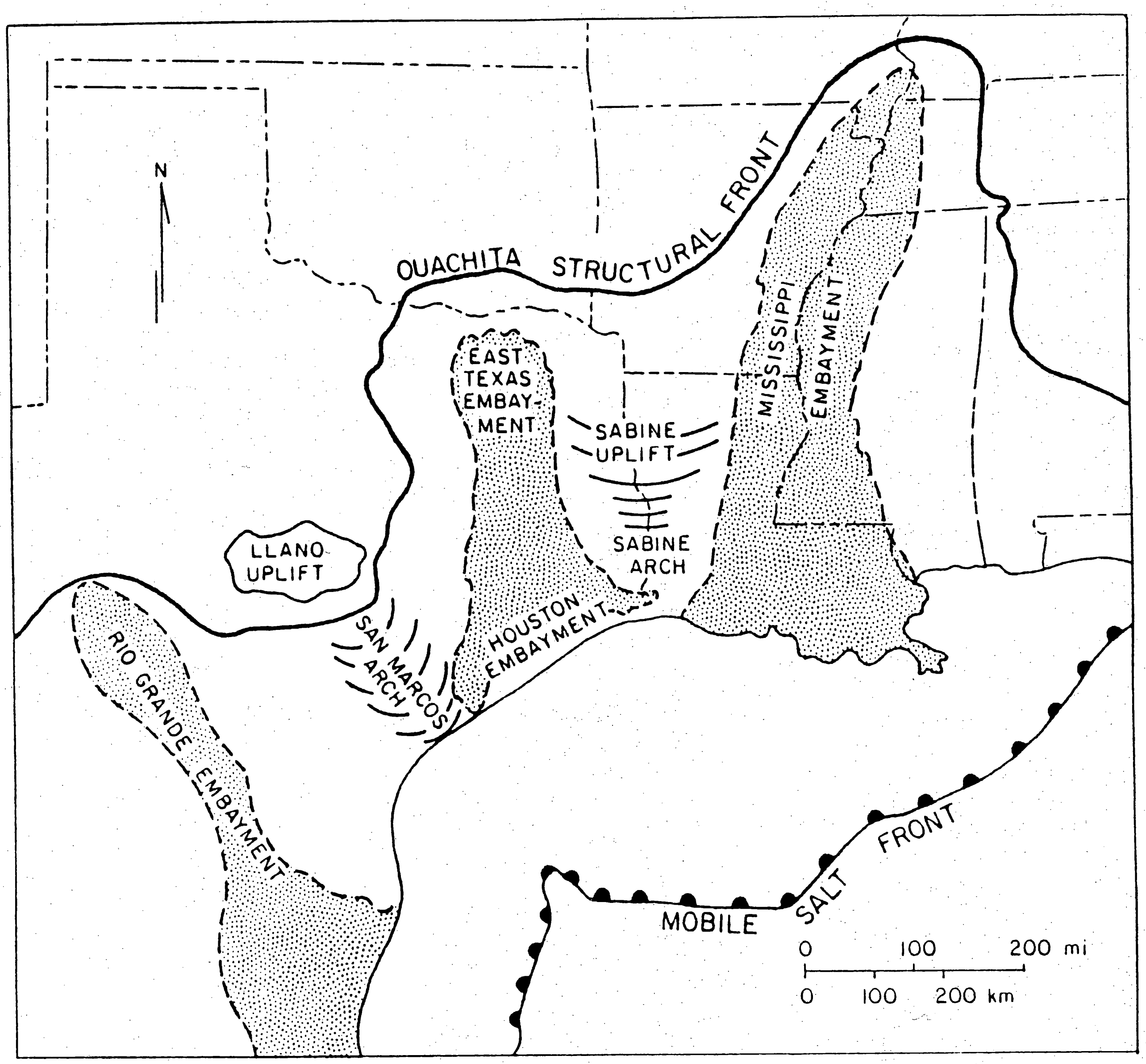


Figure 17. Structural elements that affect Texas coastal rivers and basins. (From Wright, 1980, after Murray, 1961, and Hardin, 1962.)

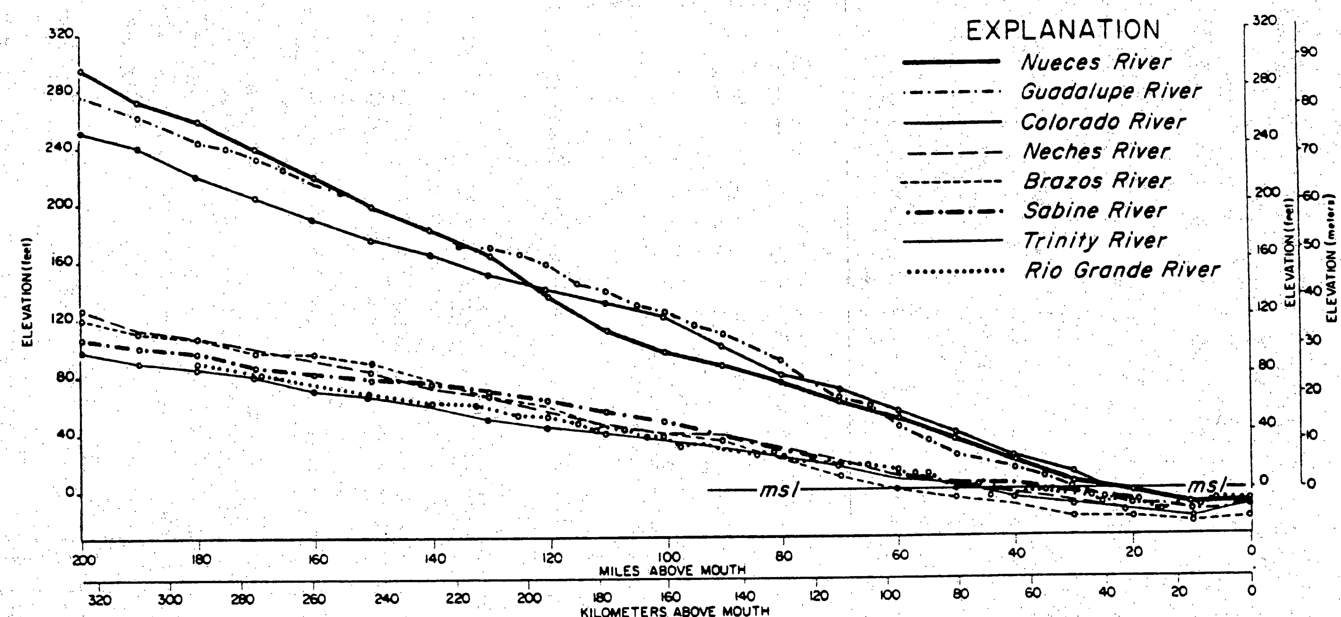


Figure 18. Longitudinal channel profiles for major Texas rivers. (From Morton and Donaldson, 1978.)

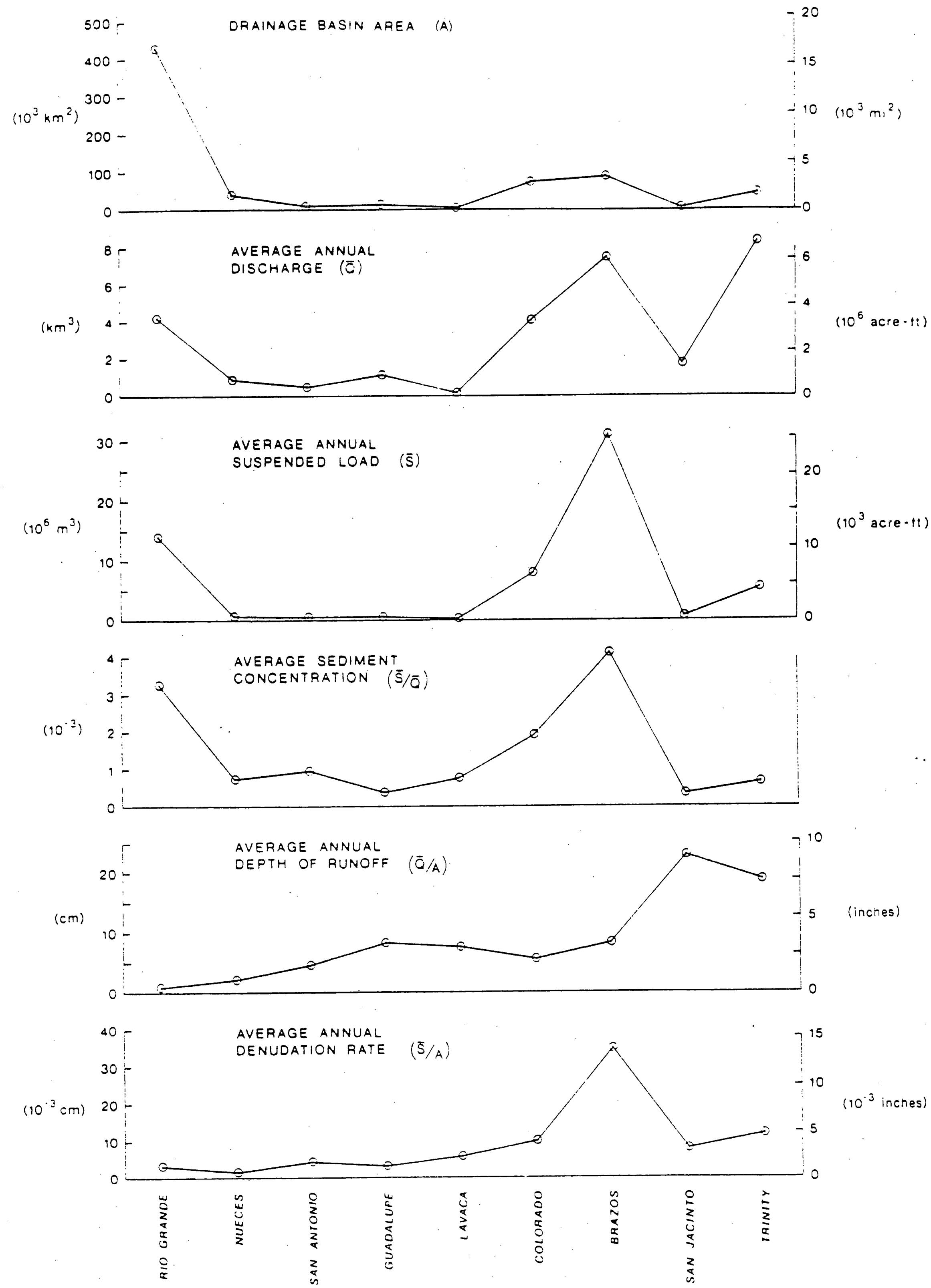


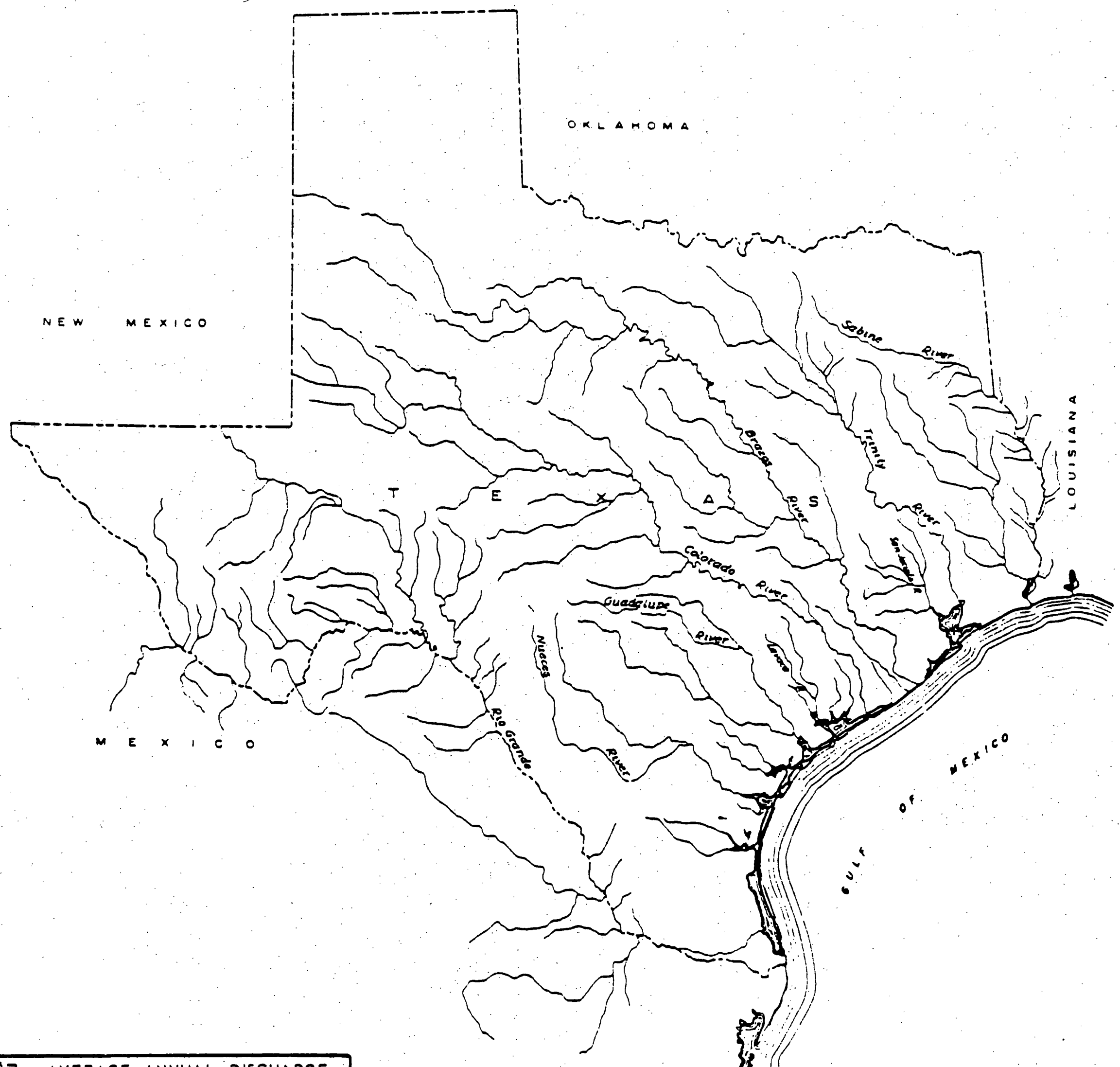
Figure 19. Characteristics of some major coastal rivers in Texas. (From Winker, 1979; average annual discharge and suspended sediment load based on stream-gauge records near the coast, Stout and others, 1961; years of record range from 1925 to 1952, and were chosen, where possible, to avoid the influence of reservoirs upstream.)

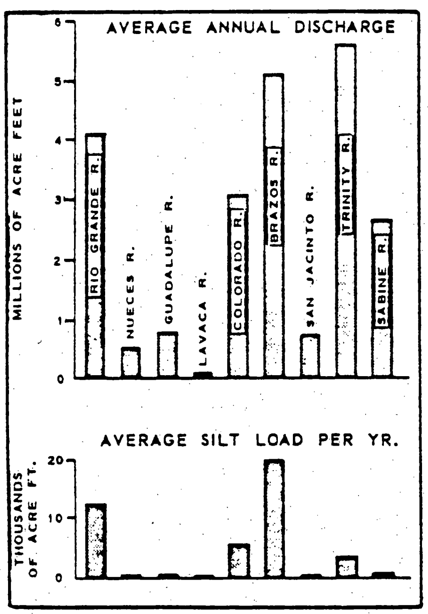
for erosion, include (1) size of drainage basin, (2) precipitation and vegetation, (3) elevation and relief, (4) rock types, and (5) man (Ritter, 1967).

The Soil Conservation Service, U.S. Department of Agriculture, has estimated the amount of sediment eroded from Texas land areas (based on general land use and soils maps) using the universal soil loss equation (Greiner, 1982). The universal soil loss equation uses factors related to those listed above for denudation including rainfall, soil erodibility, topography, crop management, and erosion control. In addition, the Soil Conservation Service has investigated sedimentation by water in Texas (USDA, 1959; Greiner, 1982). A comparison of previous sedimentation surveys (median date of which is 1947) with a study in 1979 for various lakes and reservoirs in Texas indicated that rates of sediment accumulation had declined (Greiner, 1982). Greiner attributed the lower rates in sedimentation to several factors, including changes in land use, by noting there have been (1) significant decreases in amount of cropland (a large producer of sediment), (2) continuous implementation of soil conservation measures since 1935, and (3) construction of flood-prevention dams and other trapping elements since about 1954.

Historical Discharge and Load

A comparison of discharge and sediment load of the major Texas rivers, based on early records (up to 1954), shows that the Trinity, Brazos, Colorado, and Rio Grande had (in the past) the highest average annual discharge, ranging from about 2.7 million acre-ft to slightly more than 5 million acre-ft; these same rivers also had the highest annual silt loads, ranging from the Brazos with 20,148 acre-ft to the Trinity (3,622 acre-ft) (Fig. 20). The Trinity River load is only about 18 percent of that of the Rio Grande. The high sediment loads characterizing the Rio Grande and the ancestral Brazos–Colorado couplet contributed to the filling of their respective paleovalleys (see preceding section on origin of bays), and to the progradation of deltas into the Gulf. Suspended sediment load transported by the other streams was considerably lower, ranging from 88 acre-ft/yr (Lavaca River) to 636 acre-ft (Sabine River). None of these rivers, including the Trinity, have completely





SUMMARY OF DATA ON DISCHARGE AND SILT LOAD
FOR SOME MAJOR TEXAS STREAMS

Stream	Period	Annual Average Discharge of Stream Acre-feet	Annual Avg. Amt. of Silt Acre-feet	Tons	Net Drainage Area Sq. Miles
Brazos	1924-54	5,186,640	20,148	30,756,580	34,810
Colorado	1937-42	3,167,710	5,898	8,991,960	29,140
Guadalupe	1945-54	799,662	303	461,214	5,311
Lavaca	1945-54	122,837	88	133,945	887
Nueces	1942-54	555,636	116	177,690	–
Rio Grande	1929-43	4,166,619	12,588	19,192,311	157,204
Sabine	1932-33 35-54	2,762,345	636	970,766	4,858
San Antonio	1942-54	395,231	373	568,218	3,918
San Jacinto	1932-33; 37-54	703,528	213	325,280	1,811
Trinity	1936-54	5,689.331	3,622	5,520,960	17,192

Figure 20. Annual discharge, silt load, and drainage area for some major Texas rivers. Based on data prior to 1955. (From LeBlanc and Hodgson, 1959.)

filled their paleovalleys (estuaries), but all have constructed deltas at the head of the bays (bayhead deltas).

Historical trends in streamflow and sediment loads of major Texas coastal rivers are shown in Figures 21 to 32. Major reductions in sediment load are apparent for the Trinity, Brazos, Colorado, Nueces, and Rio Grande. For example, for the Brazos River, Curtis and others (1973) reported the annual sediment load to be about 16 million tons; as noted by Milliman and Meade (1983), this load is one-half of that (32 million tons) presented in an earlier report by Holeman (1968). The annual average sediment load of the Brazos River presented by Winker (1979) was 31 million tons (Fig. 19) for the period of 1925 to 1947 (Richmond station). The average annual load for the Brazos from 1947 to 1979 was approximately 12 million tons/yr (Fig. 23), or less than 40 percent of that for the earlier period. Although reductions in river sediment load can be partly attributed to land use changes and to the continuous implementation of soil conservation measures in the drainage basins since 1935 (Greiner, 1982), the apparent major contributing factor to the decreased sediment supply in many of the streams in Texas is reservoir development (Fig. 33).

Effect of Reservoir Development on River Sediment Load

The reduction of stream sediment load downstream from reservoirs is well documented. One of the most often cited examples is the Colorado River that discharges into the Gulf of California (Milliman and Meade, 1983). The average annual suspended-sediment load of the Colorado River at Yuma, Arizona, for the period 1911–1916 was about 235 million tons; as a result of reservoir development and increased use of water for irrigation, the suspended-sediment load was reduced (for 1965–1967) to an average annual discharge of about 0.153 million tons, or about seven-hundredths of a percent of the past value (Curtis and others, 1973) (Fig. 34). It should be noted that Meade (1969) (in discussing the Atlantic Coastal Plain) concluded that because of the vast amount of sediment contributed to river systems by human activities such as agriculture (estimated to have increased sediment yield tenfold) and urbanization, that even though sediment is partly intercepted

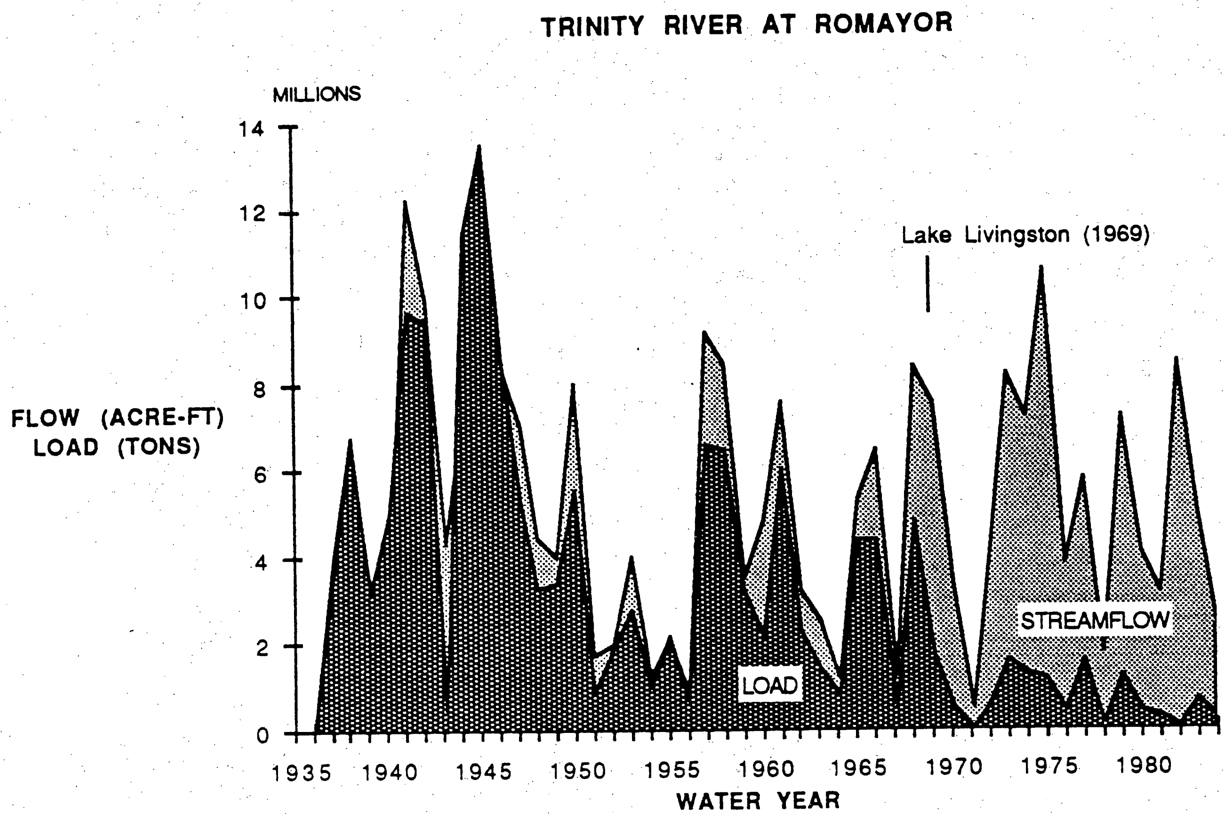


Figure 21. Annual streamflow and suspended sediment load of the Trinity River. (Data from Stout and others, 1961; Adey and Cook, 1964; Cook, 1967; Cook, 1970; Mirabal, 1974; Dougherty, 1979; and unpublished records from Texas Water Development Board, available through TNRIS.)

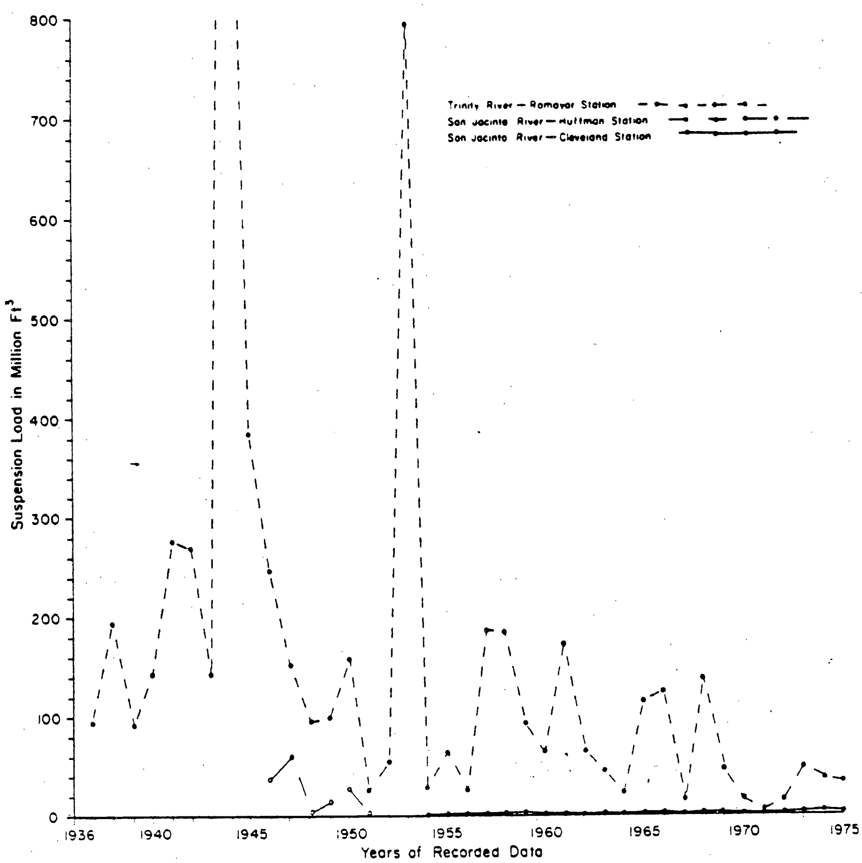


Figure 22. Suspended-sediment load for the Trinity and San Jacinto Rivers for the period 1936 through 1975. (From Morton and McGowen, 1980.)

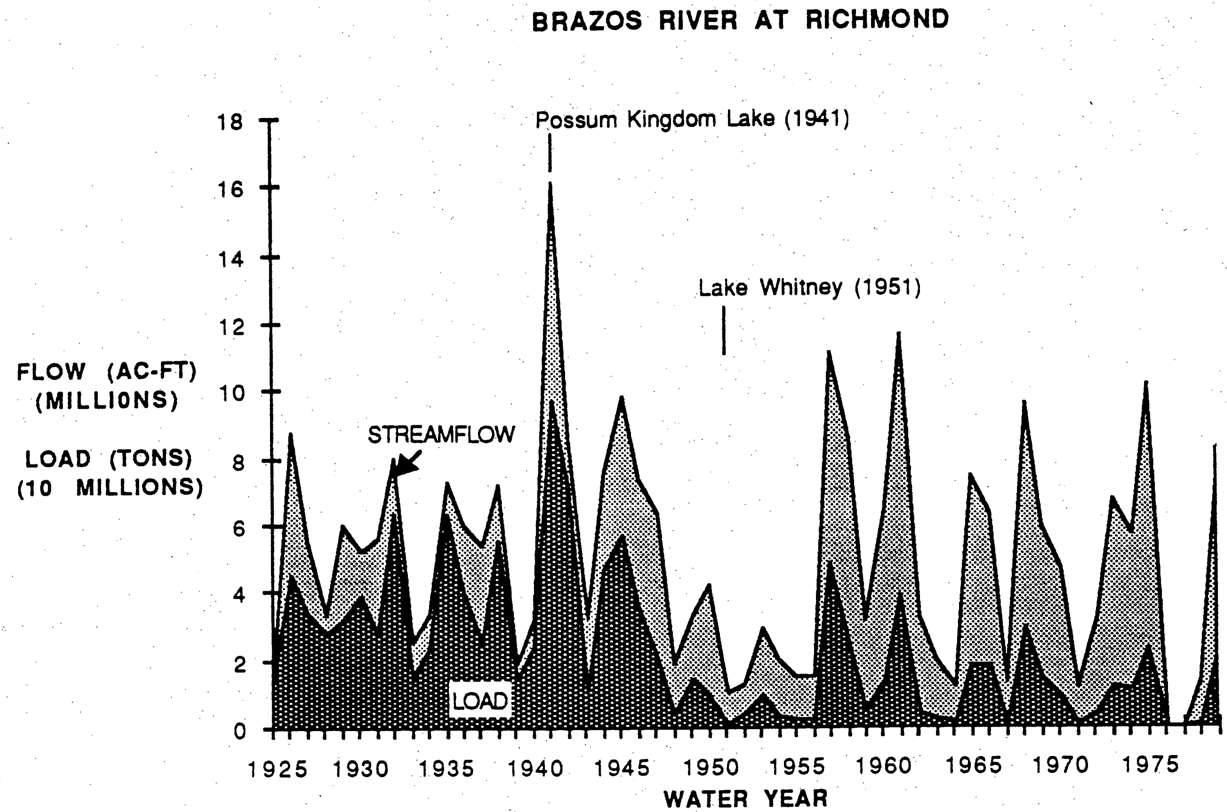


Figure 23. Annual streamflow and suspended load of the Brazos River. (Sources of data same as for figure 21.)

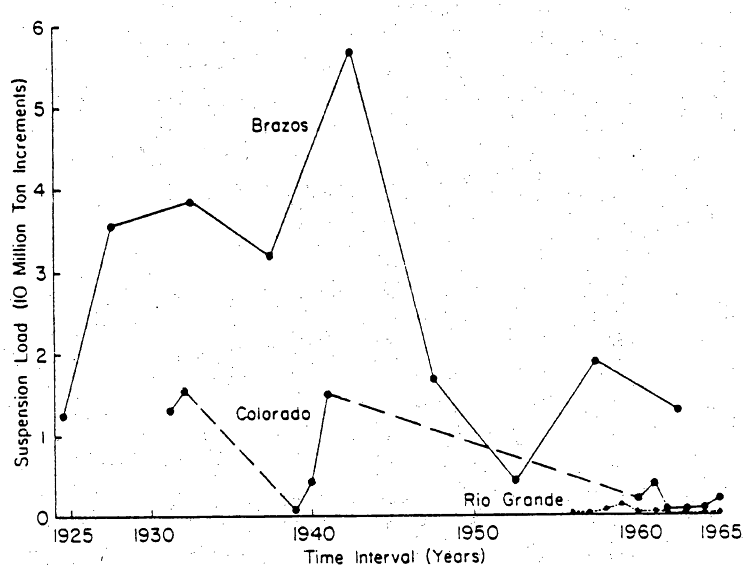


Figure 24. Comparison of suspended-sediment loads of the Brazos and Colorado Rivers, and Rio Grande; gauging stations on the Brazos at Richmond, on the Colorado near Eagle Lake, and on the Rio Grande at Brownsville. Both the Brazos and the Colorado show a decrease in suspended load. The load at Brownsville was monitored only after the completion of Falcon Dam (1954). (From McGowen and others, 1977.)

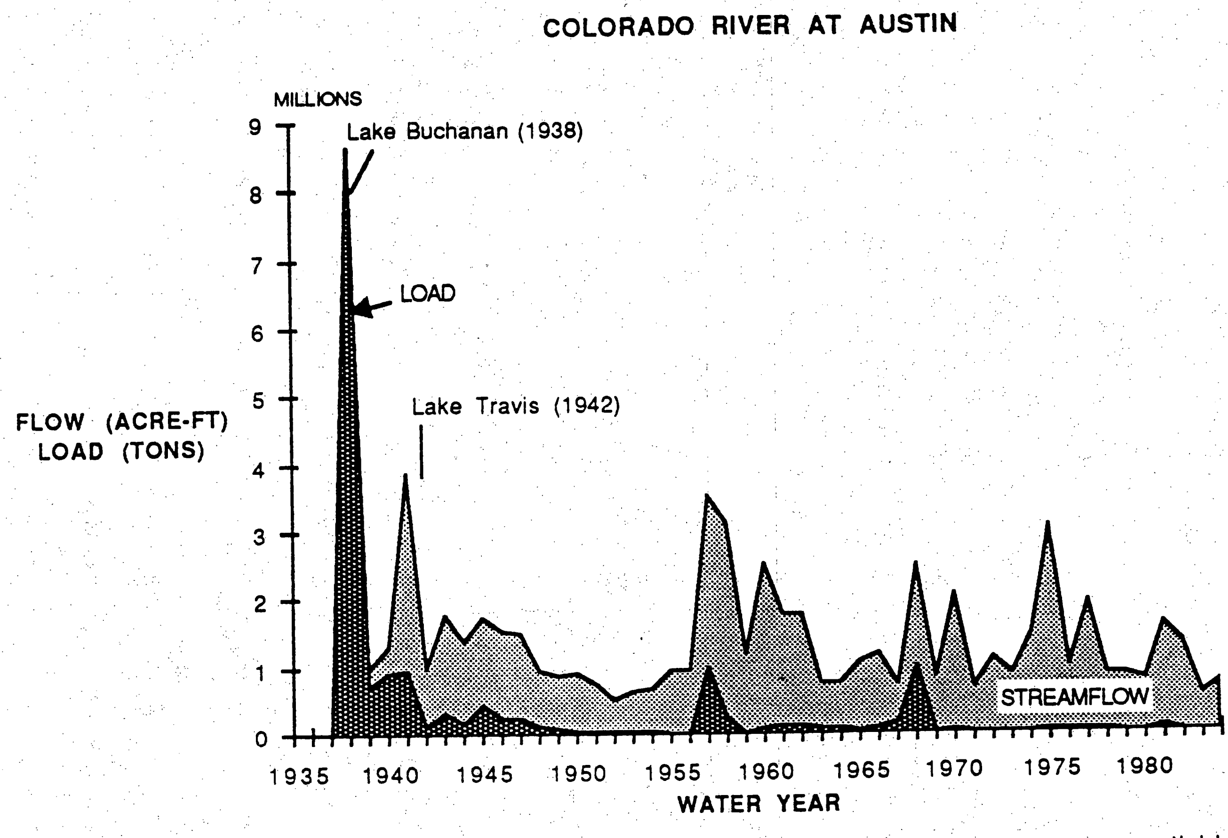


Figure 25. Annual streamflow and suspended load of the Colorado River. Data were not available for the following years: streamflow for 1962 and 1963, and suspended load for 1962, 1963, 1976, and 1977. (Sources of data same as for figure 21.)

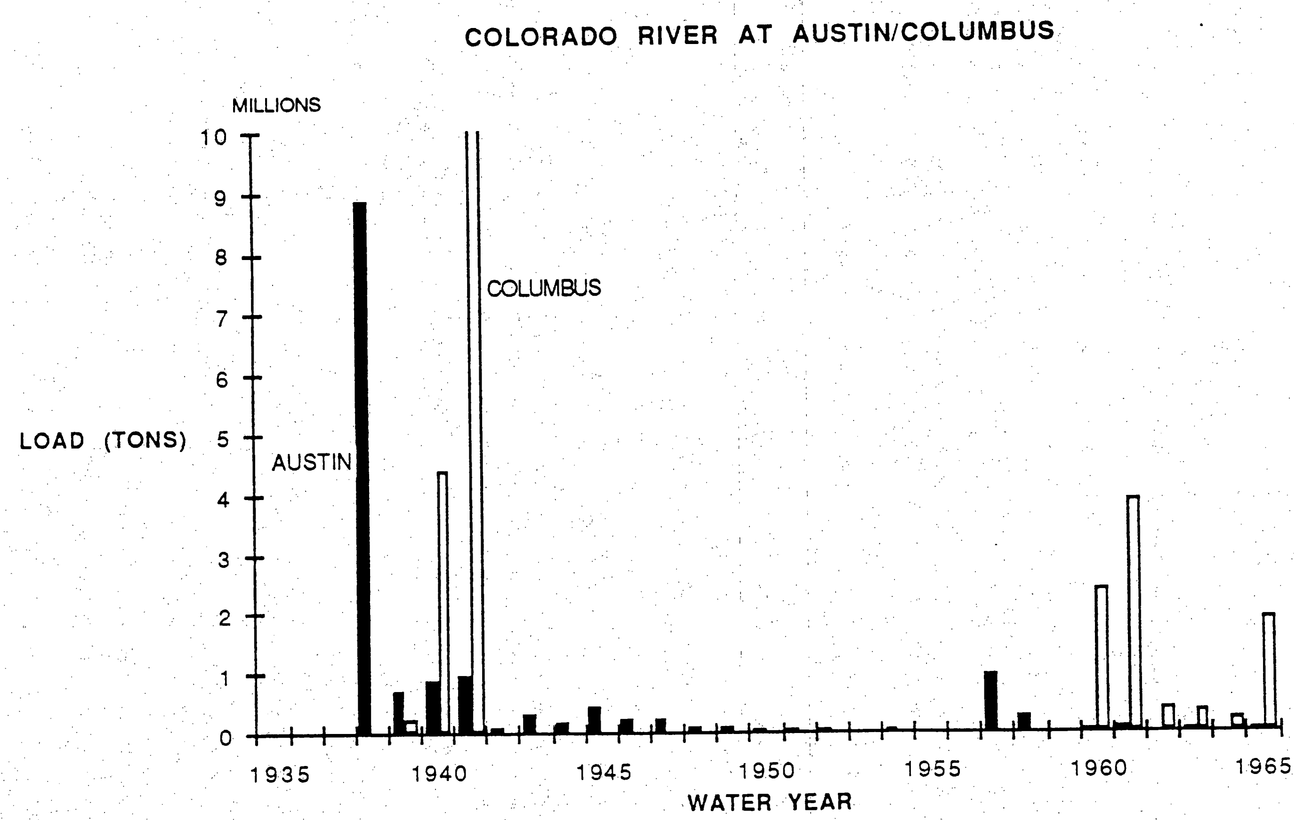


Figure 26. Suspended load of the Colorado River at Austin and Columbus. (Sources of data same as for figure 21.)

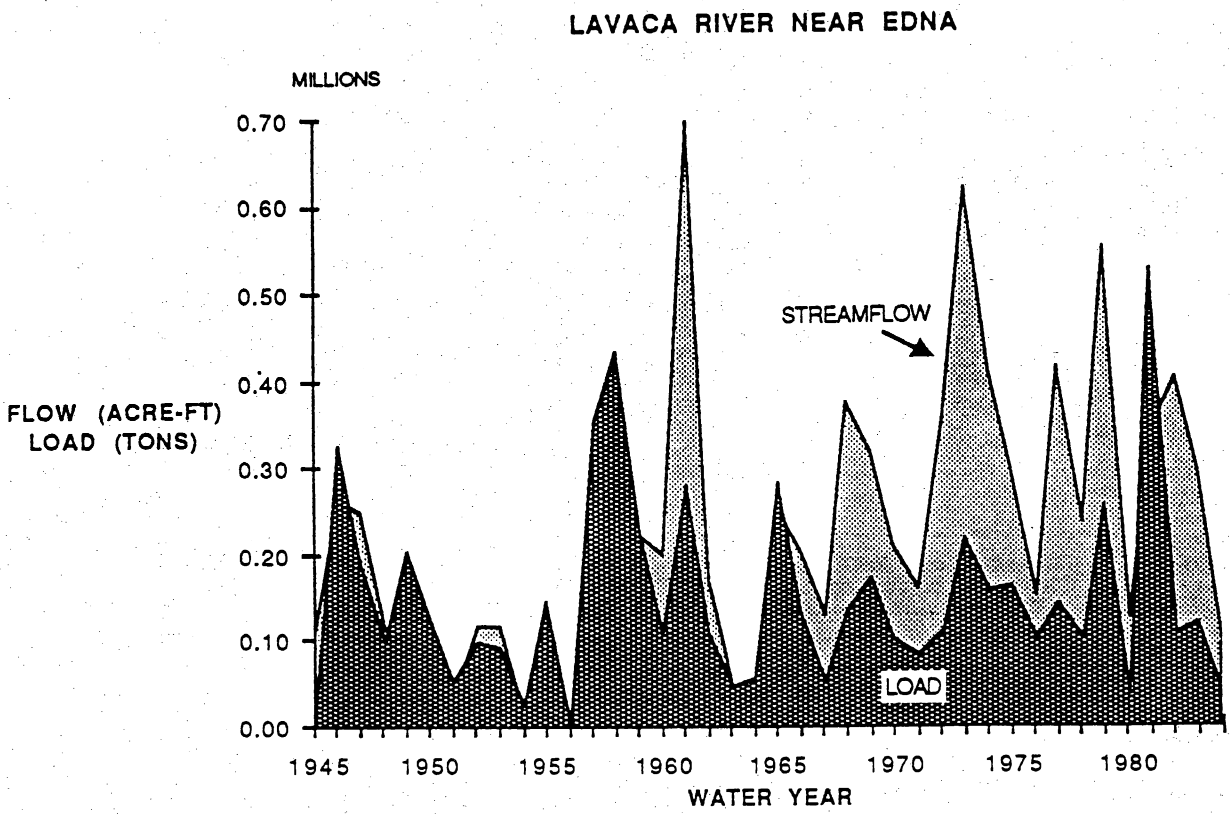


Figure 27. Annual streamflow and suspended load of the Lavaca River. (Sources of data same as for figure 21.)

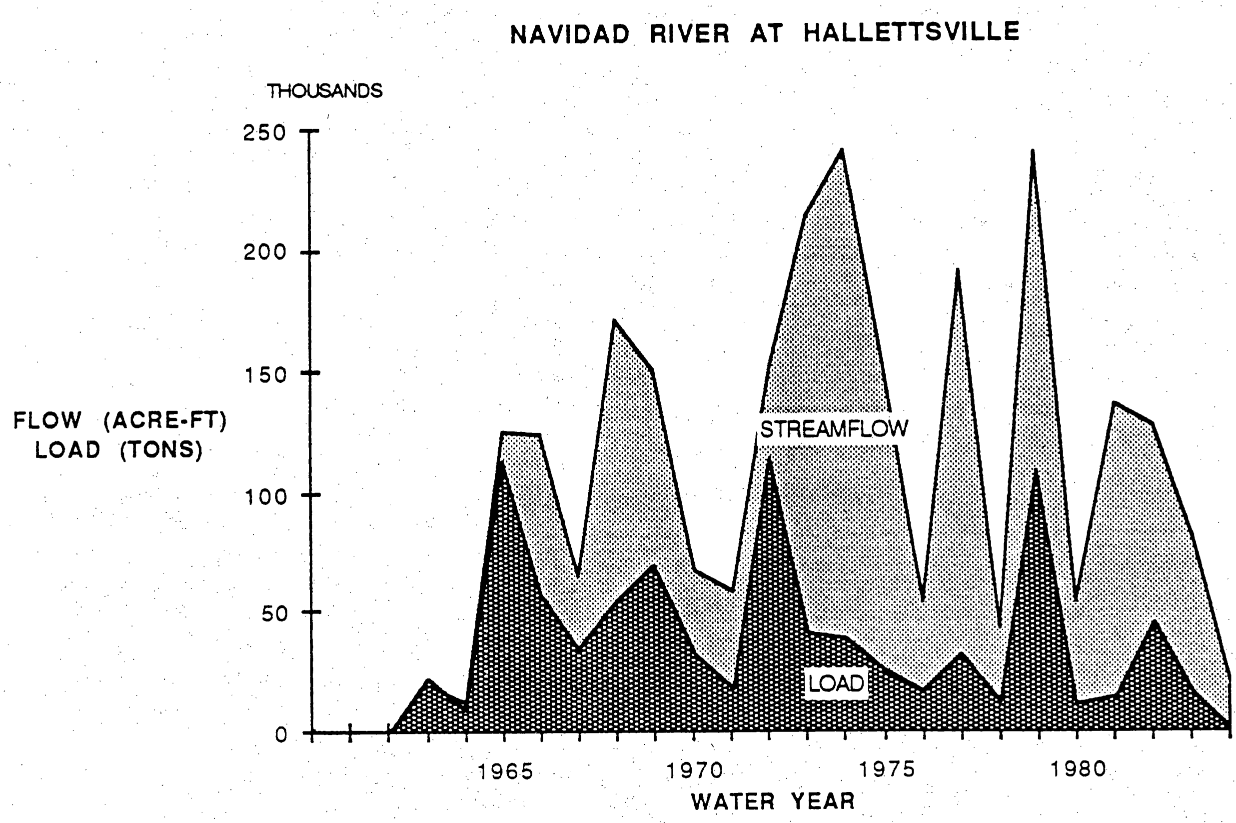


Figure 28. Annual streamflow and suspended load of the Navidad River. (Sources of data same as for figure 21.)

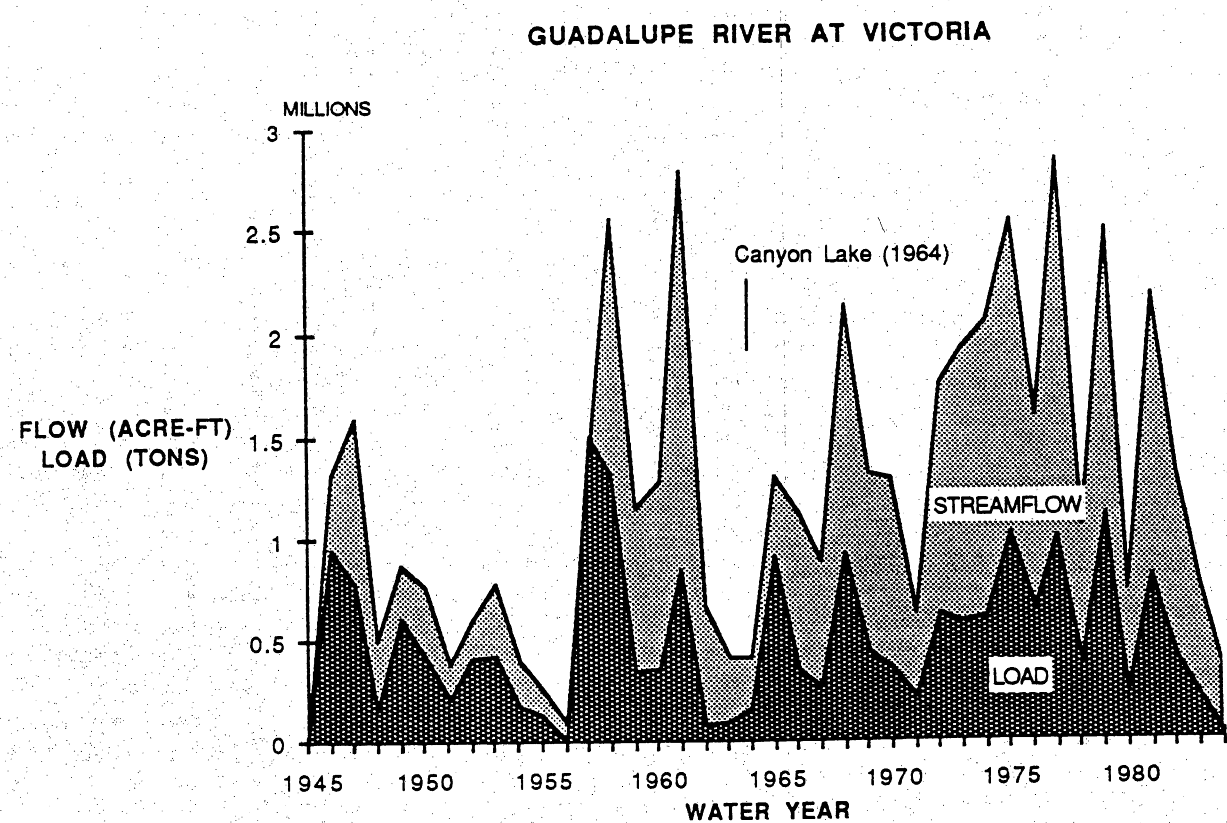


Figure 29. Annual streamflow and suspended load of the Guadalupe River. (Sources of data same as for figure 21.)

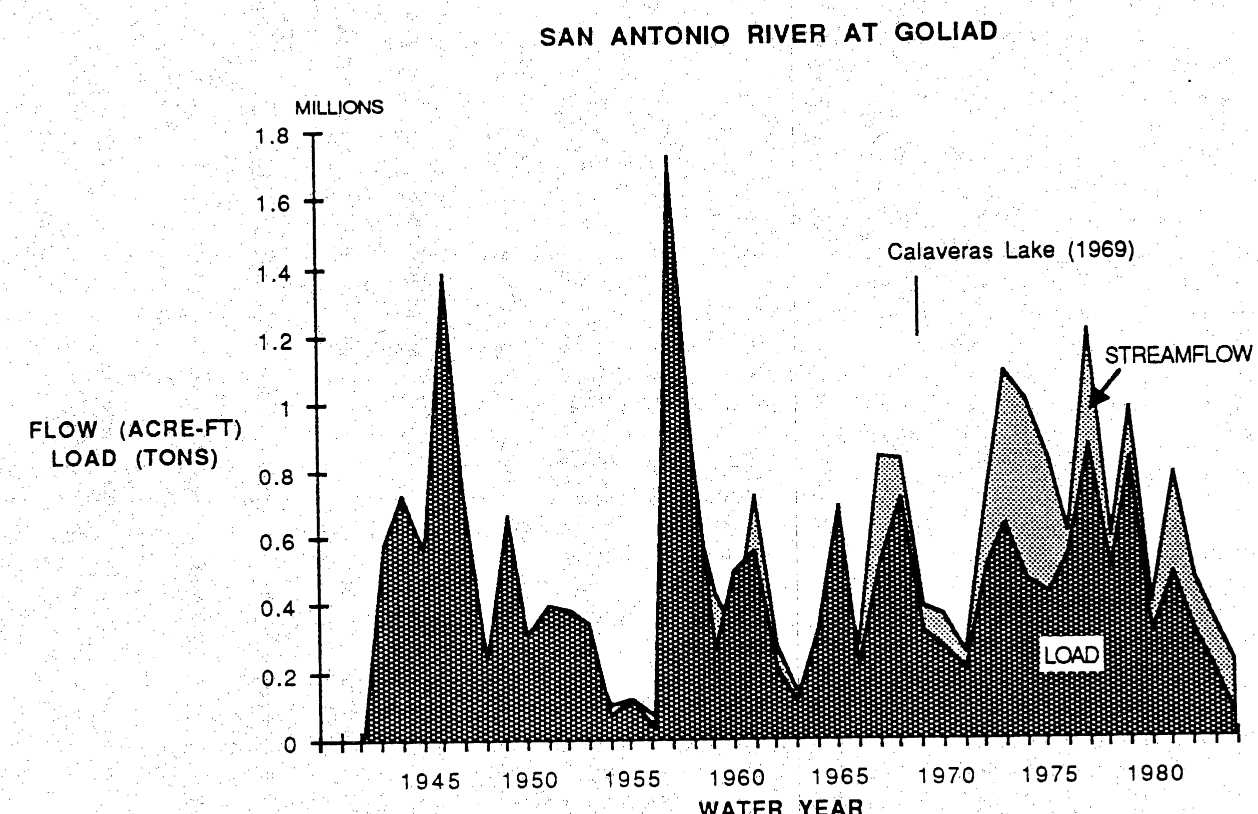


Figure 30. Annual streamflow and suspended load of the San Antonio River. (Sources of data same as for figure 21.)

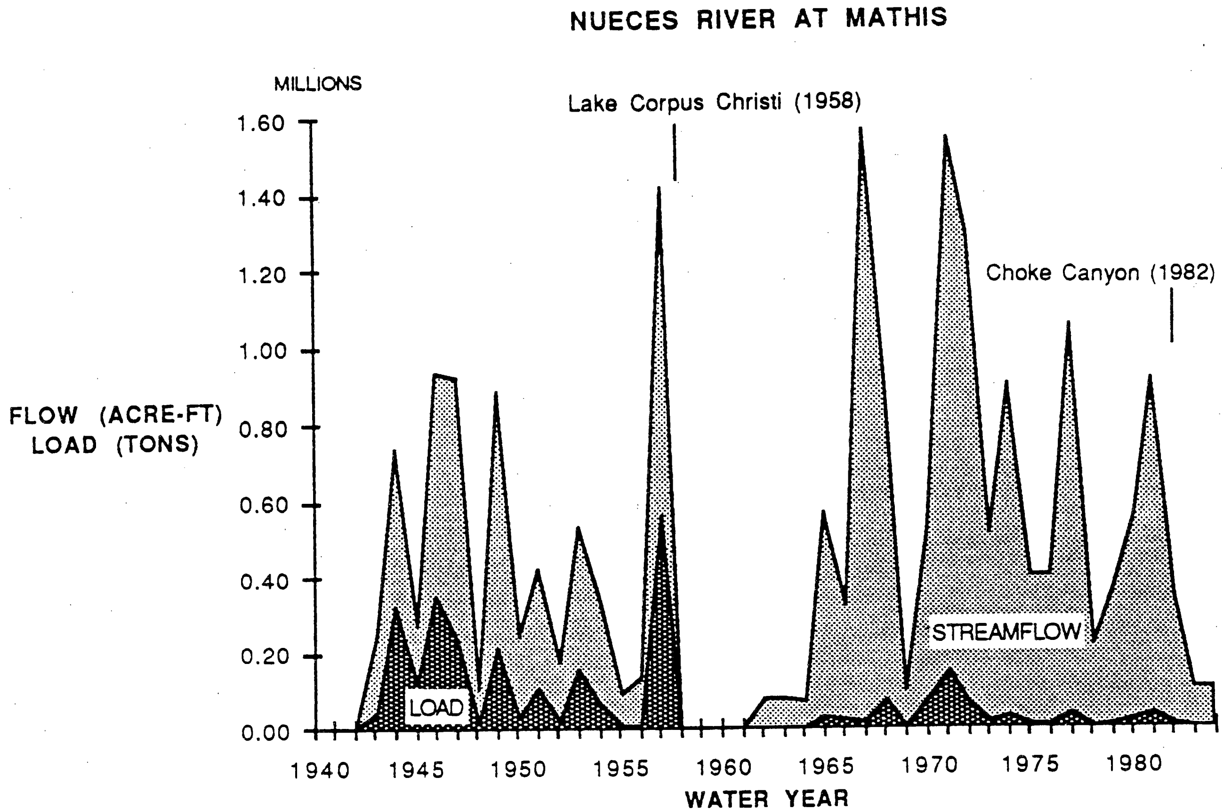


Figure 31. Annual streamflow and suspended load of the Nueces River. Data were not available for the years 1958-1961. (Sources of data same as for figure 21.)

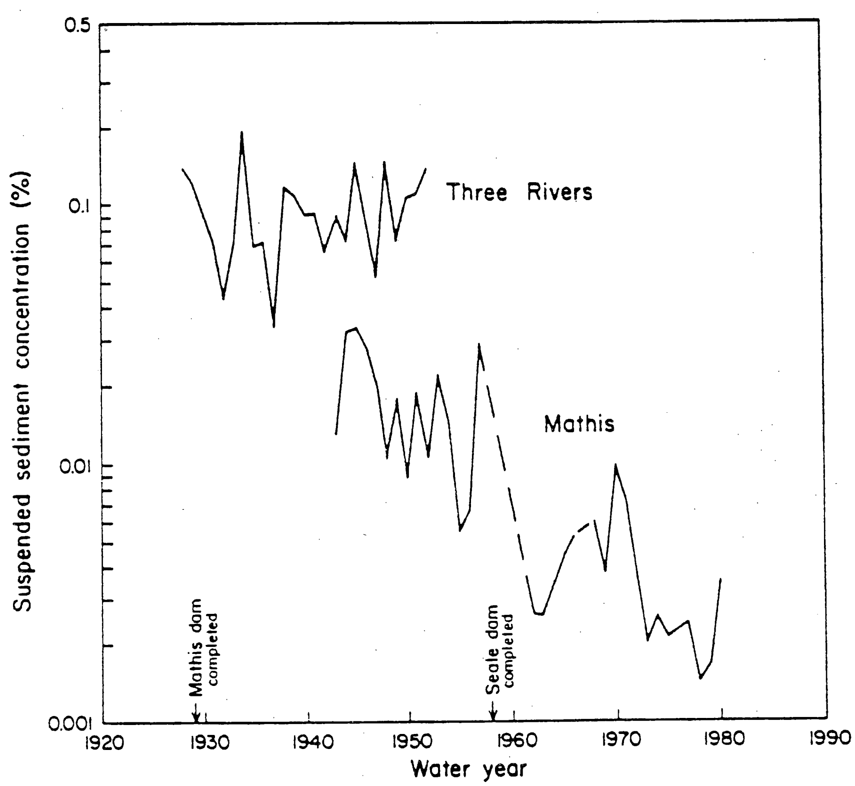


Figure 32. Concentration (by weight) of suspended sediment for the Nueces River. The Three Rivers measuring station is upstream from Lake Corpus Christi, and the Mathis station is downstream. No data for dashed segments. Data from sources given in figure 21. (From Morton and Paine, 1984.)

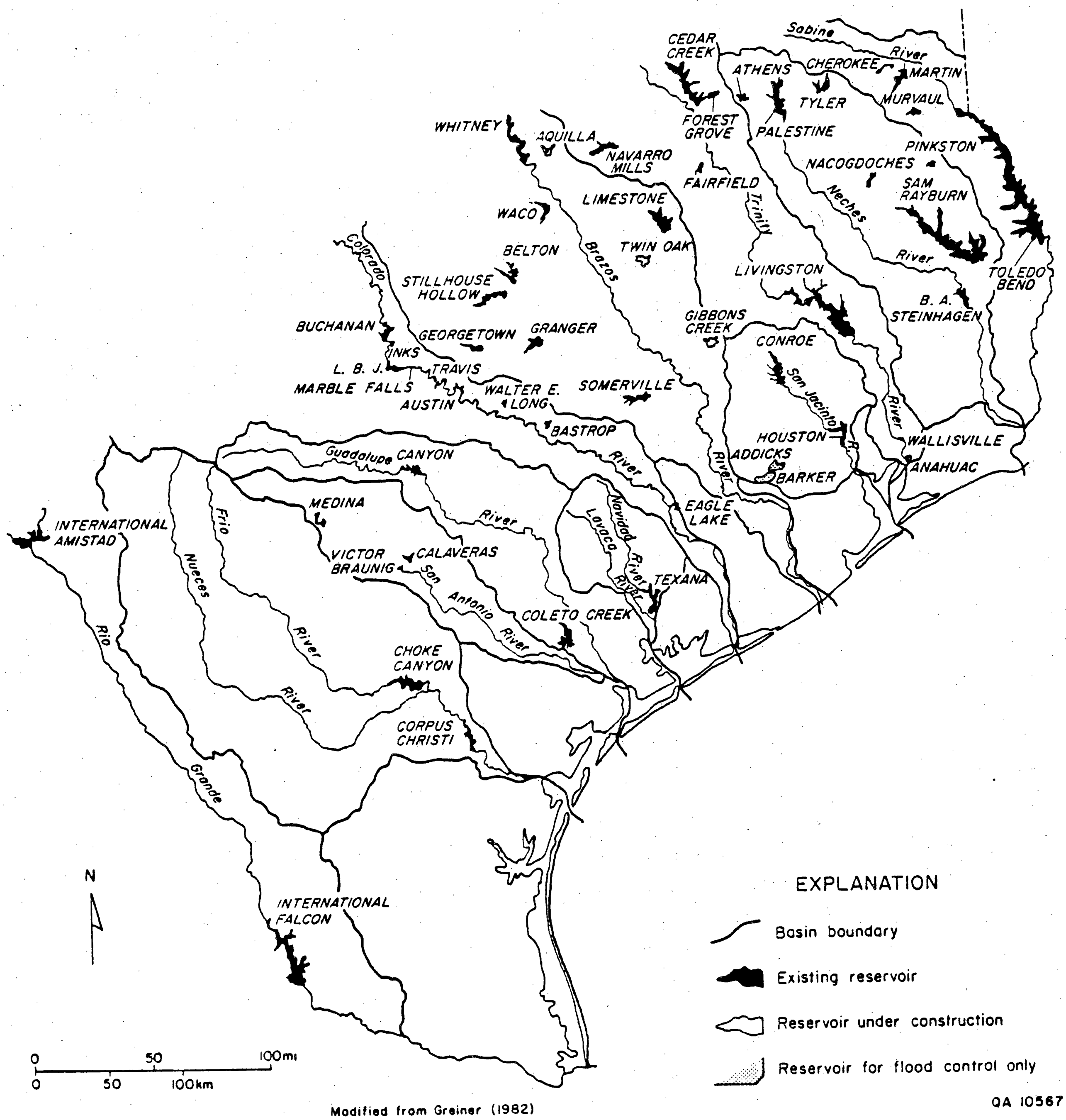


Figure 33. Reservoirs located on the Gulfward half of Texas. (From Greiner, 1982.)

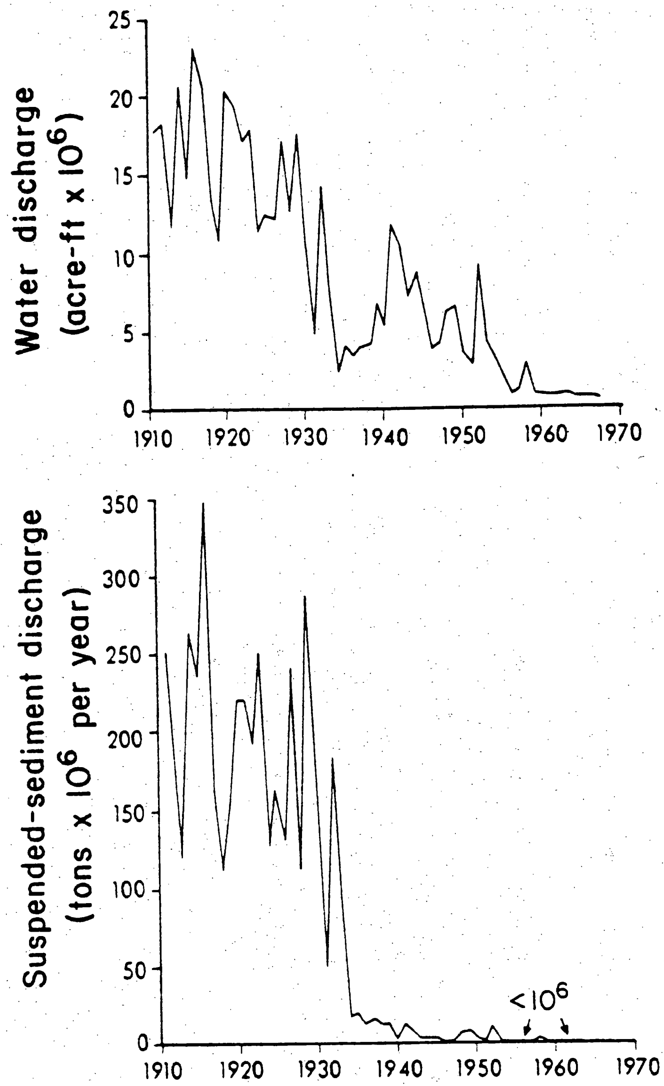


Figure 34. Water discharge and suspended-sediment load for the Colorado River at Yuma, Arizona. The sharp decline in the curves in the mid-1930's is related to reservoir development. (From Curtis and others, 1973.)

by reservoirs on rivers, the sediment loads reaching the Atlantic coast are larger than if the drainage areas were still in a natural state. However, more recently published data (Stevenson and others, 1988) indicate substantial reductions in sediment delivered to nearshore areas along the Atlantic and Gulf coastal plains (see page 94).

Sediment Trapping

Reservoir development in the drainage basins of the Nueces, Trinity, Brazos, and Colorado Rivers in Texas shows significant effects on sediment load and discharge (Figs. 32, 35, and 36). A considerable amount of data has been accumulated on reservoir sedimentation and trapping efficiency in order to determine reservoir life expectancy (Vanoni, 1975; Strand and Pemberton, 1982). The trap efficiency of a reservoir is a measure of the amount of inflowing sediment that is deposited or retained in the reservoir. Large reservoirs (storage capacities of greater than 10,000 acre-ft) trap virtually 95 to 100 percent of the incoming sediment (Leopold and others, 1964; Vanoni, 1975). Williams and Wolman (1984), citing the U. S. Army Corps of Engineers (1960), reported that the trapping efficiency of Denison Dam on the Red River (Texas and Oklahoma) was 99.2 percent during the first 12 yr after closure. The amount retained is primarily controlled by two factors, the average velocity of flow through the reservoir and the sediment size; clay may remain in suspension long enough to pass through the reservoir, but sand will not (Vanoni, 1975). Brune (1953) developed curves for estimating the trapping efficiency of reservoirs by plotting sediment trapped in 44 reservoirs against the ratios of reservoir capacity to annual inflow (Fig. 37). This concept is discussed in a later section with respect to trapping efficiency of estuaries. Churchill (1948), studying reservoirs of the Tennessee Valley Authority, presented a method of calculating reservoir trapping efficiency by relating the percentage of incoming sediment that passes through a reservoir to the reservoir's sediment index, which is the period of retention (reservoir capacity/daily inflow) divided by mean velocity of the flow through the reservoir. Strand and Pemberton (1982) compared curves of the two methods (Fig. 38). Several Texas reservoirs are included in figures 37 and 38. The trap efficiencies of Possum Kingdom on the Brazos River and Buchanan on the

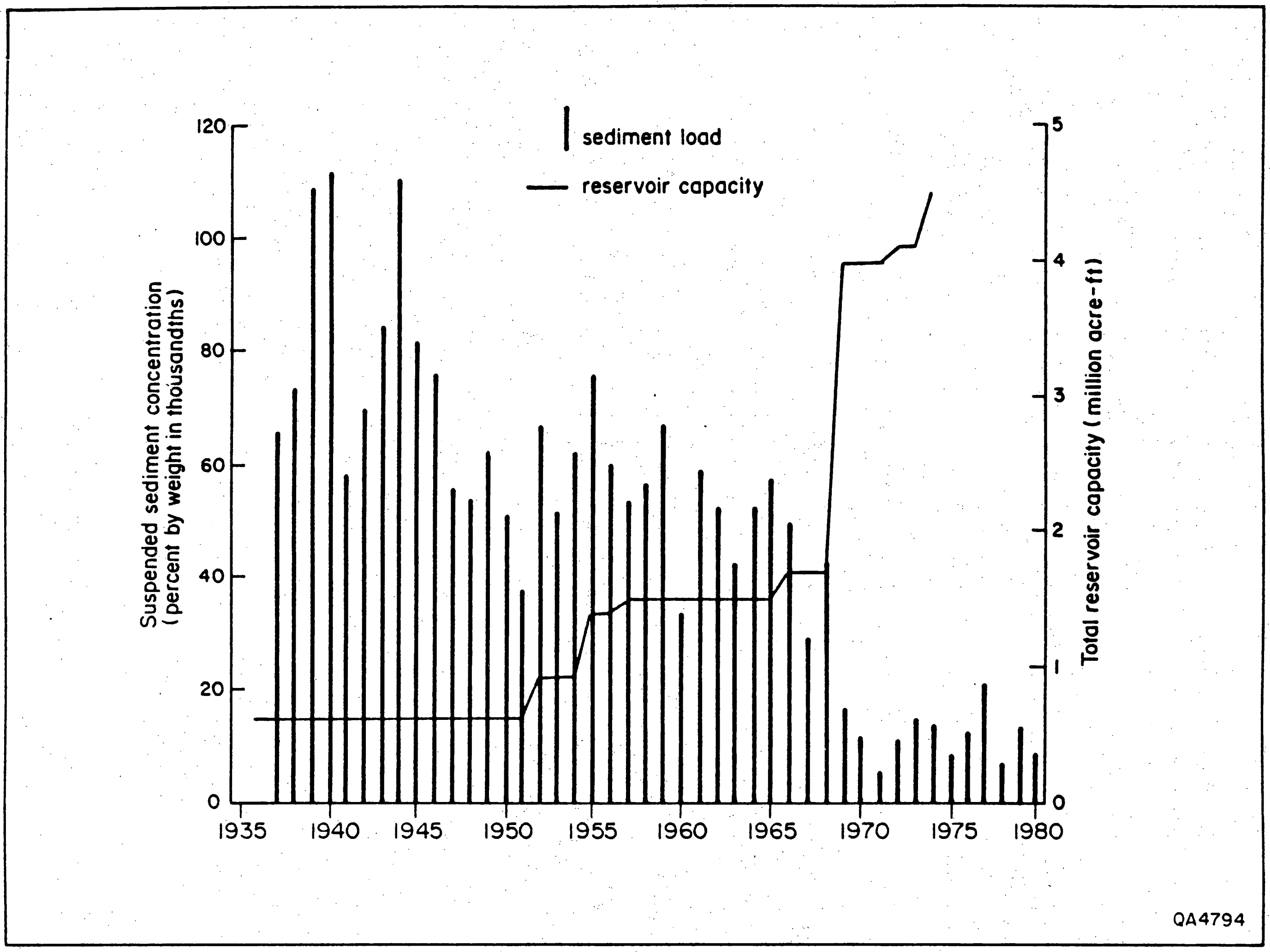


Figure 35. Suspended-sediment load (percent by weight) of the Trinity River at Romayor, and cumulative authorized water storage in reservoirs of the Trinity River basin. (Sources of data for suspended load same as in figure 21; reservoir data from Texas Water Development Board, 1973.) (From Paine and Morton, 1986.)

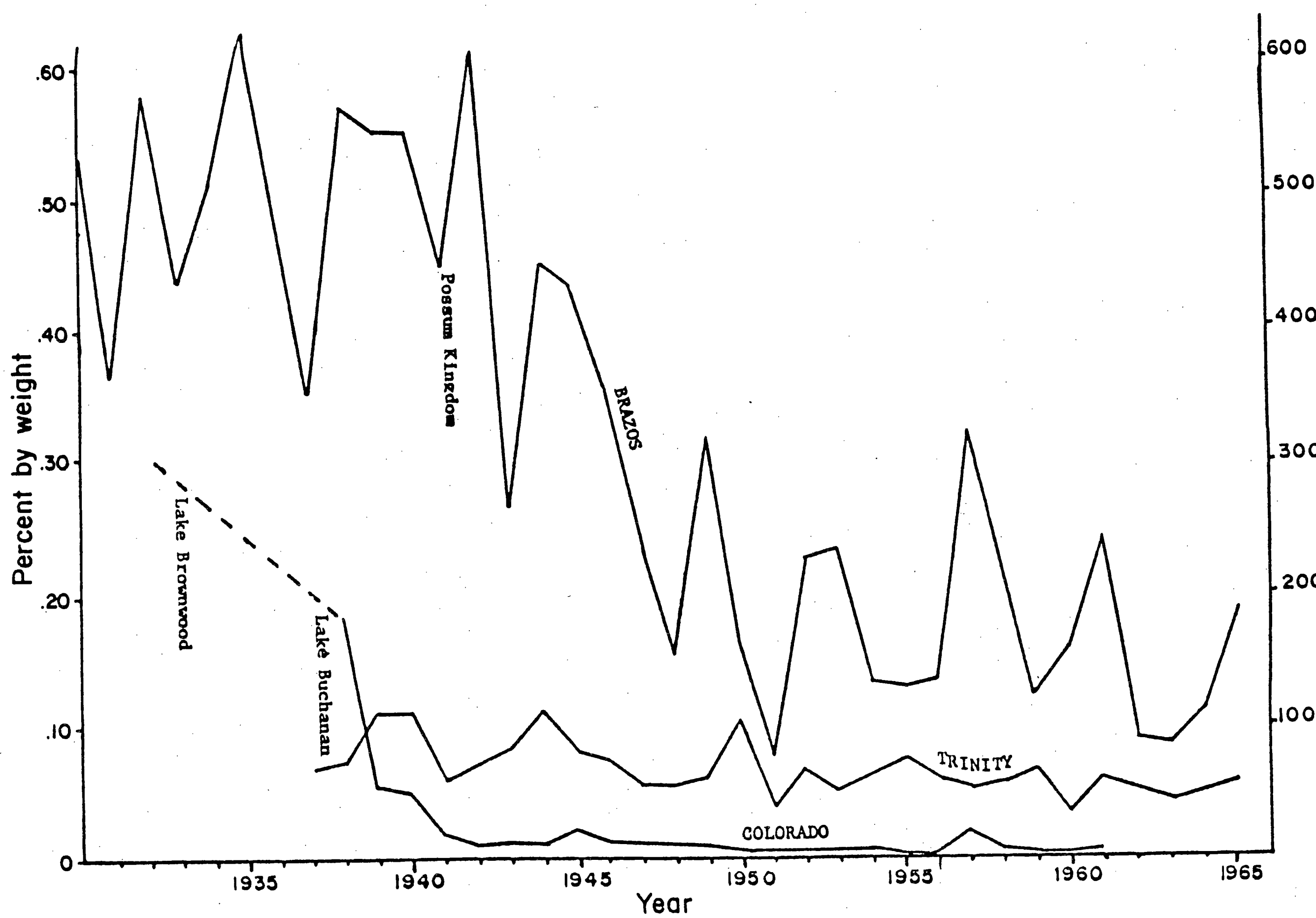


Figure 36. Annual suspended sediment loads of the Brazos, Trinity, and Colorado Rivers, with declines in load related to reservoir development. Sediment load along the vertical axis is concentration in percent by weight. (From USDA, 1972.)

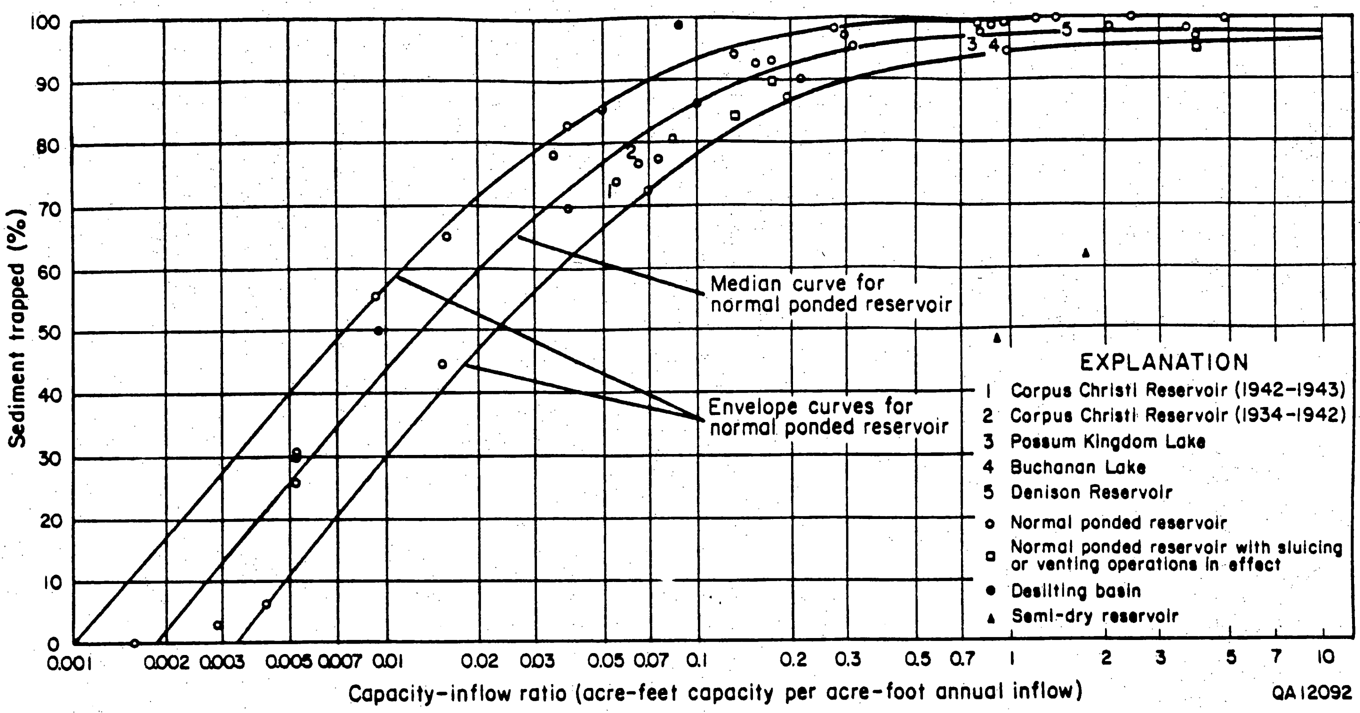


Figure 37. Trap efficiency curve. (Modified from Brune, 1953.)

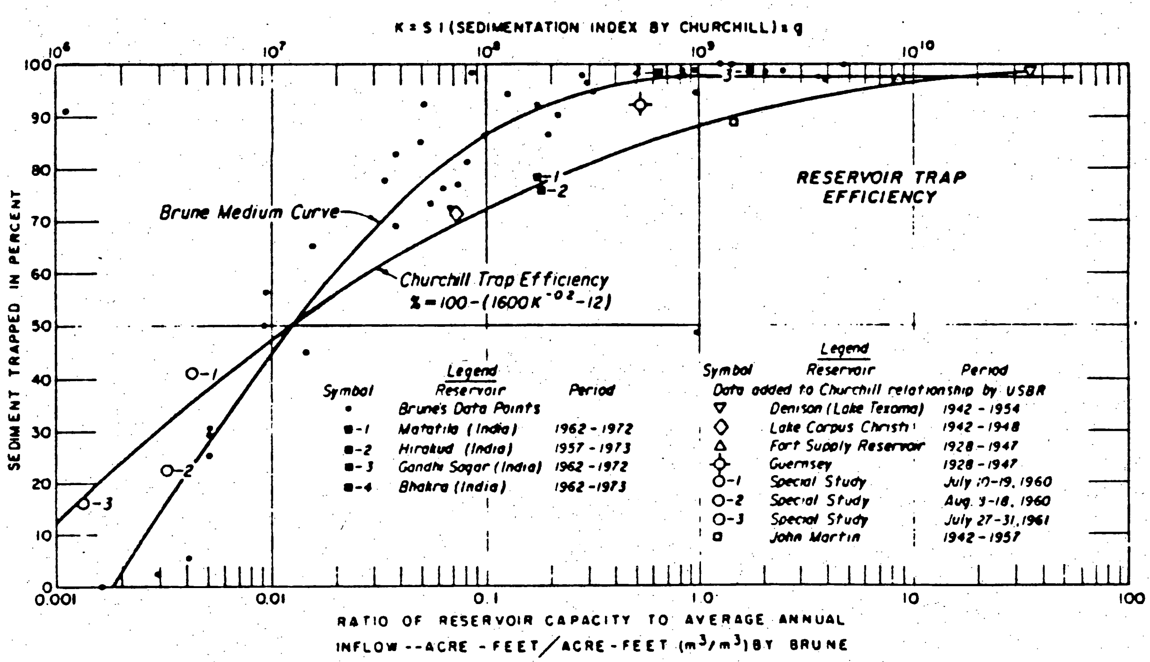


Figure 38. Comparison of trap efficiency curves from Brune (1953) and Churchill (1948). (From Strand and Pemberton, 1982.)

Colorado River are about 98 percent. The trap efficiency of Lake Corpus Christi, which is about 74 percent for 1942–1948 (Brune, 1953), has increased more recently.

The Soil Conservation Service of the U.S. Department of Agriculture (USDA, 1959) in an inventory of sedimentation in Texas, has presented data on reservoir sedimentation, including reservoir trap efficiency, within the various river basins. In addition, a considerable amount of data on sediment outflow from three to four reservoirs in Texas has been accumulated through measurements by Texas Water Development Board (Cook, 1970; Mirabal, 1974; Dougherty, 1979). Data on reservoir trap efficiency defined by ratios of reservoir capacity to reservoir inflow are presented by Cook (1970) and Mirabal (1974). Leibbrand (1987) estimated the sediment input into Lake Corpus Christi for the 1972–1985 period to be 5,320 acre-ft (dry); the sediment output during the same period was 117 acre-ft (dry). The amount deposited in the lake was about 5,140 acre-ft (dry), or approximately 97 percent of the sediment that entered.

Detailed sedimentation surveys have been conducted in some reservoirs in Texas. For example, Govin (1973) conducted a sedimentation survey of Lake Buchanan on the Colorado River in North-Central Texas. He found that about 101,400 acre-ft of sediment has accumulated in the lake since 1937 when impoundment began. Among his findings are that (1) average rate of sedimentation was 2,800 acre-ft/yr, which represented an annual average loss in reservoir capacity of 0.29 percent, (2) maximum accumulation occurred in the upper lake and in the deep river channel of the lower lake, (3) sediments were predominantly medium and fine-grained silt and clay from the river, and (4) the river's bed load and the coarser fraction of the suspended load must be deposited well above the lake. The average annual denudation rate of the drainage basin was calculated at a minimum of 0.11 acre-ft/mi^2, which Govin considered to be in reasonable agreement with a rate of 0.32 acre-ft/mi^2 calculated by Ritter (1967) for the Colorado River in Central and West Texas.

Effects Downstream from Reservoirs

Williams and Wolman (1984) compiled a large data set detailing the downstream effect of dams by analyzing 1,817 measurements of 287 cross sections downstream from 21 dams. Although there is wide variation in the post-dam water-discharge characteristics from river to river, flood peaks were generally decreased by the dams. Average annual peak discharges were decreased from 3 to 91 percent of the pre-dam values, with the average decrease being about 39 percent. Suspended sediment load decreased markedly for hundreds of kilometers downstream from dams, and on some rivers annual loads did not equal pre-dam loads anywhere downstream. The distance downstream from dams that is required for a river to regain the sediment load equivalent to the pre-dam load varies. In some major rivers, annual sediment loads may not equal pre-dam values for hundreds or thousands of kilometers downstream, if at all. Degradation of channel beds generally occurred during the first decade or two after dam completion. The magnitude of degradation varied from negligible to approximately 7.5 m (24.6 ft). The general trend was for bed material to coarsen as degradation proceeded, although this trend may change in later years. Channel width downstream from a dam can increase, decrease, or remain constant. Vegetation commonly increased downstream from dams, probably as a result of reduction in peak flows after dam closure.

Degradation or erosion downstream from dams is a common occurrence (Williams and Wolman, 1984), as noted above. However, sedimentation below dams can also occur (for example, Minter, 1976; Woolley, 1985; Hobbs, 1987). Minter (1976) concluded that among the effects caused by construction of dams on the Brazos River and its tributaries were a reduction in peak flood flows and a great reduction in the river's sediment carrying capacity. The construction of Lake Whitney on the Brazos has decreased peak floods by 50 to 65 percent, resulting in a decrease in sediment transporting power (competency) and a local build-up of gravel deposits in the channel at the mouths of tributary streams (Woolley, 1985; Hobbs, 1987). Although a reduction in sediment load below the dam has promoted channel degradation or erosion, net aggradation has occurred near tributary junctions (Fig. 39). The inability of the Brazos to transport the size of the gravel delivered to it by its tributaries has led to armoring and stabilization of the clastic deltas (Hobbs, 1987).

Minter (1976) concluded that major dam and reservoir development within the Brazos River basin has significantly reduced the amount of suspended load and bed load reaching the Texas

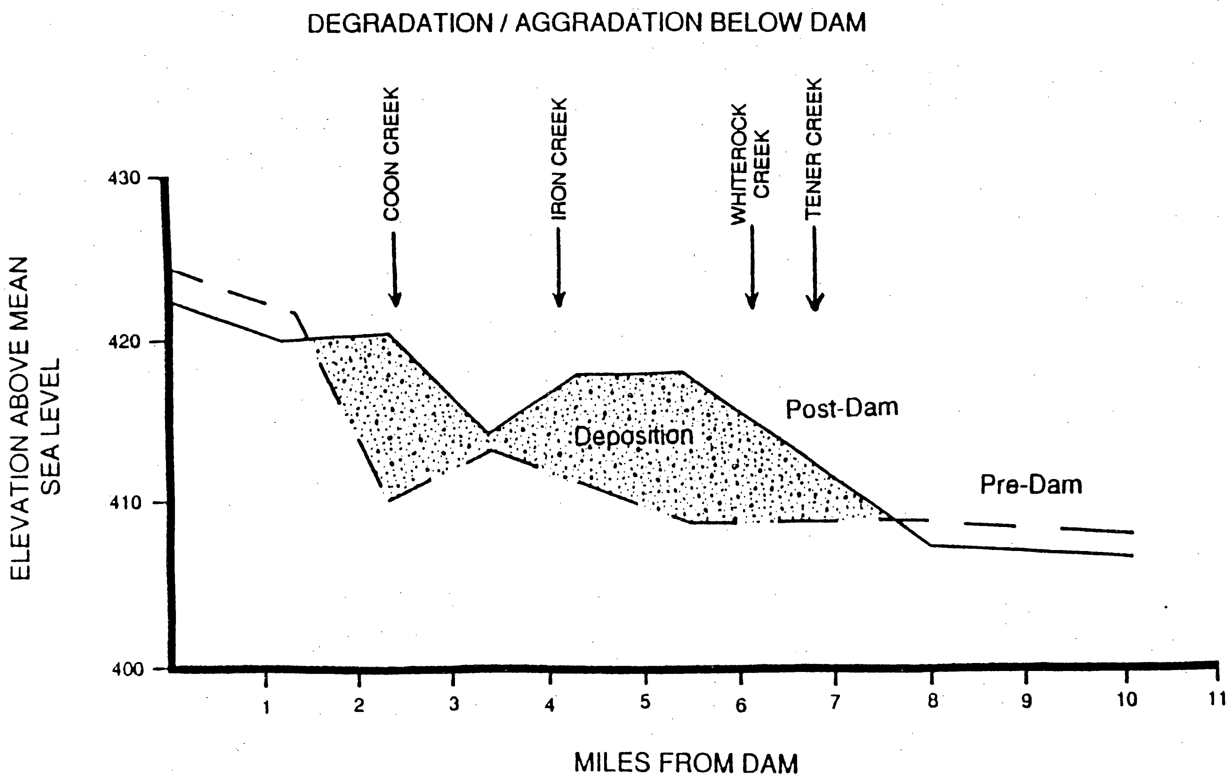


Figure 39. Pre-dam and post-dam longitudinal profiles of the Brazos River below Lake Whitney Dam, indicating net deposition along several miles of the channel. (From Woolley, 1985; illustration from Hobbs, 1987.)

coast. It was estimated that reservoirs are trapping approximately 76 percent of all sand derived from the drainage basin. According to Minter (1976), the amount of sand lost by entrapment plus that lost due to declines in river transporting power can account for the entire increase in rates of Gulf shoreline erosion (Seelig and Sorensen, 1973; Morton and Pieper, 1975) near the mouth of the Brazos, at least since 1937.

In contrast to the findings of Minter (1976), who reported a loss in sand delivered to the coast after reservoir development, Isphording (1986), in a study of sediments deposited in Apalachicola Bay, Florida, reported an increase in sand delivered to the bay after reservoir development. Isphording found abrupt changes in sediment regimen in the bay, including increases in sand and clay but a striking decrease in silt, which he traced to construction of a number of dams on rivers that discharged into the bay. Noting that the reservoirs trap sand and silt, he hypothesized that clay is washed over the spillways and continues down the river eventually to settle out in bay waters. He attributed increases in sand in the bay to channel erosion downstream from the reservoirs. The dam nearest the bay was approximately 160 km (100 mi) upstream.

Classification of Rivers by Sediment Load

Rivers can be classified in terms of their sediment load (Schumm, 1968, 1972) as suspended load, mixed load, and bed load streams. Bed load is that part of the load transported along the bed of the channel, and is usually composed of the coarser materials such as sand, gravel, and larger material that slide, roll, or bounce (saltate) along the bottom of the stream; suspended load is that which is held in the water column, or suspended above the channel bed, and is composed primarily of the finer sized materials—mostly silt and clay but also some sand.

Suspended-load streams transport on average less than 3 percent sand size or larger sediments (bed load), and are generally characterized by relatively narrow, deep, and sinuous channels. Bed-load streams transport on average greater than 11 percent sand size and larger material, and are relatively wide, shallow, and straight. Mixed-load streams have sediment characteristics and morphology intermediate between these two end members (Schumm, 1972). The Guadalupe and San Antonio Rivers are examples of suspended load streams, and the Brazos and Colorado Rivers vary from predominantly bed-load streams in their upper reaches to mixed-load streams on the coastal plain (Morton and McGowen, 1980).

Sediment dispersal patterns and modern depositional systems along the Gulf Coast are depicted in Figure 40. Deltas fed by modern rivers are significant depositional features along the coast.

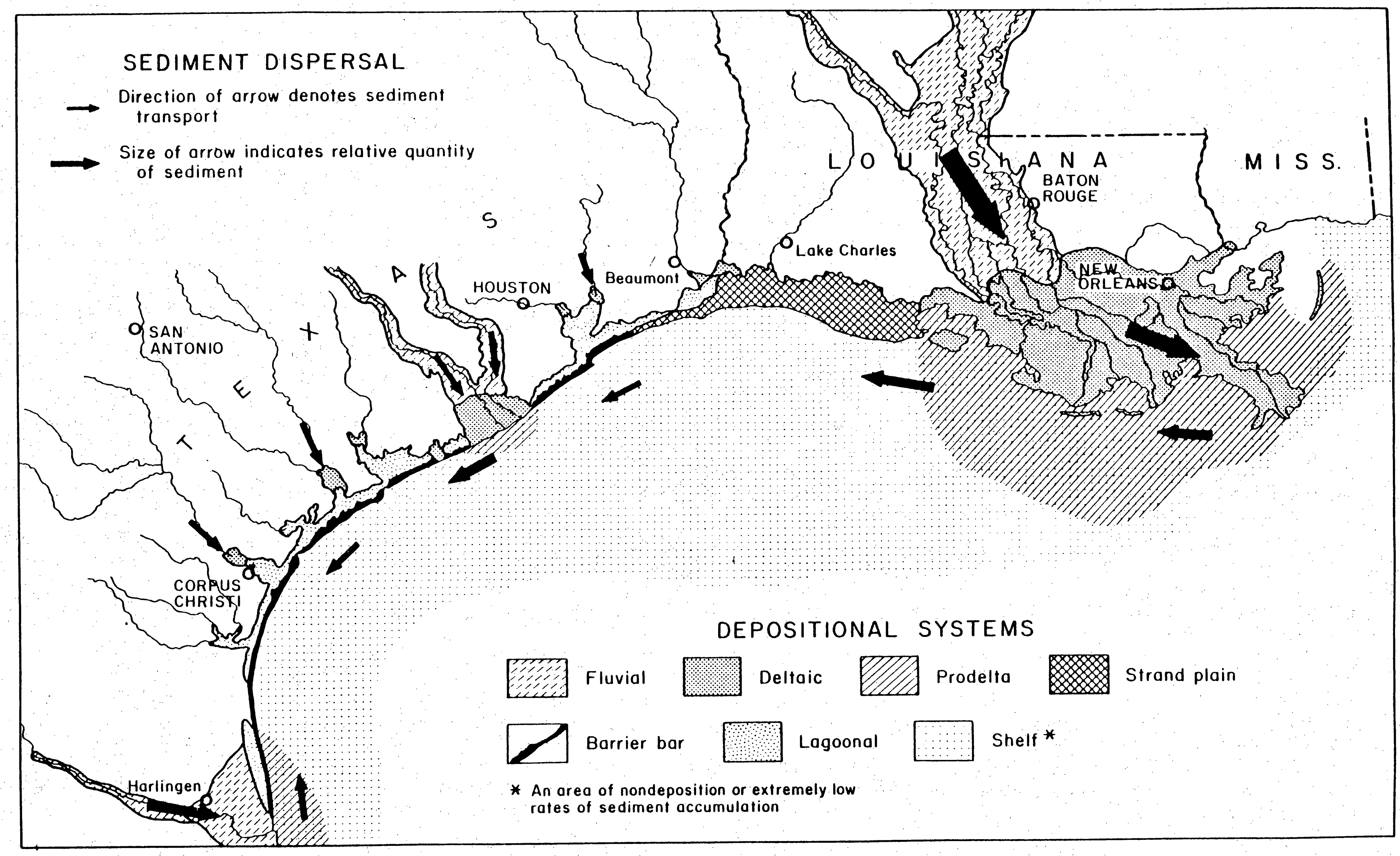


Figure 40. Modern depositional systems and sediment dispersal patterns, northwest Gulf of Mexico; by A. J. Scott. (From Fisher and others, 1969.)

FLUVIAL-DELTAIC-WETLAND SEDIMENTATION

"Of all geologic processes, fluvial sediment transport and deltaic deposition are among the most dynamic..." (Morgan, 1970a). A delta as defined by Fisher (1969) is simply "a river-fed depositional system that results in irregular progradation of a shoreline. A complex of delta lobes comprise a delta system." In a broader definition of deltas, Wright (1985) included both subaqueous and subaerial accumulations of river-derived sediments that are deposited at or near the source stream; his definition includes these deposits when reworked by waves, currents, and tides. Although modern deltas may have general similarities in that they are formed from sediment deposited as a stream loses its velocity upon reaching base level (the receiving reservoir), deltas may vary greatly in terms of their composition, size, shape, origin, and sedimentary properties (Morgan, 1970a). Deltas range from those that are large-scale depositional units supplied by rivers with high discharge and large sediment loads and that prograde into oceanic waters, to those that are relatively small scale, such as bayhead deltas, that prograde into bays and estuaries (Fisher, 1969).

Delta Development

Among the variables that affect delta formation are: (1) sediment input load such as amount, rate, variation in rate, and suspended/bedload ratio; these factors are affected by climate and extent and nature of the drainage basin, (2) nature of the discharging and reservoir water bodies, especially regarding relative water densities, (3) reservoir energy such as the kind and degree of waves, currents, tides, particularly as related to the amount of sediment input, (4) depth of water into which the delta progrades, (5) nature of substrate of the receiving reservoir, which affects subsidence and storage of prograding sediments, and (6) structural nature of the receiving basin (Fisher, 1969).

The first prerequisite for delta formation is the existence of a major river system composed of a drainage basin within which sediments are supplied by erosion from precipitation and runoff. Individual tributaries coalesce to produce a larger trunk stream, which is housed within an alluvial valley that connects to the coastal receiving basin (Fig. 41). A general discussion of the history of deltaic studies and delta formation is presented by Morgan (1970a), from which the following general discussion of delta formation was derived.

As sediment-laden water from the river enters an estuary or other receiving basin, density differences cause the river water to spread and gradually mix with the water of the basin. In the mixing zone, suspended sediment, including clays that flocculate, are deposited in subaqueous levees. Flocculation occurs as the fresher riverine water enters the more saline estuarine water. Clay particles are characterized by large surface areas that normally have a high negative surface charge. The like charges cause the individual particles to repel each other and remain dispersed. However, when the particles enter water with high concentrations of cations such as marine water, the particles become destabilized and form floccules held together by electrochemical bonding forces. The flocculation process aids sedimentation because the larger floccules settle faster than the individual smaller particles. Clay minerals flocculate and settle at different rates depending on salinities. Kaolinite and illite, under laboratory conditions, flocculate rapidly as salinity increases between 1 and 2 ppt; montmorillonite flocculates over a much wider range of salinities and accordingly settles at a slower rate. Flocculated particles may deflocculate if returned to fresh water. Turbulence can bring more particles together, thus enhancing flocculation, but floccules can also be torn apart by high turbulence that produces shear stresses exceeding the attractive forces of the floccules. This discussion on flocculation was derived principally from Nichols and Biggs, (1985); studies of flocculation in estuaries include Postma (1967) and Kranck (1984).

The coarser bed load is also moved into the receiving basin, where as current velocity decreases, it is deposited at the mouth of the river, forming a river mouth bar. Most of the suspended load and some of the bed load is transported beyond the river mouth bar and deposited in an environment referred to as the delta front. Beyond the delta front is an area called the prodelta

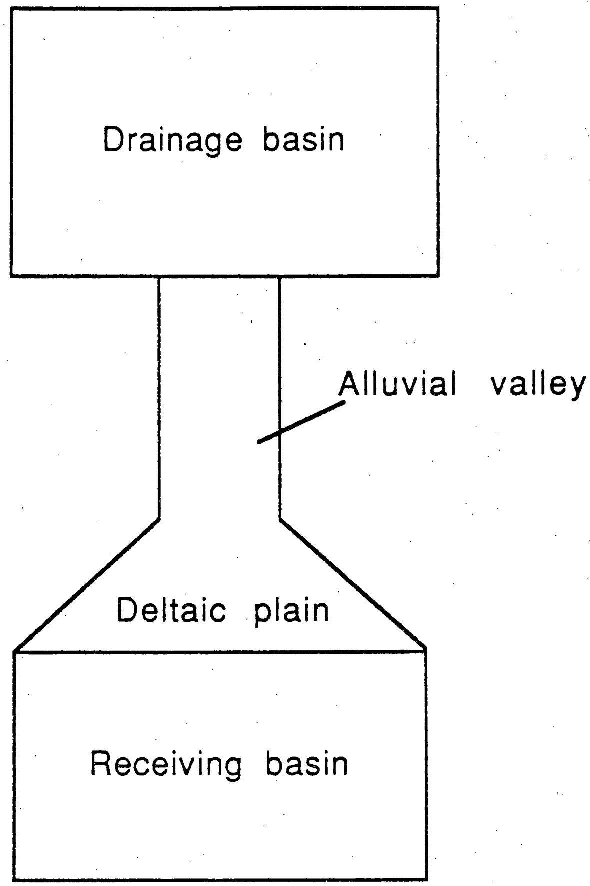


Figure 41. Major components of a river system (from Coleman and Wright, 1971).

depositional environment, where finer fractions of the suspended load gradually settle to the bottom, forming a blanket of clay and silt. The rates of deposition of sediments around the river mouth are dependent not only on the rate of water mixing and corresponding reductions in water velocities but also on the energy characteristics, such as strength of tides, waves, and currents, of the receiving basin. The interrelated factors that have a bearing on deltaic sedimentation are summarized in table 8.

As deposition at the river mouth continues, the subaqueous levees and river mouth bar eventually become subaerial. Deposition of the river mouth bar causes the channel to divide, or bifurcate, and new subaqueous levees are constructed along the distributaries (Fig. 42). The process of subdivision is repeated in geometric progression (2, 4, 8, etc.) if not opposed by wave erosion and longshore currents (Russell, 1967). The result is a delta, characterized by branching distributaries (Fig. 43), that progrades seaward (Fig. 44).

Successive flooding and deposition of sediment continues to build up the delta and develop the environments depicted in Figures 45 A and B. Coarser sediments (sand) characterize the channels and channel mouth bar deposits, sand and silt characterize the delta front sediments, and clay and silty clay the prodelta deposits. During floods along the distributary channel, coarser sediments are deposited closest to the channel as the river currents diminish rapidly when the flood water leaves the channel. Finer sediments are carried into the flood basins or interdistributary basins, where they are deposited in slack water. The natural levees formed along the channel are composed of silt and sandy silt, sediments finer than those in the active channels but coarser than the floodbasin deposits.

Deltas formed at the heads of estuaries—for example, the Guadalupe and Trinity deltas along the Texas coast—prograde over muds that were deposited in the estuarine environment. As the delta moves seaward, the fine-grained prodelta sediments override the bay muds (silt and clay), and they themselves are overridden by the successive deltaic deposits, including the delta-front sands (Fig. 44). However, the distributary channels may cut down into the bay muds and occupy a lower elevation than the bay floor (Donaldson and others, 1970).

Table 8. Factors influencing deltaic sedimentation. (From Morgan, 1970b.)

RIVER REGIME (Variations influence sediment load and transport capacity	Flood stage	Sediment load	Quantity of suspended load and bed load (that is, stream capacity) increases during flood
		Particle size	Particle size of suspended load and bed load (that is, stream competence) increases during flood
	Low river stage	Sediment load	Stream capacity diminishes during low river stage
		Particle size	Stream competence diminishes during low river stage
COASTAL PROCESSES	Wave Energy		High wave energy with resulting turbulence and currents erode, rework, and winnow deltaic sediments
	Tidal range		High tidal range distributes wave energy across an extended littoral zone and creates tidal currents
	Current strength		Strong littoral currents, generated by waves and tides, transport sediment alongshore, offshore, and inshore
STRUCTURAL BEHAVIOR (With respect to sea level datum)	Stable area		Rigid basement precludes delta subsidence and forces deltaic plain to build upward as it progrades
	Subsiding area		Subsidence through structural downwarping coupled with sediment compaction allows delta to construct overlapping sedimentary lobes as it progrades
	Elevating area		Uplift of land (or lowering of sea level) causes river distributaries to cut downward and rework their sedimentary deposits
CLIMATIC FACTORS	Wet area	Hot or warm	High temperature and humidity yield dense vegetative cover, which aids in trapping sediment transported by fluvial or tidal currents
		Cool or cold	Seasonal character of vegetative growth is less effective in sediment trapping; cool winter temperature allows seasonal accumulation of plant debris to form delta plain peats
	Dry area	Hot or warm	Sparse vegetative cover plays minor role in sediment trapping and allows significant aeolian processes in deltaic plain
		Cool or cold	Sparse vegetative cover plays minor role in sediment trappings; winter ice interrupts fluvial processes; seasonal thaws and aeolian processes influence sediment transportation and deposition

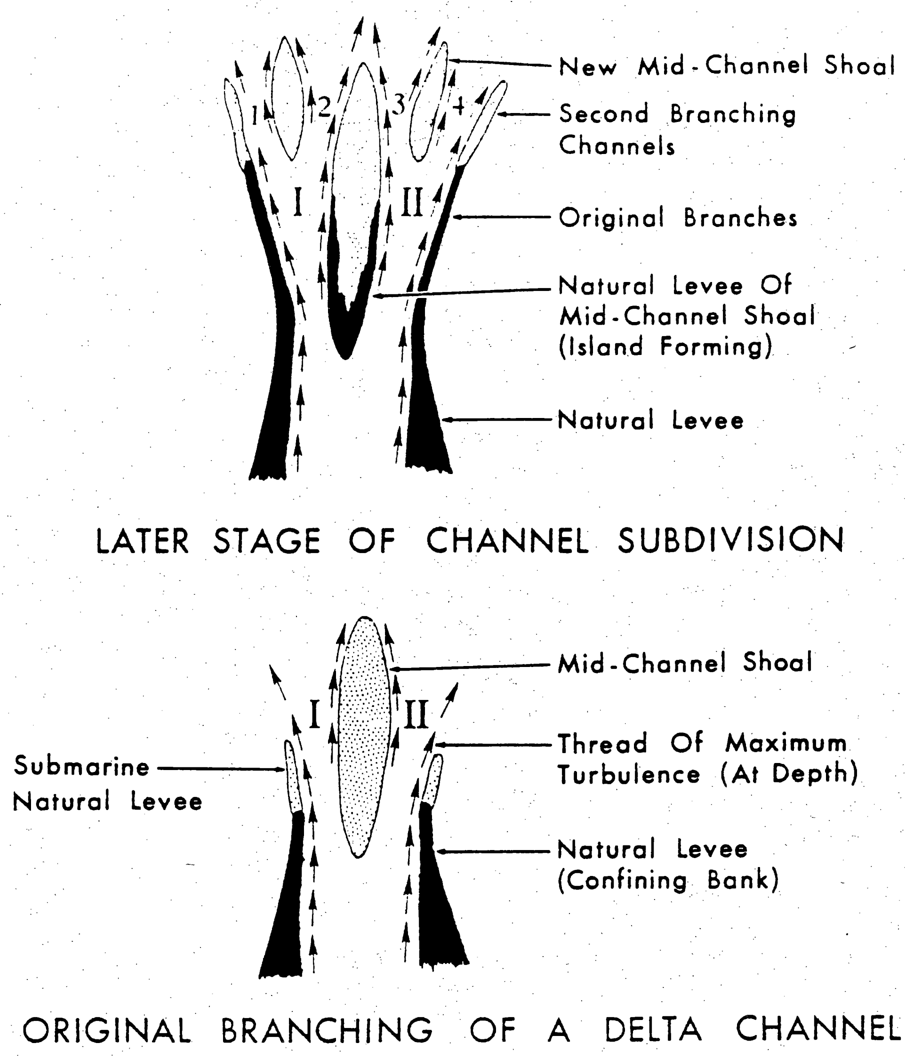


Figure 42. Development of mid-channel shoals and branching distributaries. (From Russell, 1967.)

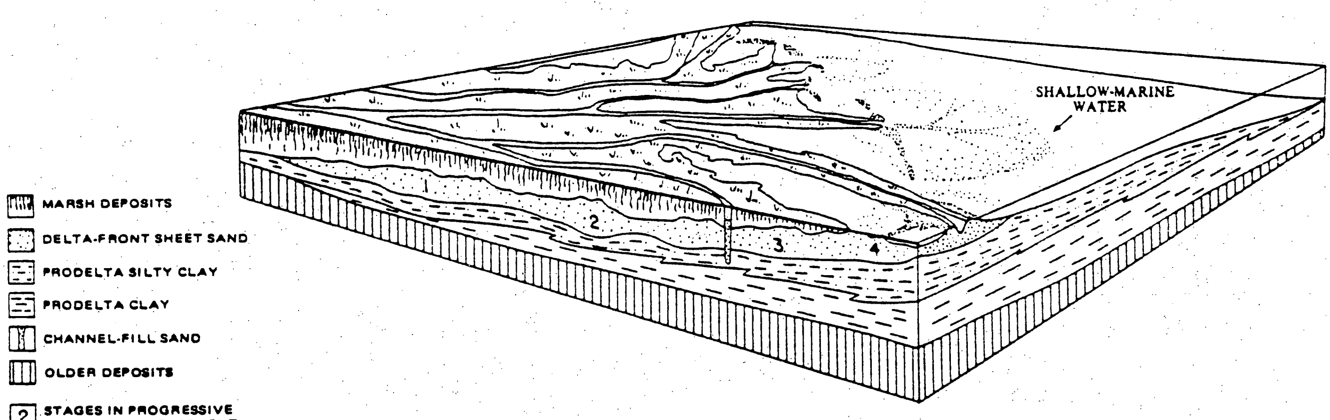


Figure 43. Components of a delta developing in shallow water. (From Gould, 1970.)

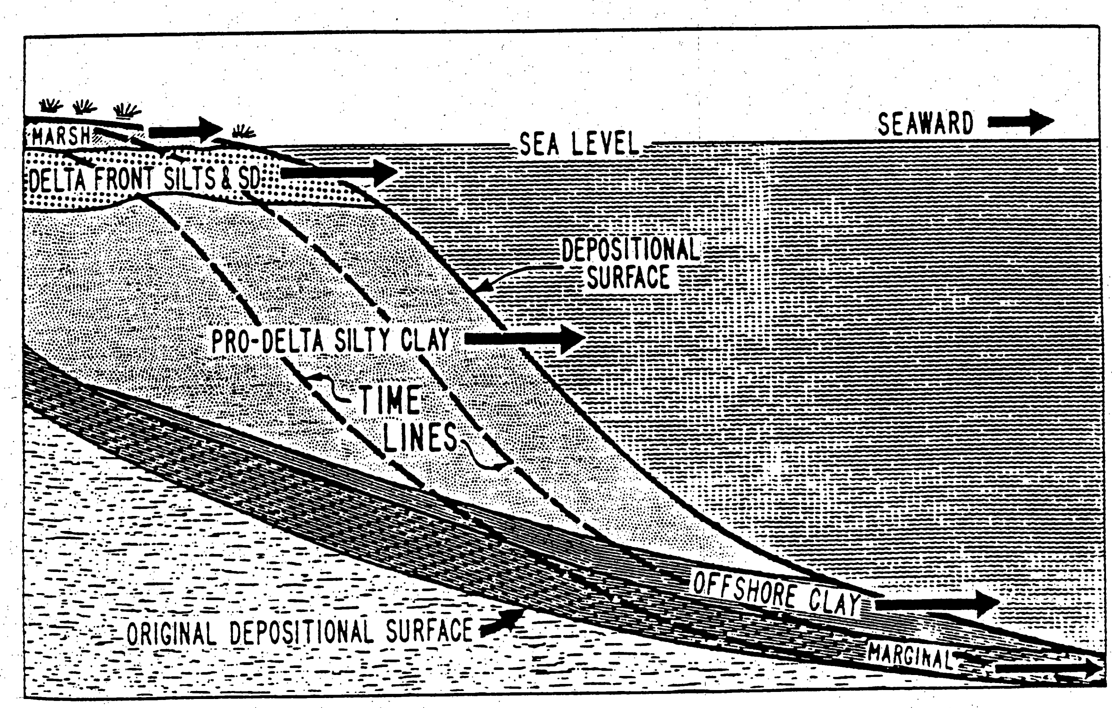


Figure 44. Seaward migration of depositional environments in high-constructive deltas, typical of bay-head deltas (Fisher and Brown, 1972). (From Scruton, 1960; reprinted by permission.)

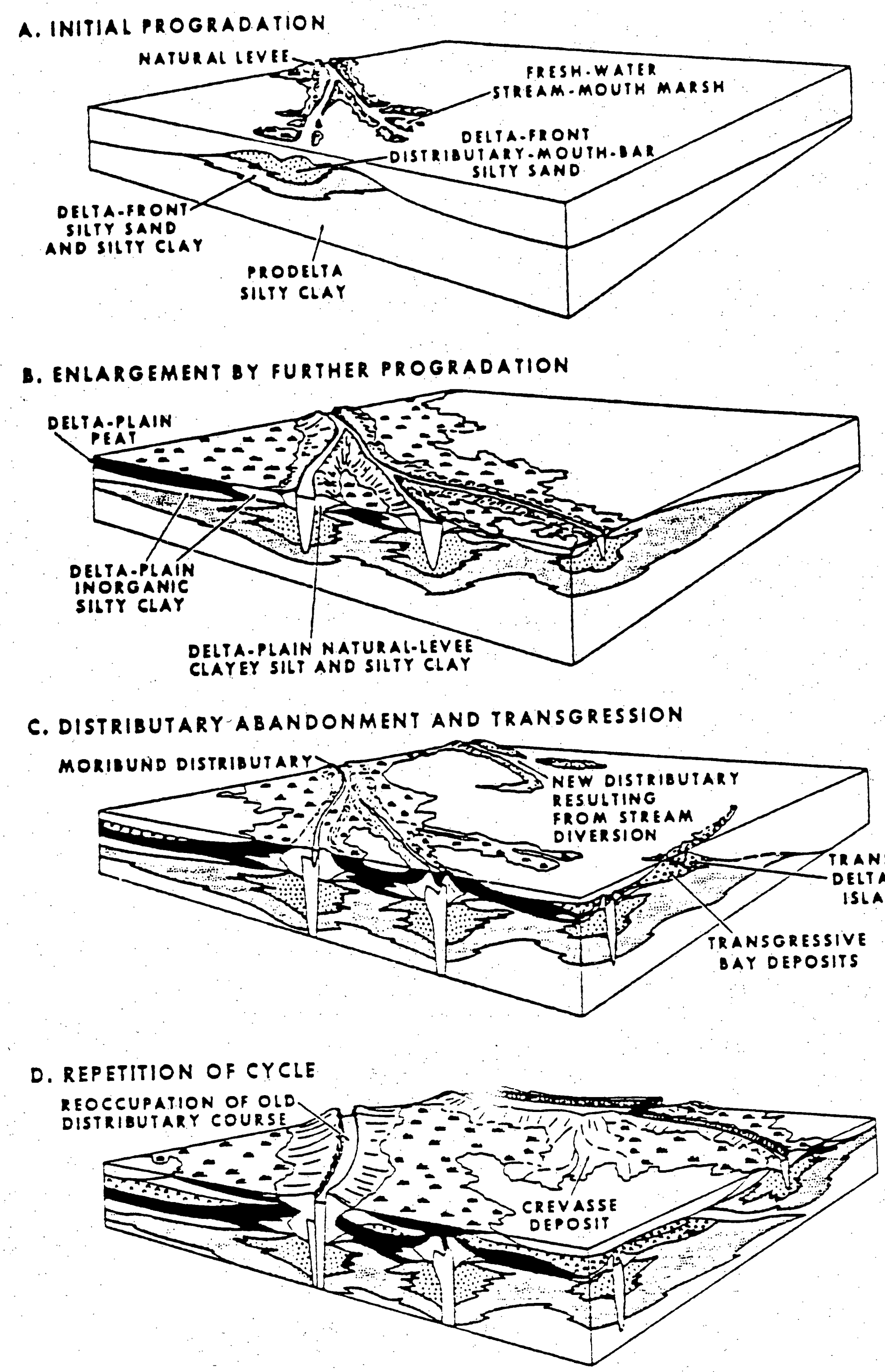


Figure 45. Development of deltaic facies, including constructional and destructional cycles. (From Frazier, 1967; reprinted by permission.)

As noted by Morgan (1970a, p. 112), deltaic distributaries prograde when the distributary "finds a shorter route of steeper gradient to base level; as a rule this occurs when the river crevasses or breaks through its own levees into an adjacent bay." Typical longitudinal gradients before crevassing may be from 1.8 to 2.8 cm/km (0.10 to 0.15 ft/mi); the cross-levee gradient at the point of the crevasse may exceed 100 cm/km (5 ft/mi) (Morgan, 1970a). This steep gradient produces high velocities and turbulence that allows the crevassing stream to scour a channel through its levee and deposit a sediment splay (crevasse splay) in the adjacent bay water (Fig. 45 C). If the channel is scoured deeply enough it will continue to supply sediment to the newly formed deposit during the succeeding low-water stages. Morgan (1970a) notes that subdeltas of the modern Mississippi River are classic examples of crevasses that have filled shallow bays along principal distributaries.

Development of Deltaic Marshes

As the delta environments aggrade, salt-water or brackish to fresh-water marsh plants become established in the intertidal and higher areas. The plants help baffle currents and trap sediments, thus building up the marsh (Redfield, 1972; Gleason and others, 1979; Stumpf, 1983). In intertidal areas, natural levees initially aggrade more rapidly than back marshes located in interdistributary basins, but as the levees reach higher elevations (along tidal channels), flooding and depositional events are less frequent and aggradation on the back marshes may start to catch up, thus producing a more level surface (Redfield, 1972). This general trend toward decreasing accretion rates and increasing elevations of the marsh surface has not been supported by some investigations (Oenema and DeLaune, 1988).

The higher marsh substrates generally contain more sand and silt than the lower marshes, which are typically characterized by mud (Kanes, 1970; Frey and Basan, 1985; Fig. 46). In addition to being topographically above the lower marsh, the higher marsh is older and is influenced more by terrestrial conditions (Frey and Basan, 1985). The lower marsh is often intertidal and more frequently inundated by waters carrying suspended sediments from the bay–estuary–lagoon system.

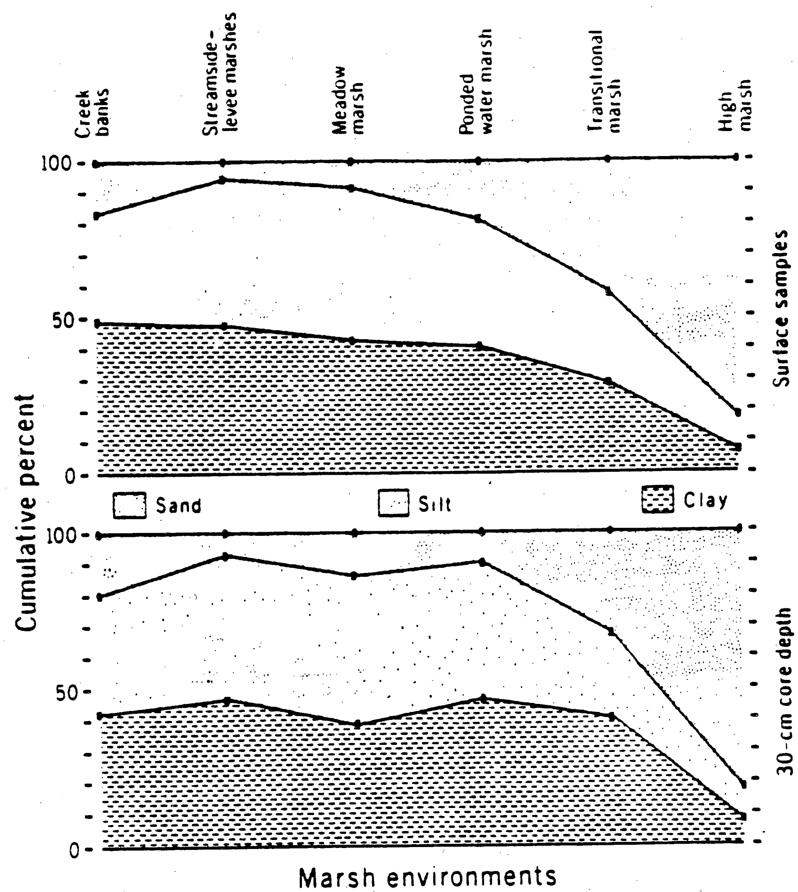


Figure 46. Relative percentages of sand, silt, and clay in salt marsh sediments, Sapelo Island, Georgia. (From Frey and Basan, 1985, after Edwards and Frey, 1977; reprinted by permission.)

Marsh Sediments

Frey and Basan (1985) noted that very few marshes along the Gulf coast have been studied in detail geologically and suggest that the marsh sediments (except for the Mississippi delta complex) seem to be similar to those of the southeastern Atlantic coast, which were described as dominantly inorganic with insignificant amounts of peat. Suggested reasons for the absence of thick peat deposits are (1) tidal flushing, (2) rapid degradation of plant material by intense biologic activity, and (3) extremely slow rates of coastal warping or submergence (Frey and Basan, 1985). Comparison of total organics in sediments in Texas brackish marshes and salt marshes (White and Calnan, 1989) indicates concentrations lower than those reported by Frey and Basan (1985) for Georgia salt marshes.

Inorganic sediments of a representative southern Coastal Plain salt marsh (near Sapelo Island) contained approximately constant proportions of silt and clay, where maximum amounts were 60 percent and 55 percent, respectively; sand is uniformly low in the low marsh and sand and muddy sand predominate in the high marsh (Frey and Basan, 1985) (Fig. 46).

In a study of a delta lobe on the eastern half of the Colorado River delta along the Texas coast, Kanes (1970) reported that the low-marsh sediments were characteristically finer grained than those on the higher marsh. He suggested that the sediments on the higher marsh were derived from the river during floods, while the low-marsh sediments were derived from turbid bay waters. In samples collected from salt marshes on the western half of the Colorado River delta, White and Calnan (1989) found clay (ranging to 80 percent) to be significantly more abundant than silt (maximum of about 30 percent) in the low marshes, while in a levee marsh, silt (at about 50 percent) was slightly more abundant than clay (about 40 percent). Sand was not a significant component of the marshes sampled except on nearby Matagorda Peninsula where it was predominant (ranging to almost 90 percent). Sediment samples in a brackish marsh system on the Trinity River delta along the upper Texas coast were more variable, maximum concentrations of clay and silt being about 80 percent and 20 percent in low (back) marshes; high marshes along levees had higher silt content, ranging to a maximum of near 55 percent, and a maximum clay content of near 40 percent. Marshes with sandier substrates were located on delta substrates undergoing active progradation near the bay margin. At one site near a distributary channel, the concentrations of sand, silt, and clay were 67, 18, and 15 percent, respectively (White and Calnan, 1989). The marsh at this latter site was only partly vegetated; sediment samples contained less than 0.5 percent organic carbon.

Marsh Degradation

In deltaic areas where fluvial sediments are abundantly supplied to the marshes during flood events, the marshes may follow a geologic succession from low marsh to high marsh to uplands as levee deposits aggrade through successive depositional events. When active distributaries are abandoned and the delta progrades into new areas, the fluvial-sediment supply to the older delta lobe is cut off or is reduced. Marshes may continue to receive sediments through tidal inundation, but will not remain emergent unless the sediment supply is enough to offset destructive processes. Processes that tend to degrade a delta such as waves and currents and subsidence begin to dominate and the abandoned delta lobe undergoes deterioration through erosion and submergence. Upland levee deposits may revert to marshes as the delta lobe subsides (Fig. 47).

Life Cycle of a Subdelta in a Delta System

The extensive loss of wetlands on the Mississippi River Delta (Gagliano and others, 1981) has made this area the center of deltaic wetlands research along the Gulf coast. Wells and Coleman (1987) noted that although the causes of wetland loss are complex, most scientists agree that a combination of natural processes (subsidence, sea-level rise, and changing depositional sites) and human-induced causes (artificial canals and levees and sediment diversion) are the major causes of

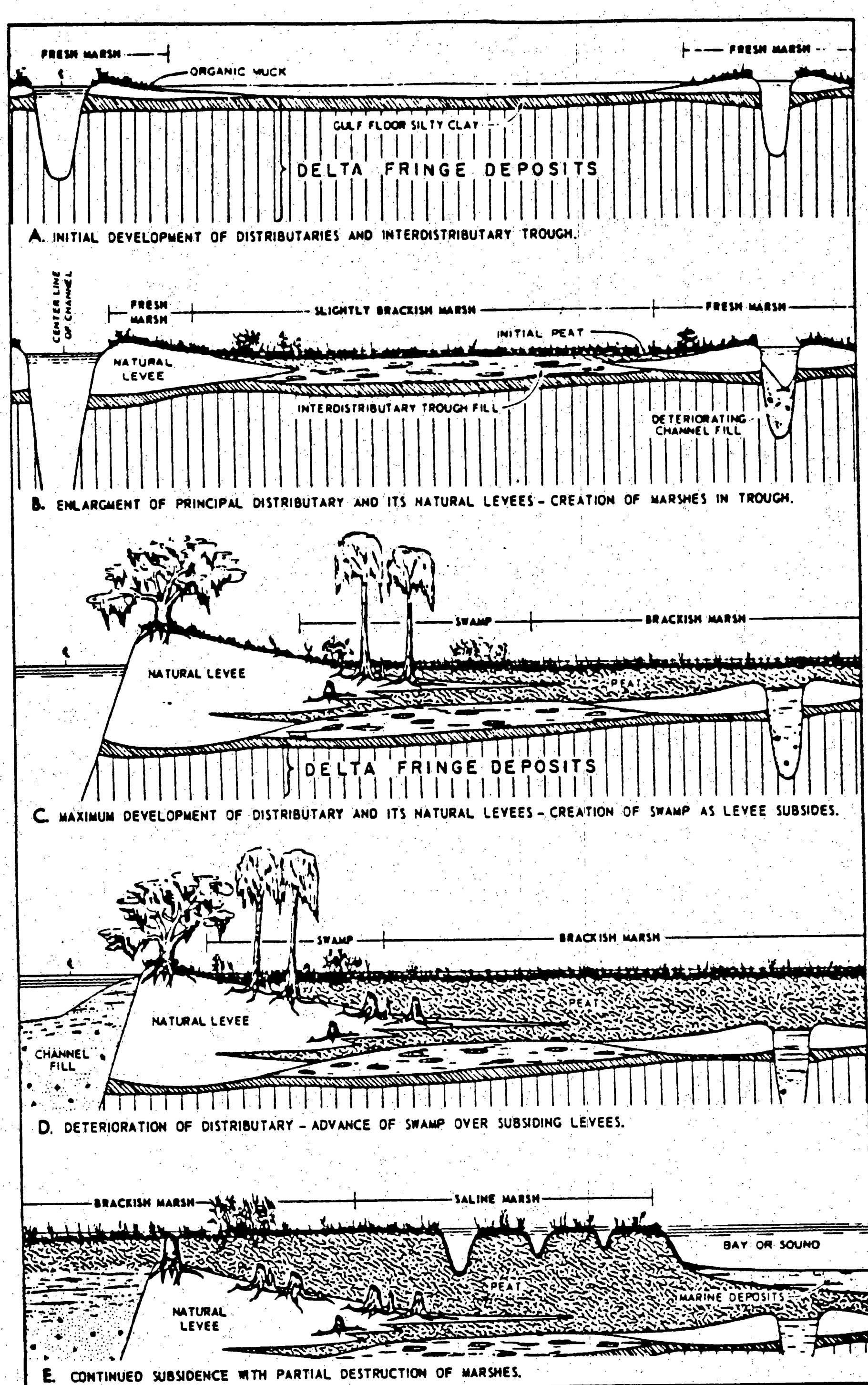


Figure 47. Development and deterioration of deltaic marshes. (From Bernard and LeBlanc, 1965, modified from Fisk, 1960.)

wetlands loss. Similar causes are affecting other deltas. For example, the combined effects of subsidence, sea-level rise, and a sharp reduction in sediment input because of the Aswan High Dam will likely submerge much of the northern Nile Delta within 30 km (about 20 mi) of the coast by the end of the next century (Stanley, 1988).

The life history of a subdelta, which undergoes a constructional and destructional phase (Scruton, 1960), is characterized as follows by Wells and Coleman (1987): (1) subdeltas follow a natural cycle of growth and deterioration that lasts for a period of about 115 to 175 years; (2) growth rate is regulated by depth of the receiving basin, sediment discharge through the crevassing channel, amount of sediment deposited in the receiving basin, and the efficiency of the channel network in the subdelta to deliver the sediment; (3) there is an initial period of slow infilling (10 to 30 yr), during which a subaqueous channel system is being constructed; this period is followed by rapid subaerial growth as channels extend farther into the receiving basin by elongation and bifurcation, until a peak is reached as the channel network becomes too complex (about 25 channels) to efficiently deliver sediments; (4) rapid subsidence is characteristic of the deteriorating phase, which allows for continuous infilling during subaerial deterioration; reversion to open water initially occurs in the older parts of the subdelta, and expands from there. The Mississippi delta has a complex evolutionary history (Fig. 48).

A significant process that leads to the destruction of abandoned delta environments is subsidence. Morgan (1967) noted that in addition to regional downwarping, subsidence can result from the effect of overloading and attendant compaction and water loss in underlying sediments. When sedimentation ceases following abandonment of a delta lobe, subsidence continues. The effects are apparent, first, in interdistributary marshes, which begin to be replaced by open water (Fig. 49). Because the distributary levees are more massive than other delta environments, they subside more rapidly and drag down adjacent marshes forming elongate ponds that parallel the levees (Morgan, 1967). As subsidence continues, the action of waves and currents in the expanding ponds accelerates the process through erosion, and the smaller ponds coalesce. The marshes ultimately are converted back to marine or estuarine open-water habitats.

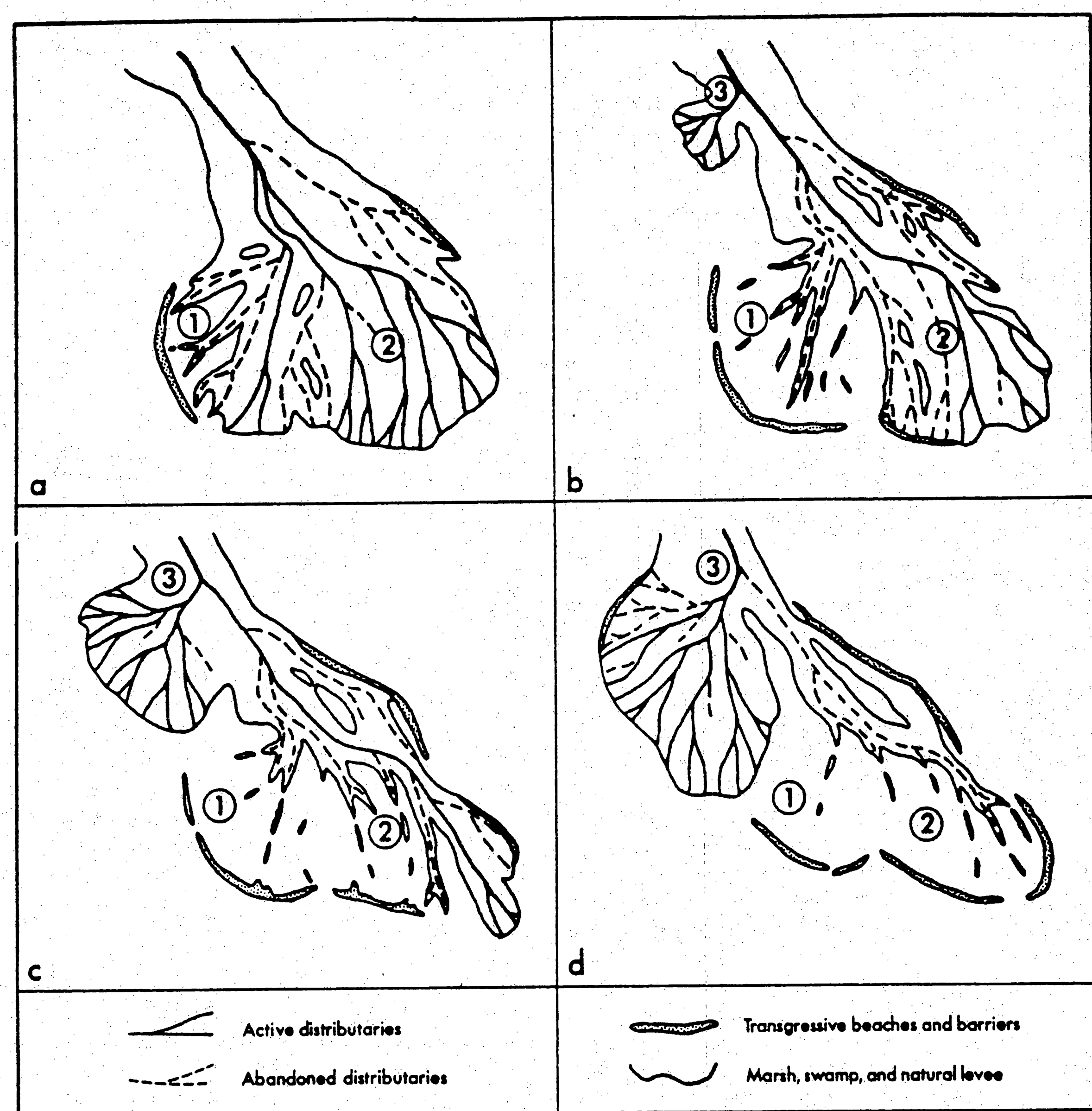


Figure 48. Evolutionary sequence of the Mississippi River delta distributaries. (From Morgan, 1977, in Boesch and others, 1983.)

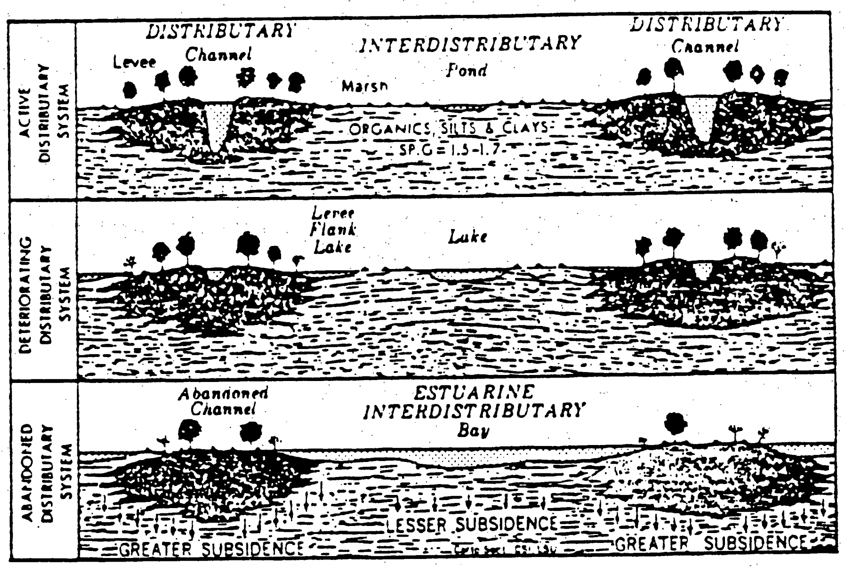


Figure 49. Effect of subsidence from sediment compaction on distributary and interdistributary areas. (Reprinted by permission, from Fig. 2, James P. Morgan, "Ephemeral estuaries of the deltaic environment," *Estuaries*, George H. Lauff, ed., p. 117, Copyright 1967 by AAAS.)

The deltas in the bay–estuary–lagoon system along the Texas coast have developed in a significantly different setting than the much larger Mississippi River delta, which has prograded over thick muds on the continental shelf. As noted by Winker (1979), deltas formed by the Guadalupe River (Donaldson and others, 1970) and Trinity River (McEwen, 1969) are examples of relatively thin deltas that have prograded into shallow, protected estuarine systems. Still, these deltas have gone through stages similar to the Mississippi River delta (Fig. 48). For example, the Guadalupe River delta at the head of San Antonio Bay has had a history of delta lobe construction, abandonment, and subsidence (Fig. 50). The reversion of marsh to open water occurs in the abandoned parts of the delta as subsidence outpaces sediment supply.

Marsh Maturation and Aggradation

After marsh vegetation begins to colonize deltaic deposits as described in the preceding discussions, the marsh substrate may continue to aggrade (accrete vertically) as sediments are delivered with each inundation. The rate of aggradation, or vertical accretion, depends on many factors such as (1) frequency, depth, and period of inundation, (2) amount of sediment transported into the marsh with each inundation, (3) type of inundation, whether from river flooding or estuarine tides, (4) location and elevation of the marsh, whether streamside (levee) or backmarsh, and (5) biological factors such as density and type of vegetation.

In his classic study of Barnstable marsh in Massachusetts, Redfield (1972, p. 224) reported that marsh aggradation rates were variable and depended "on the elevation which the intertidal marsh had reached, the availability of waterborne sediment, and the distance from open water." Rates of vertical accretion were highest in the initial stage of development of an intertidal marsh, as exemplified by a rate of 5.1 cm/yr. The rate of accretion diminished as the surface rose, producing an average rate of about 1.8 cm/yr. In the high marsh at greater distances from open water, long-term (approximately 1,000 yr) rates average about 0.21 to 0.15 cm/yr. Redfield concluded that the vertical accretion and the transformation of intertidal marshes to high marshes are dependent on

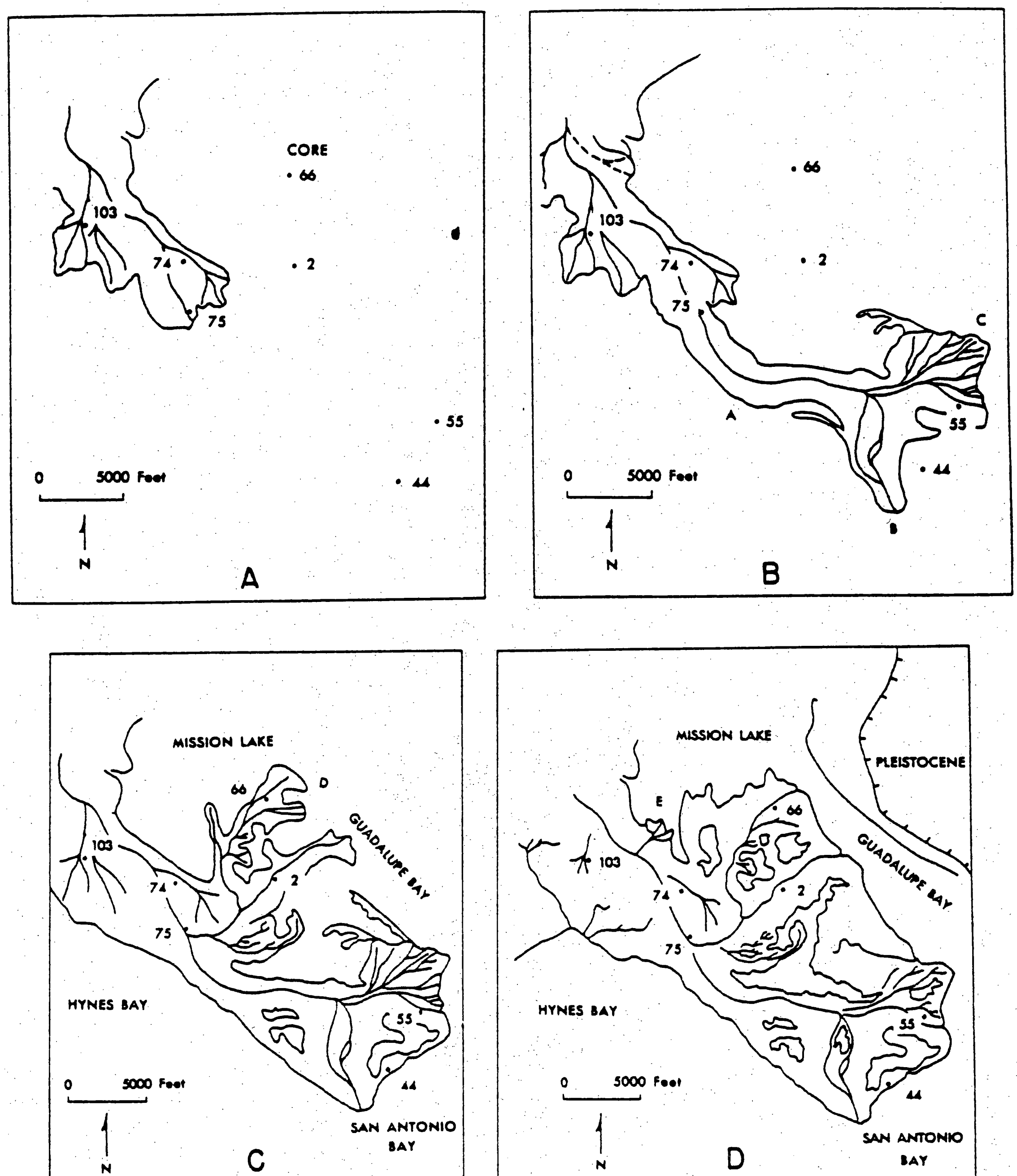


Figure 50. Growth and development of the Guadalupe delta at the head of San Antonio Bay during the past 2,000? years. Stages of growth and subdelta development progress from (A), which is the oldest subdelta, through (B) and (C), and finally to (D), which illustrates the present delta. The most recent development has occurred in Mission Lake at Traylor Cut subdelta located at (E) in illustration (D). (From Donaldson and others, 1970.)

availability of sediment. The accretion rate is presumably more rapid along the margins of the marsh where levees form, and less rapid in backmarsh areas. As the marsh surface becomes more extensive and reaches higher elevations, it becomes more inaccessible and the rate of accretion slows (Redfield, 1972).

In a study of tidal salt marshes in England, Pethick (1981) concluded that there is a positive statistical relationship between marsh elevation/age and vertical accretion rates. His model showed that young marshes (10-yr-old marshes) had accretion rates of up to 1.7 cm/yr, whereas older marshes (> 500 yr old) had rates as low as 0.002 cm/yr. The concept of the model is that younger marshes are lower in elevation, which amplifies the depth and frequency of inundations compared to older marshes. Older marshes have reached higher elevations, which decreases the frequency and period of inundation, resulting in less sediment deposition and lower rates of vertical accretion through time. Pethick reported that the age/height relationship defines an asymptotic curve where the asymptote is apparently controlled by the frequency of tidal maxima, which lies about 80 cm (2.6 ft) below the level of the highest spring tides in the North Norfolk, England, marsh that he investigated.

Stages of development, or maturation, of marshes have been recognized by many researchers (Fisk, 1960; Gould and Morgan, 1962; Chapman, 1974; Frey and Basan, 1985). As mentioned by Frey and Basan (1985), the maturation process is a reflection of floral and faunal succession, as well as size, position, and differentiation of the marsh in the coastal system. Geologically, succession occurs through progradation and the resulting lateral and vertical displacement of low marshes by high marshes (Frey and Basan, 1985). The elevation and age of a marsh may follow a chronology like that described by Pethick's model discussed in the preceding paragraph, but actual ages of low (young) and high (old) marshes may be different from those detailed by Pethick. Progradation and development of salt marshes at the mouth of the Colorado River along the central Texas coast, for instance, occurred primarily after 1929, when a logjam was removed from the river (the delta prograded approximately 6.4 km [4 mi] in about 6 yr (Wadsworth, 1966). High levee marshes and low interdistributary marshes, although still being modified, have developed during this relatively brief period. Frey and Basan (1985, p. 238) pointed out that while a scale of relative maturation, from youthful to mature to old, can be established for marshes, the emphasis "clearly is upon physiographic "stage" and not "finite" age". The Colorado River delta marsh can be characterized as youthful by the fact that low-marsh environments make up most of the total area. A marsh with a large terrestrial sediment supply and rapid deposition may generally favor high-marsh development, whereas the stable phase of a marsh with a low terrigenous sediment supply, pronounced tidal range and tidal sedimentation, and growth restricted to bay or estuarine margins may be submature to supermature (Frey and Basan, 1985).

Marsh sedimentation and rates of marsh aggradation have been investigated by many researchers, including Van Straatan and Kuenen, 1958; Ranwell, 1964; Schou, 1967; Pestrong, 1972; Redfield, 1972; Armentano and Woodwell, 1975; Flessa and others, 1977; Harrison and Bloom, 1977; Richard, 1978; DeLaune and Patrick, 1980; Letzsch and Frey, 1980; Nixon, 1980; Pethick, 1981; Hatton and others, 1982, 1983; DeLaune and others, 1983; Stumpf, 1983; Baumann and others, 1984; Boesch and others, 1984; Frey and Basan, 1985; Smith and Frey, 1985; Stevenson and others, 1985, 1986, 1988; Cahoon and others, 1987; and Oenema and DeLaune, 1988. Studies of accretion rates of Gulf Coast marshes have been conducted principally in Louisiana marshes where loss of wetlands has occurred at the alarming rate of more than 100 km^2/yr (39 mi^2/yr) (Gagliano and others, 1981). There is a lack of information on the rate of accretion in Texas marshes, although a loss in marsh area has been documented in the Galveston–Houston area (Johnston and Ader, 1985; White and others, 1985) and the Beaumont–Port Arthur area (Gosselink and others, 1979; White and others, 1987).

Marsh aggradation is an important process in which inorganic and organic sediments that are deposited on the marsh surface provide nutrients for plant growth and help keep the marsh emergent (DeLaune and others, 1978; 1981; DeLaune and Patrick, 1980). Some of the highest rates of marsh aggradation have been documented along the Gulf Coast in Louisiana (DeLaune and others, 1978; Boesch and others, 1983). The rates are considerably higher than along the east coast (table 9), and presumably are related to the high rates of subsidence and relative sea-level rise in coastal Louisiana

Table 9. Marsh aggradation (vertical accretion) rates measured in coastal Louisiana and along the U.S. Atlantic coast. (From Boesch and others, 1983.)

Location	Marsh type	Marsh accretion rate (mm/yr)	Mean sea-level rise (mm/yr)	Source
Louisiana Deltaic Plain	Freshwater		11.0	Hatton and others (1983)
	streamside	10.6		
	backmarsh	6.5		
	Intermediate (*Spartina patens*)			Hatton and others (1983)
	streamside	13.5		
	backmarsh	6.4		
	Brackish (*Spartina patens*)			Hatton and others (1983)
	streamside	14.0		
	backmarsh	5.9		
	Saline (*S. alterniflora*)		13.0	DeLaune and others (1978); Baumann (1980)
	streamside)	13.5		
	backmarsh	7.5		
Chenier Plain	Salt-brackish (*S. patens)*	7.0	12.0	Baumann and DeLaune (1982)
Georgia	*S. alterniflora*	3–5		Summarized by Hatton and others (1983)
Delaware	*S. alterniflora*	5.0–6.3	3.8	Summarized by Hatton and others (1983)
New York	*S. alterniflora*	2.3–6.3	2.9	Summarized by Hatton and others (1983)
Connecticut	*S. alterniflora*	8–10	2.5	Summarized by Hatton and others (1983)
	S. patens	2–5		
Massachusetts	*S. alterniflora*	2–18	3.4	Redfield (1972)

(DeLaune and others, 1978; Hatton and others, 1983). It has been proposed by Rusnak (1967) that the accumulation of sediments in estuaries must be equal to sea-level rise. Although the rate of aggradation in marshes may be decreased by compaction, it may be increased by subsidence and rise in sea level (Letzsch and Frey, 1980). The higher rates in areas of more rapid subsidence are probably due to more frequent inundations and sediment deposition; that is, there is a natural feedback loop (comparable to Pethick's, 1981, model describing the relationship between marsh accretion and its height and age) in which a submerging (and thus topographically low) marsh attempts to retain an equilibrium condition by trapping sufficient sediments to remain emergent. If there is insufficient sediment, however, the rate of aggradation will not keep pace with the relative rise in sea level and the marsh will eventually be replaced by open water. Rates of marsh aggradation in many areas of coastal Louisiana have apparently not kept pace with rates of sea-level rise and subsidence (DeLaune and others, 1978, 1983; Baumann and others, 1984; Penland and others, 1988).

Average aggradation rates (determined from analysis of ^{137}Cs in cores) in *Spartina alterniflora* salt marshes in Barataria Basin in Louisiana vary from 1.35 to 0.75 cm/yr (0.53 to 0.30 in/yr) in levee and backmarsh areas, respectively (DeLaune and others, 1978). Measurements of aggradation rates in fresh, intermediate, brackish, and salt marshes in Barataria Basin (also using ^{137}Cs) provided similar results of 1.35 to 0.75 cm/yr (0.53 to 0.30 in/yr) in levee and backmarsh environments (Hatton and others, 1983). Rates of aggradation in brackish marshes on the Chenier Plain near Calcasieu Lake to the east of Sabine Lake averaged 0.8 cm/yr (0.3 in/yr), as determined by ^{137}Cs and artificial marker horizons (DeLaune and others, 1983). The rate of relative sea-level rise (based on tide-gauge records) in these areas of Louisiana ranges from about 1.1 to 1.3 cm/yr (0.43 to 0.51 in/yr) (DeLaune and others, 1978, 1983; Hatton and others, 1983). A comparison of marsh accretion rates with mean sea-level rise for the Louisiana Deltaic Plain and Chenier Plain, as well as for marshes along the Atlantic coast, was summarized by Boesch and others (1983) in table 9. While the accretion rates of levee marshes in the Deltaic Plain are near the rate of relative sea-level rise, accretion rates in inland areas, or backmarshes, are not keeping pace and the marsh sites are deteriorating into small, open-water areas (DeLaune and others, 1978; Hatton and others, 1983). In the Chenier Plain, DeLaune and others (1983) conclude that an accretion rate of 0.8 cm/yr (0.3 in/yr) is not sufficient to maintain the elevation of the marsh in an area that is submerging at 1.2 cm/yr (0.47 in/yr).

Baumann and others (1984) investigated marshes in the Mississippi Deltaic Plain and found that although accretion rates in marshes of Barataria Bay ranged from 9 to 15 mm/yr (0.35 to 0.59 in/yr) and marshes in Fourleague Bay had accretion rates of from 6 to 13 mm/yr (0.24 to 0.51 in/yr) in inland and streamside marshes respectively, the marshes in Barataria Bay were deteriorating whereas marshes in Fourleague Bay were more stable. The authors attributed the closer balance between marsh aggradation and relative sea-level rise in Fourleague Bay marshes to fluvial sediments delivered by the Atchafalaya River. Annual river flooding was the major depositional process in Fourleague Bay marshes. Annual flood cycles contributed 91 and 69 percent of the streamside and inland sediments, respectively, during 1981 and 1982 (Baumann and others, 1984). Basinal sedimentary processes at work in Barataria Bay, where deteriorating marshes received most of their sediment during storm events, were insufficient to counter the effects of relative sea-level rise. Baumann and others (1984) further concluded that major accretion in Barataria Bay effectively ended with the damming of Bayou Laforche in 1904, and with the construction of an artificial levee system along the Mississippi River after the 1927 flood. Sediments delivered to the Barataria Basin marshes were principally a result of hurricanes, tropical storms, and winter storms (Fig. 51). Hurricanes and winter fronts, the latter occurring at an average frequency of 5.7 per winter month, are apparently major depositional and erosional agents in both bay areas. Baumann and others (1984, p. 224) stated, "the transformation from marsh to open bay cannot be halted unless there is a reintroduction of riverine sediment."

The delivery of sediments by river flooding is an important process, especially in areas characterized by low-tidal regimes. As pointed out by Frey and Basan (1985), the difference, in terms of hydraulics, between river and tidal flooding is one of duration; there is a longer slack-water period during river flooding, which allows more time for suspended sediments to settle out.

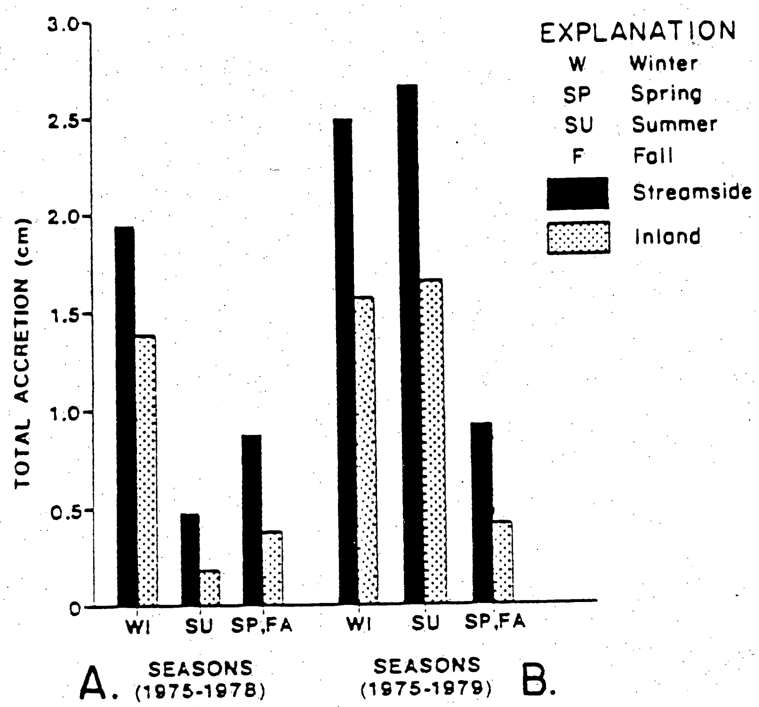


Figure 51. Seasonal variations in aggradation rates of salt-water marshes in Barataria Basin, Louisiana. Mean seasonal aggradation for 1975–1978 (A) indicates that rates were highest in winter, compared with mean seasonal aggradation for 1975–1979 (B), when rates were elevated by a hurricane and tropical storm that made landfall during the summer of 1979. (From Baumann and DeLaune, 1982).

Van Heerden and others (1981), using cores and bathymetric charts to study the evolution of the Atchafalaya Delta, found sedimentation rates of levee environments to be associated with annual flood events. They identified deposits with high sedimentation rates (> 0.3 m [1 ft] per annual flood), medium sedimentation rates (0.15 to 0.30 m [0.5 to 1 ft] per annual flood), and low sedimentation rates (< 0.15 m [0.5 ft] per annual flood).

In an investigation of the emerging Atchafalaya Delta, Louisiana, DeLaune and others (1987) reported that marsh accretion rates were as great as 1.4 cm/yr (0.55 in/yr). The higher rates were in bay bottoms and marshes closest to the active delta. Fluvial sediment input causes the delta to prograde and also provides the source of marsh nutrients supporting vegetation growth, which in turn provides organics for vertical marsh accretion. Marsh areas updrift, away from the prograding delta, had a smaller mineral sediment input as well as organic input and were accreting at a slower rate; the authors concluded that these areas will likely continue to deteriorate.

In the Barataria Basin, which receives insignificant riverborne sediment, there is a landward decrease in inorganic sediments as one progresses from the gulfward lying salt marshes to the landward lying fresh marshes (Hatton and others, 1982). The inorganic sediments, which increase in a seaward direction in response to hydraulic energy, are apparently locally derived from erosion of adjacent marshes. The organic matter accumulating in the marsh shows an inverse relationship to the inorganic sediment; that is, the accretionary role of organic matter increases in a landward direction, from saline- to fresh-water marshes (Hatton and others, 1982). These inorganic-organic relationships are reversed near active deltas where rivers and distributary channels are sources of inorganic minerals (Fig. 52).

Stevenson and others (1986) reviewed 15 areas (Atlantic and Gulf coasts) in a comparison of rates of marsh aggradation with rates of relative sea-level rise. They found that in at least four of the sites, three of which are along the Gulf coast, marsh aggradation rates were not keeping pace with relative sea-level rise (Fig. 53). Rates of relative (apparent) sea-level rise at the Gulf coast sites, however, are more than two times the rates at the other sites reviewed (Fig. 53). The authors found a strong correlation between mean tidal range and accretionary balance (the four sites that

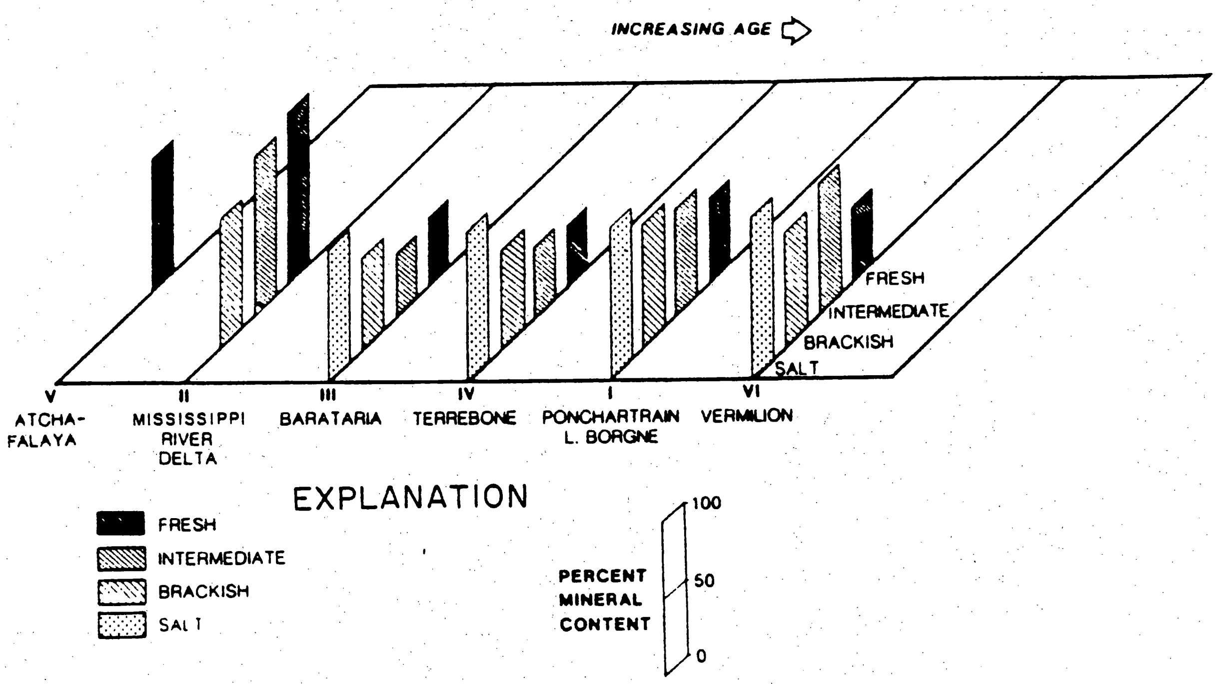


Figure 52. Percentages of inorganic (mineral) sediments in marsh soils in Mississippi delta basins arranged in order of increasing age. (From Gosselink, 1984, based on data from Chabreck, 1972.)

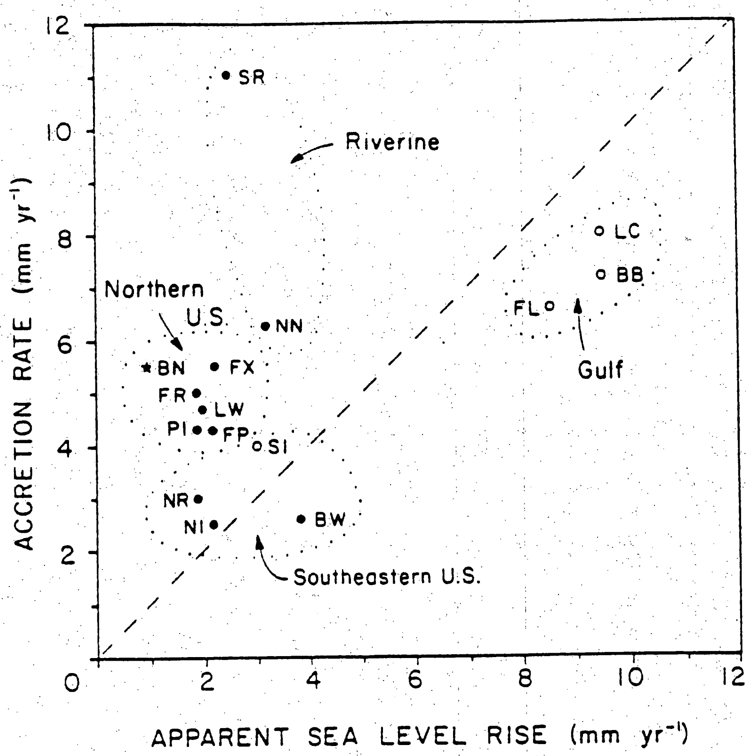


Figure 53. Relationship between rates of marsh aggradation and relative sea-level rise between 1940 and 1980. Aggradation rates were determined from ^{210}Pb (solid circles) and ^{137}Cs (open circles) measurements, and historical measurements (asterisk for Barnstable marsh). BB = Barataria Bay; BN = Barnstable marsh; BW = Blackwater marsh; FL = Fourleague Bay; FR = Farm River; FX = Flax Pond; FP = Fresh Pond; LC = Lake Calcasieu; LW = Lewes; PI = Prudence Is; NI = North Inlet; NN = Nanticoke River; NR = North River; SI = Sapelo Island; and SR = Savannah River. (From Stevenson and others, 1986; reprinted by permission.)

were not keeping pace with relative sea-level rise are in microtidal areas), but concluded that more data are needed from areas of low tidal energy to confirm the relationship. Nevertheless, Stevenson and others (1986) postulated that, where tides are weak and irregular and sediment inputs are low or reduced relative to past levels, marshes have difficulty in keeping pace with relative sea-level rise. They further suggest that hurricanes and storms, which can sporadically add sediments to marshes, may play a critical role in the sediment budget in areas with low tides, which is in agreement with Stumpf's (1983) conclusion that storms control sediment supply on microtidal marshes, and that sedimentation depends directly on storm frequency and sediment availability.

Nixon (1980) presented data on marsh accretion rates and sea-level rise to support his conclusion that salt marshes are sinks of suspended sediments "at least on the time scale of years to centuries." Using Chesapeake Bay, he concluded that over 15 percent of the annual sediment input from the Susquehanna River and other sources was deposited on the marshes. A review of the Chesapeake Bay sediment budget by Stevenson and others (1988) indicated that Nixon's estimate was high and that the marshes trapped only 5 to 11 percent of the total sediment input. Citing other evidence, Stevenson and others (1988) concluded, "although the concept that marshes act as major sediment sinks may be accurate over the last few millenia, tidal transport studies suggest considerable variability with most marshes presently exporting material on an annual basis." These authors suggest that estuarine sedimentation occurs principally in subtidal flats, which are below the limit of emergent marsh vegetation. Stevenson and others (1988) further concluded that (with respect to north Atlantic, south Atlantic, and Gulf coast marsh systems) southern marshes are apparently more susceptible to erosion and export of materials than northern marshes because of (1) differences in tidal dynamics, (2) seasonal changes in sea levels, and (3) higher temperatures. Another critical factor in the erosion and export processes, they hypothesized, is the reduction of the terrigenous sediment supply from the rivers along the southern U.S., which has led to sediment starvation and undernourishment of the coastal wetland systems over the last half century (Fig. 54).

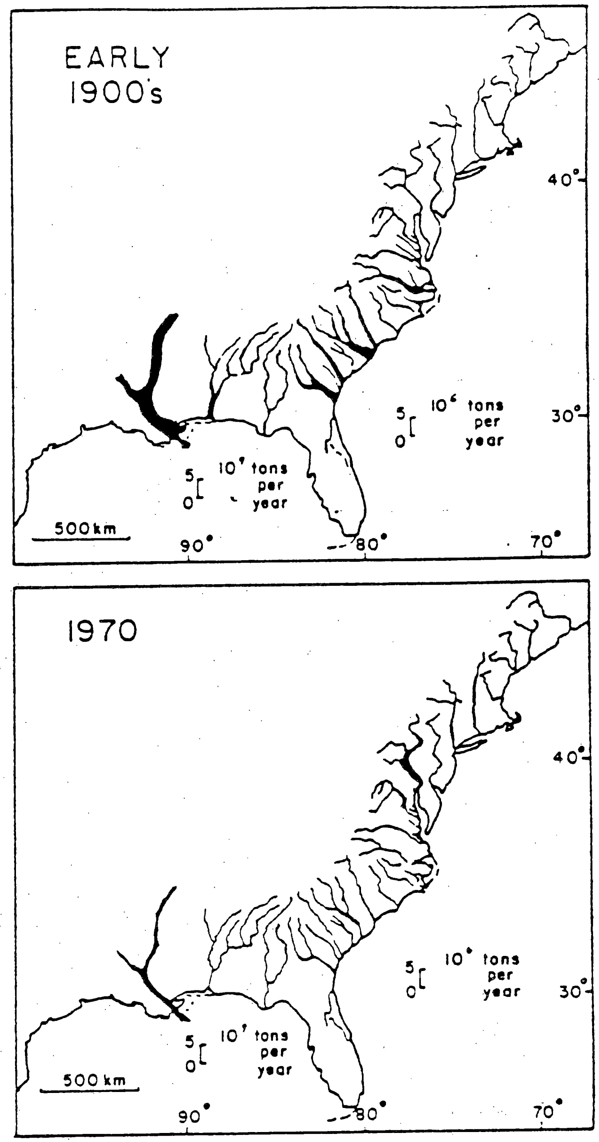


Figure 54. Comparison of sediment delivery to nearshore areas along the eastern U.S. and Mississippi Delta coasts between the early-1900's and 1970. Along the Gulf Coast the scale is in 10^9 tons/yr, and along the Atlantic Coast 10^6 tons/yr. (From Stevenson and others, 1988; reprinted by permission.)

Biodeposition

Biodeposition is an important part of sedimentation in marshes. Smith and Frey (1985) have proposed that suspension-feeding invertebrates, which actively filter suspended particles from flood waters and bind and deposit the particles on the marsh surface, may be responsible for much of the net sedimentation in the marsh. Feces and pseudofeces of these invertebrates help stabilize the sediment surface by forming large aggregates of silt and clay-size particles (Haven and Morales-Alamo, 1968), providing the marsh with a means of retaining nitrogen, phosphorus, and trace elements. The invertebrates also recycle organic materials within the detrital food chain (Kraeuter, 1976): ingestion and digestion of sediment by organisms changes the apparent grain-size and organic and inorganic chemistry of the sediment (Carney, 1981).

Many suspension-feeding invertebrates, such as oysters, barnacles, tunicates, and copepods, ingest large quantities of small particles in the 1- to 5-micron range, and after passage through the digestive tract, the particles are voided into the water as fecal pellets that range in length from 50 to 3,000 microns (Haven and Morales-Alamo, 1972). The upper size limit of particles ingested by suspension feeders is generally smaller than that of deposit feeders (Rhoads, 1974). In many molluscan species, some particulate matter is rejected before ingestion and ejected from the shell cavity in a loosely compacted mass termed pseudofeces (Haven and Morales-Alamo, 1972). Fecal pellets settle to the bottom at a faster rate than that of the particles originally in suspension, and they may be resuspended or mixed into bottom deposits by deposit feeders. Pellets and their fine component particles may accumulate in bottom areas where they would not normally settle out because of differences in hydrodynamic properties between the pellets and particles (Haven and Morales-Alamo, 1968). Feces and pseudofeces that settle to the bottom are termed biodeposits. The entire complex process, involving many groups of animals and physical and chemical factors, is termed biodeposition (Fig. 55) (Haven and Morales-Alamo, 1966).

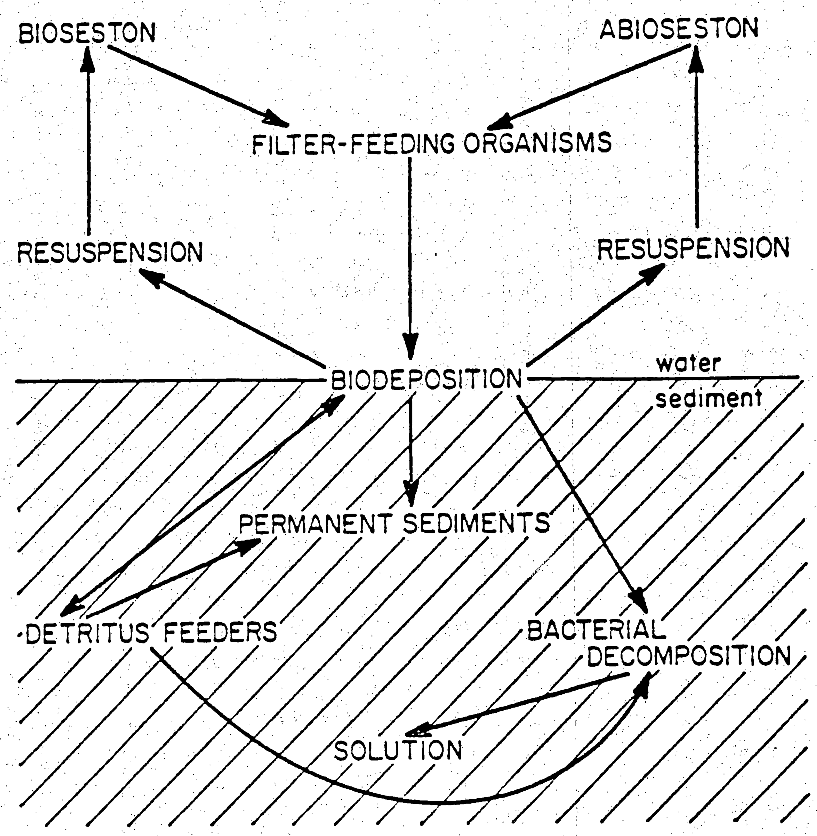


Figure 55. Theoretical biodeposition cycle in an estuary. (Modified from Haven and Morales-Alamo, 1966; reprinted by permission.)

Suspension feeders and deposit feeders may produce large quantities of biodeposits. For example, Rhoads (1974) reports that the top centimeter of mud in Buzzards Bay, Massachusetts, is almost entirely pelletal because of reworking of the mud by the polychaete *Clymenella torquata* and the bivalves *Yoldia limatula* and *Nucula annulata*. Laboratory studies indicate that oysters on 0.405 hectare of an estuarine bottom may produce up to 981 kg of feces and pseudofeces (Haven and Morales-Alamo, 1966). Lund (1957) calculated that if oysters covered 1 acre of bottom, they would deposit 8.36 tons of fecal material (dry weight) or 6 tons of dry mineral matter in 11 days. Rates of biodeposition of mud by some suspension and deposit feeders found on the Texas coast are shown in table 10.

Suspension-feeding mussels, such as *Geukensia demissa,* may ultimately enhance the productivity of a salt marsh by filtering particulate nitrogen and depositing it as feces and pseudofeces in the marsh sediment. The major role of the mussels in nitrogen flow in a marsh is to increase retention of nitrogen within the marsh by filtering particulate nitrogen from suspension (Jordan and Valiela, 1982). Jordan and Valiela (1982) report that the entire population of *Geukensia demissa* in a New England salt marsh is capable of filtering a volume of water in excess of the tidal volume of the marsh during each tidal cycle in the summer. Yearly, the mussels filter 1.8 times the particulate nitrogen exported from the marsh by tidal flushing (Jordan and Valiela, 1982). Half of this nitrogen is absorbed by the mussels and half is deposited as feces and pseudofeces. Since nitrogen limits productivity, increased retention of nitrogen may increase productivity of the marsh.

Deposit feeders probably play a more quantitatively significant role in "pelletizing" a muddy sea floor than do suspension feeders (Rhoads, 1974). The deposit-feeding polychaetes *Clymenella torquata* and *Pectinaria gouldii* can rework mud at rates of 96 to 400 ml wet mud/individual/yr (Rhoads, 1974). The rates become more impressive when the individual rates are multiplied by the standing crops found in some temperate and boreal bays, which commonly range from 100 to 10,000 individuals/m^2 (Rhoads, 1974).

Table 10. Rates of biodeposition of mud by some suspension and deposit feeders on the Texas coast. (Modified from Rhoads, 1974; Kraeuter, 1976.)

Species	Deposition (mg dry weight of feces per animal per day)
Littorina irrorata (gastropod)	0.4
Crassostrea virginica (bivalve)	18.0
Geukensia demissa (bivalve)	13.0
Balanus eburneus (barnacle)	0.4
Molgula manhattenensis (tunicate)	64.0
Clymenella torquata (polychaete)	90.0
Pectinaria gouldii (polychaete)	164.0

Biodepositional rates vary seasonally. Temperate estuaries are subjected to hot summers and cold winters, requiring organisms to undergo seasonal metabolic changes that directly affect the organisms' feeding and filtration rates (Biggs and Howell, 1984). During periods of extreme heat or cold, organisms will reduce their filtering rates to a minimum or totally "shut down" (Biggs and Howell, 1984). Smith and Frey (1985) studied biodeposition by the bivalve *Geukensia demissa* in a salt marsh and found that the following variables influenced seasonal biodepositional rates: (1) mussel density; (2) length of tidal inundation (feeding time or time available for biodeposition), and (3) individual mussel biodepositional rates (g/mussel/hr). Individual mussel biodepositional rates are in turn affected by mussel size, water temperature, and amount of material suspended in water (Smith and Frey, 1985).

Bioturbation or sediment reworking in the form of burrowing, tube building, and the production of sediment-binding exudates, along with biodeposition, are known to influence the fate and transport of sediment (Schaffner and others, 1987). These processes are, in turn, affected by sediment accumulation rates. Areas of rapid deposition (>3 cm/yr) exhibit little evidence of bioturbation, as do areas where erosion dominates (Schaffner and others, 1987). Areas with low sediment accumulation rates (0.5 to 3.0 cm/yr) exhibit the highest levels of mixing.

Texas Deltaic Marshes—Development and Current State

This literature synthesis deals principally with fluvial sediments and their role in marsh development and maintenance; thus, the following discussion focuses on deltaic areas, which are the loci of fluvial sediment deposition, and the sites of extensive marshes. As previously discussed, several rivers in Texas have constructed deltas in bay–estuary–lagoon areas (Fig. 1). Deltaic areas studied most extensively are those at the mouths of the Colorado (Wadsworth, 1941, 1966; Kanes, 1965, 1970; Manka and Steinmetz, 1971), Guadalupe (Donaldson and others, 1970), and Trinity Rivers (McEwen, 1969). Comparison of the size and shape of these deltas with others is shown in figure 56. Aggradation rates of marshes that have developed in deltaic areas have apparently not

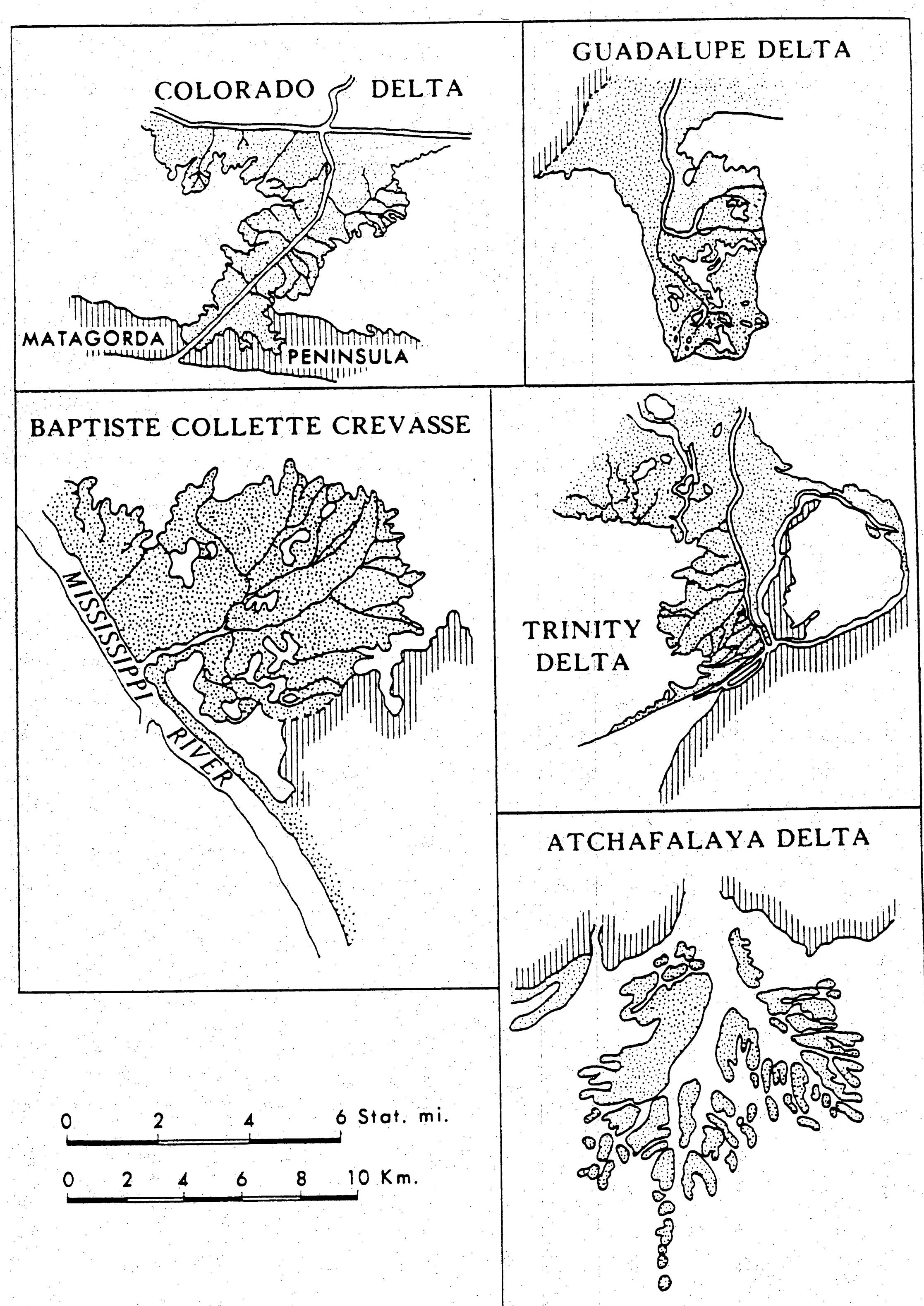


Figure 56. Comparison of size and shape of the Trinity, Guadalupe, and Colorado River deltas with the Atchafalaya River delta and crevasse-splay deposits of the modern Mississippi River. (From Cunningham, 1981.)

been previously investigated in Texas, but are currently under investigation in two areas by White and Calnan (1989).

Colorado River Delta

The Colorado River delta is the site of extensive salt-water marshes (McGowen and Brewton, 1975; McGowen and others, 1976a; Benton and others, 1977; Adams and Tingley, 1977; Ward and others, 1980; van Beek and others, 1980; White and others, 1988). In 1976, the delta was approximately 3,310 hectares (8,175 acres) in size, and about 80 percent was composed of marshes, shallow water bodies and channels, and intertidal flats; on the remaining part (20 percent) of the delta were spoil deposits, natural levees, man-made structures, and shell ridges (van Beek and others, 1980). The historical development of the Colorado River delta has been discussed by several authors, including Wadsworth (1941; 1966), Kanes (1965; 1970), Bouma and Bryant (1969), Manka and Steinmetz (1971), McGowen and Brewton (1975), USACE (1977; 1981), Ward and others (1980), and van Beek and others (1980).

The Colorado River delta has had a relative brief history of development after removal of a log raft that had trapped large amounts of sediments along its lower reaches. Removal of the raft in 1929 led to rapid progradation of the delta across the eastern arm of Matagorda Bay (approximately 6 km [3.7 mi]) between 1929 and 1935 (Wadsworth, 1966). In 1936, a channel was dredged across Matagorda Peninsula, allowing the river to discharge directly into the Gulf of Mexico. The successive stages of growth of the delta are depicted in Figure 57. Between 1929 and 1941, the average rate of growth of the delta was approximately 203 hectares/yr (500 acres/yr) (Wadsworth, 1966; Fig. 58).

Kanes (1965; 1970) conducted a detailed geological study of different environments of deposition in the northeast lobe (Egret Island) of the Colorado River delta. His investigation, which was based on almost 40 cores, revealed a platform of deltaic sediments 2.4 to 3 m (8 to 10 ft) thick. Depositional environments he delineated are delta-plain clay and silt (0.3 to 1.5 m [1 to

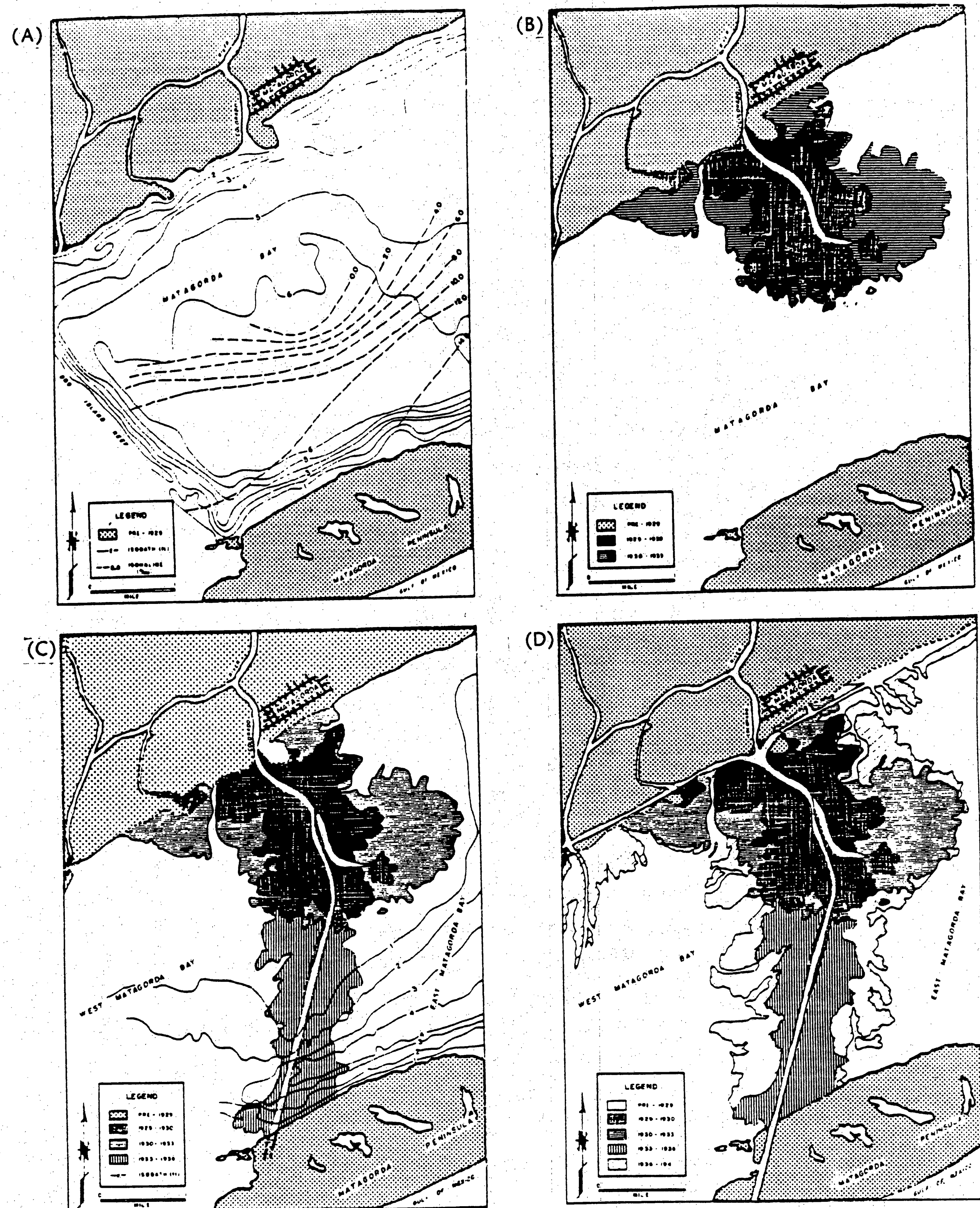


Figure 57. Historical development of the Colorado River delta between 1908 and 1941. (A) depicts pre-1929 environments; (B), pre-1929 to 1933; (C), pre-1929 to 1936; and (D), pre-1929 to 1941. Bathymetry in (C) dated 1934. (From Manka and Steinmetz, 1971; growth of the delta after Wadsworth, 1966; bathymetry after Kanes, 1965; salinity after Galtsoff, 1931; reprinted by permission.)

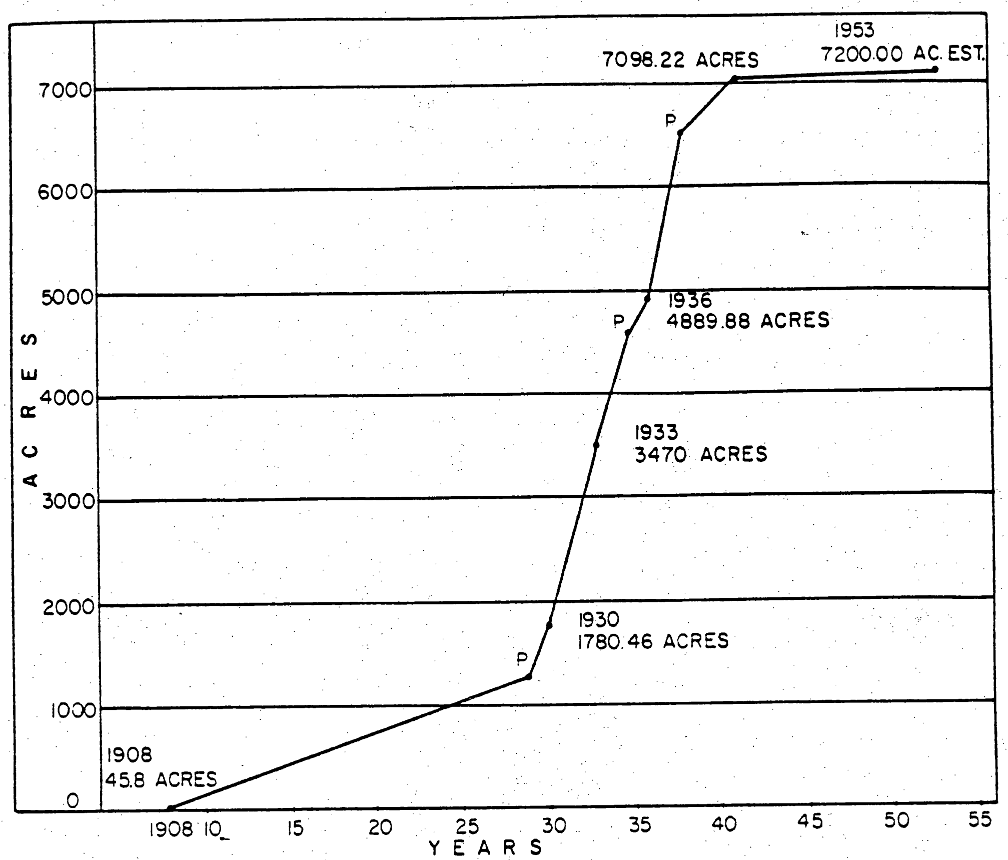


Figure 58. Increase in area of the Colorado River delta for the period 1908–1953. Postulated flood years are marked by "P". (From Wadsworth, 1966.)

5 ft] thick) , delta-front sand (0.6 to 2.4 m [2 to 8 ft] thick), prodelta silty clays (0.3 to 1.5 m [1 to 5 ft] thick), and bay silty clays and clayey silts (3 to 4.2 m [10 to 14 ft] thick). Depositional environments are similar to those of other deltas (Fig. 45). The delta plain is subdivided into low and high marshes, beach ridge, and upper and lower distributary channel. The high marshes included natural levees.

Kanes (1970) found that the early prodelta deposits of the river are fine-grained laminated sediments that lacked sand and silt. He attributed the lack of a coarser fraction to a logjam along the river, which trapped the sand and silt upstream. Removal of the logjam led to the development of a delta-front sheet sand during an initial phase of deposition, followed by a second phase in which sand was deposited in prograding distributary-mouth bars, forming bar fingers or a digitate outline (Kanes suggested that the pattern associated with this second phase of sand deposition may have been due to diversion of the river through a small manmade channel). Finer-grained sediments were deposited in interdistributary areas. High and low marshes of delta-plain environments formed on the subaerial delta lobe. Low marsh sediments were commonly finer grained and contained more laminations than the higher marsh sediment, which suggested deposition of fine particles from turbid bay waters in low marshes (Kanes, 1970).

Manka and Steinmetz (1971) investigated the depositional history of the delta's southeast lobe (gulfward of the lobe studied by Kanes). A three-dimensional reconstruction of the depositional history was based on 21 cores (the locations of 11 cores are shown in Figure 59). Cross sections of this part of the Colorado River delta indicate thickness of approximately 2.1 m (7 ft) (Fig. 60). Manka and Steinmetz agreed with Kanes in suggesting that the constructional phase of deltaic sedimentation (Scruton, 1960) ended in 1941 with the closing of all distributaries; they further suggested that the closing of the distributaries corresponded to the construction of the farm road along the eastern bank of the river channel. They concluded that the entire delta had entered a destructional phase (Scruton, 1960) in 1941. Manka and Steinmetz (1971, p. 322) felt "that if the Colorado River discharge channel had not been dredged into the Gulf of Mexico, and if flow and

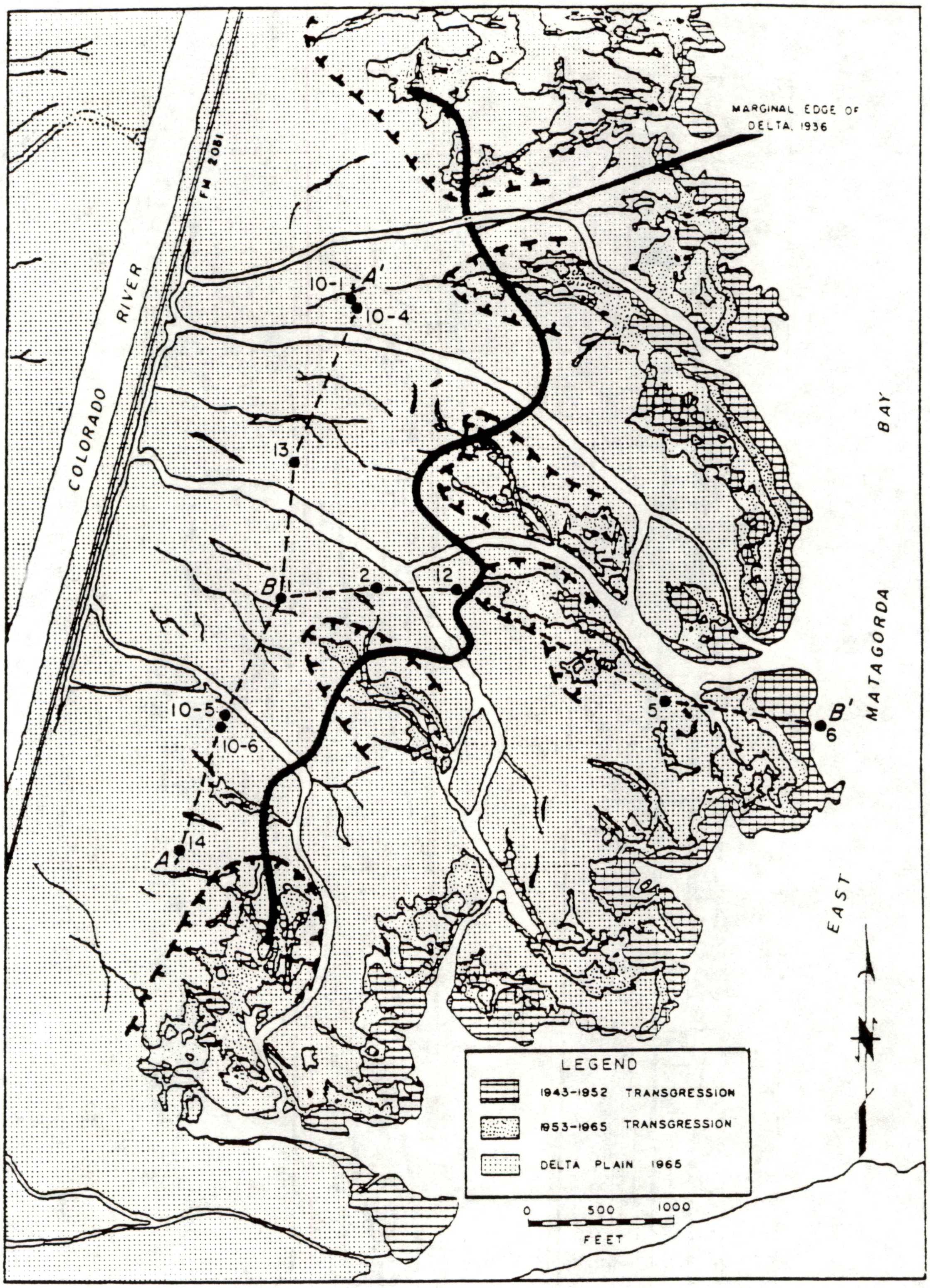


Figure 59. Deterioration or destruction of the southeast lobe of the Colorado River delta plain, as determined by comparing shoreline positions on aerial photographs taken in 1943, 1952, and 1965. Transgressive units indicate retreat of the shoreline as interdistributary bays replace marshes. Hachures indicate areas where interdistributary bays are developing. Solid circles are the locations of selected cores taken by Manka and Steinmetz. (Modified from Manka and Steinmetz, 1971; reprinted by permission.)

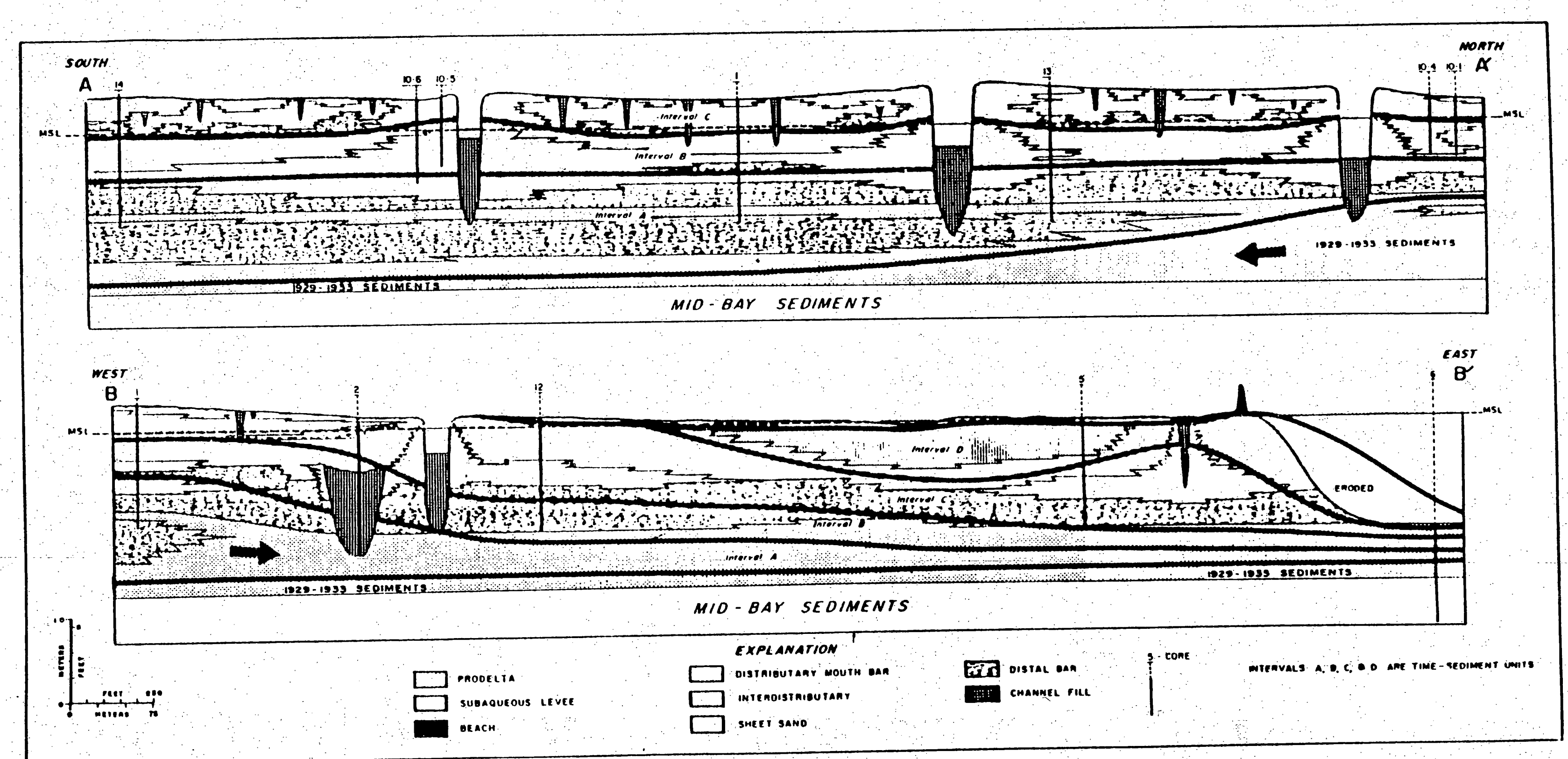


Figure 60. Cross sections of the southeast lobe of the Colorado River delta. Lines of sections are shown in figure 59. (From Manka and Steinmetz, 1971; reprinted by permission.)

sediment load were not now regulated by dams upstream, most, if not all, of the eastern half of Matagorda Bay now would be filled with deltaic sediments."

The progradation of the Colorado River produced some exceptionally high rates of bay filling near the river's mouth. The delta lobe that formed between 1929 and 1930 prograded about halfway across the bay (Fig. 57 A and 57 B). The exact depths of the bay in 1929 are not known; the bathymetry shown in Figure 57 A is for 1859–1872 (Kanes, 1970). Kanes noted that bathymetric surveys taken in 1904 and 1906 were questionable, but the surveys suggest that the bottom was shallower than shown by the 1800's survey. Assuming the depths were about 0.6 to 1 m (2 to 3 ft)[1], this indicates a sedimentation rate of about 30 to 50 cm/yr (1 to 1.5 ft/yr) (deposition of 0.6 to 1 m [2 to 3 ft] of sediment in 2 yr). This rate is similar to the sedimentation rate of the prograding Mississippi River of 300 m (1,000 ft) per 1,000 yr (Shepard, 1953; Scruton, 1960; Rusnak, 1967). However, estimates of rates of sedimentation of the Colorado River delta are considerably less if bathymetric data presented by van Beek and others (1980) are used. Changes in bathymetry in the area of the Colorado River delta between 1921 and 1935 reported by van Beek and others indicate a maximum of 1.2 m (4 ft) of sediment accumulation. This yields an aggradation rate of approximately 8 to 9 cm/yr (3 to 3.5 in/yr) for this 14-yr period. If, however, we assume that most of this change in depth occurred between 1929 and 1935, then the rate increases to 20 cm/yr (8 in/yr) (1.2 m [4 ft] in 6 yr). This assumption may be incorrect, however, because historical data indicate that the growth of the delta had begun as early as 1921 (van Beek and others, 1980).

Sedimentation rates can also be roughly calculated from analysis of time-sediment intervals defined in the southeastern lobe of the Colorado River delta by Manka and Steinmetz (1971, Fig. 60). Assuming that sequences A, B, and C (Fig. 60) were deposited between 1933 and 1936 (the postulated year in which interval D began to accumulate), a sedimentation rate of about 40 cm/yr (1.3 ft/yr) is derived (approximate total thickness of these units is about 1.2 m [4 ft] in core 5). This is similar to the 30 to 50 cm/yr (1 to 1.6 ft/yr) rate calculated for the northeastern lobe over a

[1]Galtsoff (1931), who conducted an oyster survey in this area in 1926, reported that depths decreased toward the head of the bay, and that channel depths were less than 0.9 m (3 ft).

2-yr period (see preceding paragraph). Manka and Steinmetz assumed that interval C was deposited during the 1935 flood in which river discharge reached 177,000 cfs (maximum recorded between 1916 and 1950; average discharge during this 34-yr period was 3,609 cfs). Interval C had a thickness of 0.5 m (1.7 ft) in core #13 (Manka and Steinmetz, 1971, p. 314). This indicates sediment accumulation rates of more than 50 cm (1.6 ft) as a result of a major flood event. Although Manka and Steinmetz presented no accumulation rates, with the exception of 10 cm (4 in) of sand that was postulated to have been deposited during Hurricane Carla in 1961, they do note that sedimentation rates decreased after 1936, when river discharge was diverted into the Gulf. If one assumes that interval D (Fig. 60) was deposited between 1936 and 1941, as implied by Manka and Steinmetz, then sedimentation rates of about 10 to 15 cm/yr (4 to 6 in/yr) are derived. Using the thin sheet sand that Manka and Steinmetz postulated was deposited during Hurricane Carla as an upper time line and the 1933 line as depicted in Figure 60 as the lower time line, approximately 1.8 m (6 ft) of sediment was deposited during this 28-yr period, indicating "longer term" average sedimentation rates of about 6.4 cm/yr (2.5 in/yr).

Rates of sediment accumulation near a prograding delta greatly exceed rates of accumulation in bay and estuary environments away from the delta (Shepard, 1953; Rusnak, 1967). Van Heerden and others (1981) found that as much as 1.5 m of sediment was deposited on levees of the Atchafalaya Delta during a single flood event. As stated by Scruton (1960, p. 82), "relative rapid deposition is the fundamental characteristic of deltas."

The Colorado River delta is somewhat unique in that the rapid progradation was primarily a result of the large sediment volume carried by the river (after removal of the log raft) and the shallow basin into which it discharged (Kanes, 1970; McGowen and Brewton, 1975). Storms punctuated the transportation and deposition of the sediments (Fig. 58). The rate of aggradation apparently diminished more and more in deltaic environments that were approaching sea level and becoming subaerial. Assuming that delta deposits depicted in Figure 57 D aggraded from 0.3 to 1 m (1 to 3.3 ft) (estimate based on selected bathymetric data shown in Fig. 57 C) between 1936 and 1941, the respective rates would be in the range of 6 to 20 cm/yr (2.4 to 7.9 in/yr). The higher value is equivalent to Rusnak's (1967) estimated long-term average rates of sediment accumulation of 200 m per 1,000 yr in river deltas. Where the Colorado River delta has become subaerial and marshes have developed, the rates are substantially less, perhaps in the range of 1 cm/yr (0.4 in/yr) or less (White and Calnan, 1989).

Recent Changes in State of the Colorado River Delta

The end of the constructional phase of the Colorado River delta in 1941 apparently can be attributed to several factors, including discharge of the river into the Gulf and closure of distributary channels. But, as indicated by Ward and others (1980), the year 1940 marks the period of reservoir regulation on the Colorado River. The Highland Lakes chain, which intercepts sediments from about 90 percent of the Colorado River drainage basin (Kanes, 1970; USACE, 1977), has decreased the sediment supply at the mouth of the river to a fraction of the previous load (Figs. 25 and 36). Ward and others (1980) conservatively estimated that the mean sediment load of the Colorado River is an order of magnitude below that which was responsible for the relatively constant rate of delta growth of approximately 220 ha/yr (550 acres/yr) from 1929 to about 1938.

Dredging of Tiger Island Cut in the 1950's led to the development of a subdelta that prograded into Matagorda Bay at a rate of about 8.4 m/yr (28 ft/yr) (McGowen and Brewton, 1975). Comparisons of aerial photographs taken in the mid-1950's with those taken in 1979 show extensive marsh development on the Tiger Island Cut subdelta (White and others, 1988; Fig. 61). The growth of the delta has been attributed to both river and tidal flow (USACE, 1977). Sand that has contributed to this delta is apparently derived from the Gulf and not the river (van Beek and others, 1980; White and others, 1988; White and Calnan, 1989). Model studies by the Corps of Engineers in 1959 indicate that tidal currents through Tiger Island Cut reach 1.2 m/sec (4 ft/sec); the tidal flow is flood dominated because tidal levels in the Gulf are higher than in the bay approximately 90 percent of the time; this promotes salt water flow into Matagorda Bay except during high river discharges (USACE, 1977). A significant amount of the Colorado River flow enters Matagorda Bay through Tiger Island Cut (USACE, 1970; TDWR, 1978). Another subdelta that

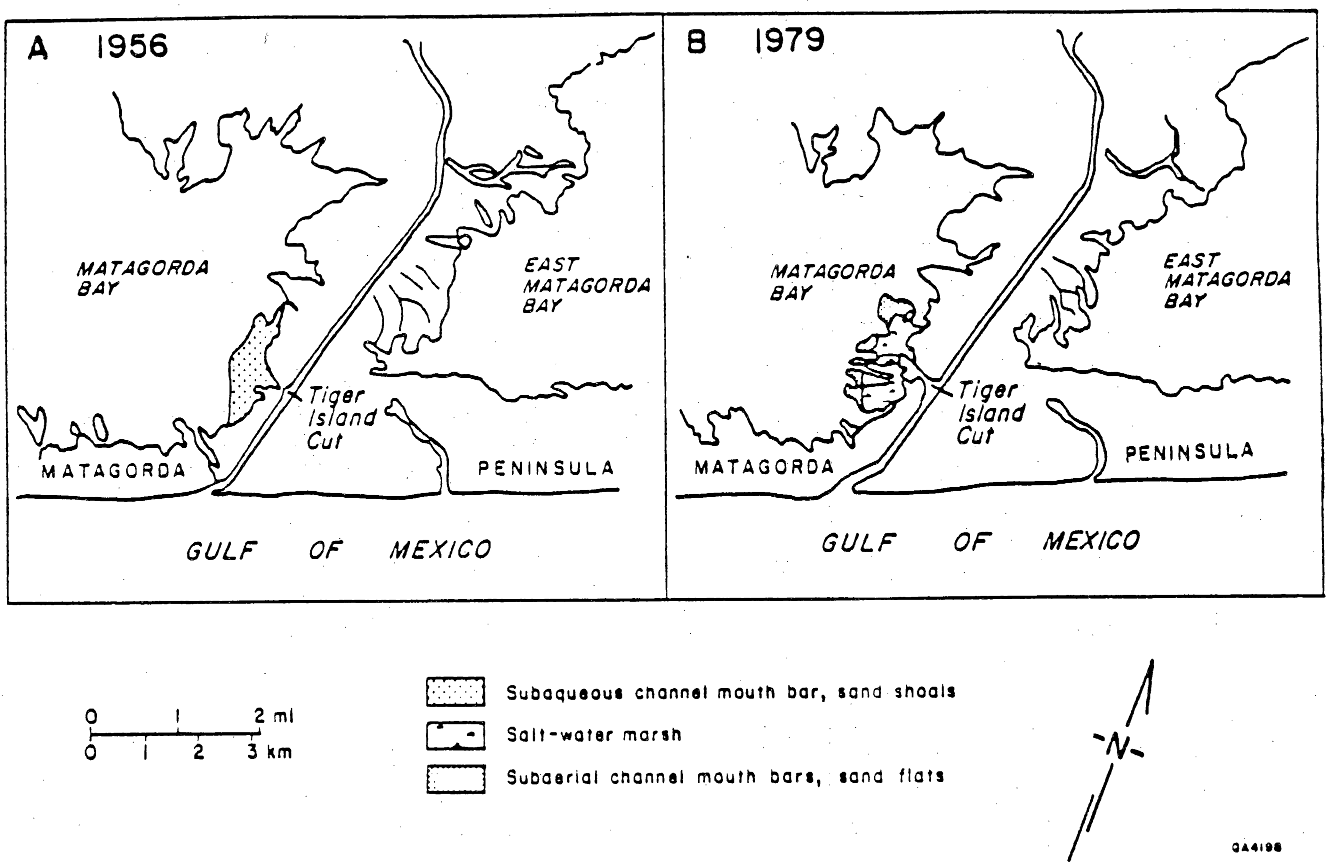


Figure 61. Changes in marsh distribution between 1956 and 1979 on the Colorado River delta at Tiger Island Cut. (From White and others, 1988.)

has grown in area since 1940 is located at Culver Cut (near the Intracoastal Waterway northwest of Tiger Island Cut); van Beek and others (1980) suggest that this area is supplied by sediment that is transported down the Intracoastal Waterway and out Culver Cut.

Manka and Steinmetz (1971) traced the destruction of the southeast delta lobe using aerial photographs taken in 1943, 1952, and 1965. These investigators reported that not only was the margin of the delta receding, or eroding due to wave attack, but compaction of underlying interdistributary sediments was promoting submergence of the delta plain and the formation of water bodies (interdistributary bays) that were enlarging through time (Fig. 59). Kanes (1970) estimated that 30 to 60 cm (1 to 2 ft) of subsidence, and locally as much as 1.2 m (4 ft), had occurred in the northeast lobe of the Colorado River delta as a result of compaction. Historical shoreline changes along the margin of the Colorado River delta determined by McGowen and Brewton (1975) indicated that the eastern margin of the delta (along East Matagorda Bay) was eroding at a rate of 1.8 to 2.4 m (6 to 8 ft) per year; the western delta (with the exception of progradation at the mouth of Tiger Island Cut) remained relatively unchanged during the period 1957–1972.

As mentioned previously, the Colorado River has discharged into the Gulf since 1936. However, sediment supply has not been sufficient to prevent erosion of the Gulf shoreline near the mouth of the river (Morton and others, 1976). Sediment is transported alongshore by littoral drift, which under the influence of dominant southeast winds is from northeast to southwest along Matagorda Peninsula (Morton and others, 1976; Morton, 1977).

Van Beek and others (1980) investigated the potential of the Colorado River to construct a delta if diverted into Matagorda Bay (USACE, 1981). Their predictive models indicate that with full diversion, a delta will prograde into the eastern arm of Matagorda Bay at a rate of about 25 ha/yr (37 acres/yr). Of their estimated 1,370 acre-ft of sediment volume supplied by the river annually, 7.7 percent would be retained in the delta (van Beek and others, 1980). Van Beek and Meyer-Arendt (1981) reported that for Atchafalaya Bay, a comparison of sediment supplied with that deposited over a 10-yr period (1967–1977) shows a retention of less than 30 percent, which is considerably below the 70 percent average calculated for the Mississippi River subdeltas.

The textural composition of the sediment delivered by the Colorado River is primarily suspended load composed of silt and clay (USACE, 1977). The Corps of Engineers (USACE, 1977) estimated that of the 1,650 acre-ft of sediment transported annually by the Colorado River, about 300 acre-ft is coarser bed load. Most of the bed load, or sand, is trapped up stream in a shallow draft navigation channel and silting basin that is periodically dredged. Welborn and Andrews (1980) investigated sediment discharge of the Colorado River at the town of Matagorda by collecting data from two floods and found that only 2.1 percent of the total load was bed load. Because little if any sand makes it to the mouth of the river, the textural composition of facies of the subdelta that forms at the mouth of the diversion channel will be considerably different from that in facies in the existing delta. In other words, there will be a fining of those facies (delta front and distributary channels) that commonly are rich in sand (Kanes, 1970; Manka and Steinmetz, 1971). Gosselink (1984) noted that reservoir construction in the drainage basin of the Mississippi River has presumably depleted the river's supply of sand, which is the main foundation material for growth of the delta; he suggests that the river, therefore, cannot support as large a delta as it has historically. As mentioned above, the Colorado River delta is primarily in a destructional stage.

Guadalupe River Delta

The Guadalupe River delta, located at the head of San Antonio Bay, is characterized by extensive brackish- to fresh-water marshes (McGowen and others, 1976b; Benton and others, 1977; White and others, 1985). Donaldson and others (1970) conducted a thorough geological investigation of the delta. Similar to the Mississippi Delta, the Guadalupe River delta, on a much smaller scale, has had a history of delta lobe growth, abandonment, and deterioration (Figs. 50 and 62). Donaldson and others (1970, p. 108) stated: "In a general sense, the Guadalupe delta represents a model of sedimentation characterized by a stream depositing its load in a shallow,

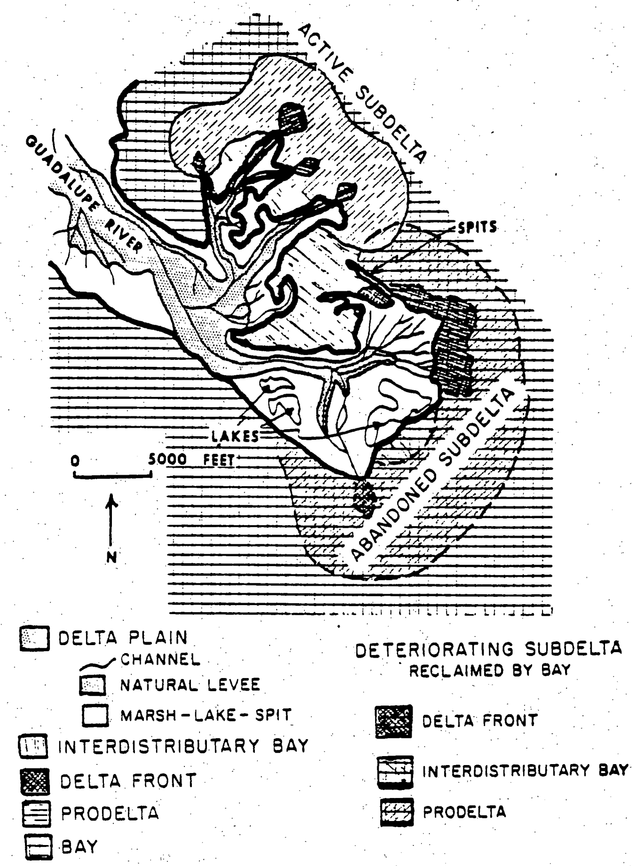


Figure 62. Map showing idealized spatial relationship of depositional environments of the Guadalupe River delta prior to formation of the Traylor Cut subdelta. (From Donaldson and others, 1970.)

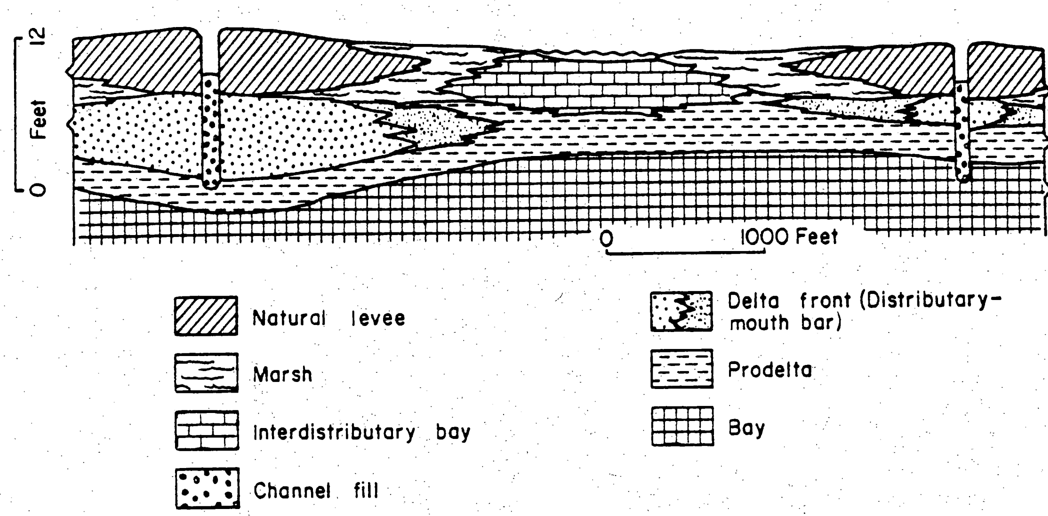


Figure 63. Generalized cross section of the Guadalupe delta showing facies. (From Donaldson and others, 1970.)

relatively quiet body of water. The delta progrades into increasingly shallow water so that the distributary channels have become deeper than the bay floor."

It is estimated that the delta began forming about 2,000 yr ago in water slightly deeper than present (Shepard and Moore, 1960). Based on this date, the average rate of progradation into the bay was approximately 12 m/yr (40 ft/yr) (Donaldson and others, 1970). Growth of the delta complex occurred in a counterclockwise direction (Fig. 50) as the major distributaries apparently migrated in response to steeper gradients in that direction. An estimated 30 cm (1 ft) of subsidence on older abandoned subdeltas resulted in the transgression of marshes and barren tidal flats over natural levees (Donaldson and others, 1970).

A cross-sectional reconstruction of the Guadalupe River delta facies reveals six major environments of deposition that are composed of distinctive sediment textures: (1) distributary channel, (2) natural levee, (3) marsh, (4) interdistributary bay, (5) delta front, and (6) prodelta (Fig. 63). Sediments of the delta front and basal part of distributary channel deposits are fine-grained sand, silty sand, or silt, while the remaining environments are composed principally of clay and silty clay (Donaldson and others, 1970). Abundant plant material occurs in the silty clay sediments of the marsh and natural levee. The configuration of the delta-front sand deposits is predominantly shoestring (barfingers), but locally, closely spaced distributary-mouth bars coalesce to form sheet sands. Average thickness of the delta (not including the channel-fill deposits, which are as thick as 4.86 m [16.2 ft]) is 2.7 m (9 ft) (Donaldson and others, 1970). The delta-plain environments, which include distributaries, natural levees, and marshes, make up the greatest volume of sediment in the delta. Thickness of the delta-plain deposits, which average 1.5 m (5 ft), decreases outward from the distributaries.

Sediment is delivered to the delta by both the Guadalupe and San Antonio Rivers, the confluence of which is less than 16 km (10 mi) upstream from the delta. Donaldson and others (1970) indicated that the average amount of suspended load transported to the delta annually is approximately 1 million tons, with nearly equal amounts of sediment contributed by the Guadalupe River (461,214 tons annually for the period 1945–1954) and San Antonio River (568,218 tons annually, 1942–1954) (Fig. 20). Textural analysis of sediments collected from the channel bed of the Guadalupe River from Victoria to its mouth indicated a substantial decrease in grain size from granules and pebbles near Victoria with changes downstream to sand and sandy silt and then to dominantly clayey silt at the river mouth (Morton and Donaldson, 1978). Composition of suspended sediment of the Guadalupe is approximately 2 percent very fine sand, 27 percent silt, and 71 percent clay (Welborn, 1967). Morton (1972) reported that clay minerals in the Guadalupe delta as well as in San Antonio Bay are predominantly montmorillonite, illite, and kaolinite, with montmorillonite most abundant. In a comparison of the suspended loads of the Guadalupe and San Antonio Rivers, Sorenson (1975) found that the suspended load of the Guadalupe is higher in organics than that of the San Antonio River. Although there is a rapid increase in inorganic sediments relative to organics during flood events, Sorenson concluded that measurements of suspended load during normal flow conditions, when organics are high, can lead to overestimates of sediment-accumulation rates in the delta. This conclusion is based on the probability that organic debris will not be preserved in substantial quantities.

Recent Changes in State of the Guadalupe River Delta

A major modification occurred on the Guadalupe delta in 1935 with the dredging of Traylor Cut, which diverted about two-thirds of the discharge of the Guadalupe River into Mission Lake (Morton and Donaldson, 1978). This diversion resulted in the rapid progradation of a subdelta into the southwestern corner of the lake. The historical development of the delta was depicted by White and Morton (1987) (Fig. 64). Donaldson and others (1970) estimated the rate of delta growth (progradation) at about 22.5 m/yr (75 ft/yr), noting that Mission Lake, which was less than 1.5 m (5 ft) deep when growth of the delta began, was less than 0.6 m (2 ft) by 1965. White and Morton (1987) reported even higher rates of progradation of the Traylor Cut subdelta during selected periods, and noted that historical progradation rates have apparently varied, in part, because of a log raft that developed along the river. Progradation rates between 1958 and 1974 were slower than after the mid-1970's when the logjam was removed. Rates of aggradation of the Traylor Cut

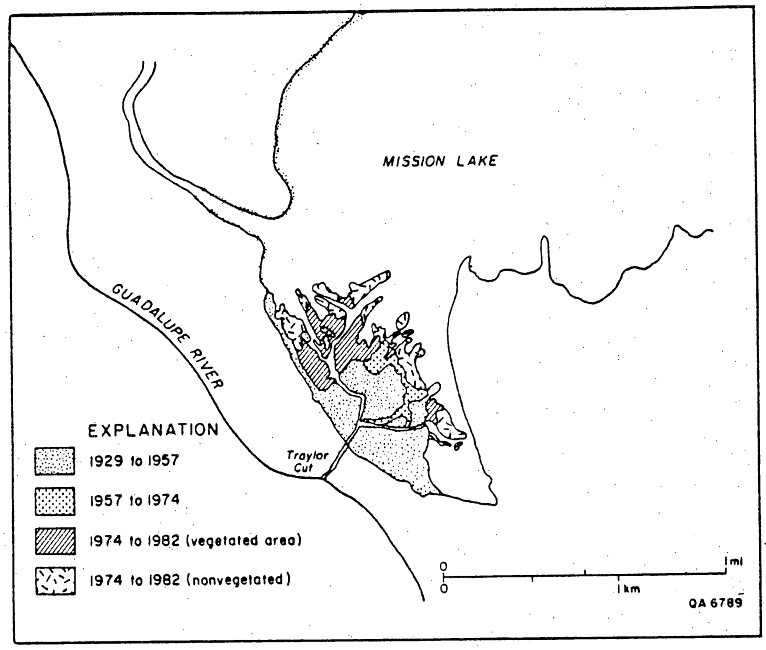


Figure 64. Shoreline changes along the Traylor Cut subdelta, based on aerial photographs, 1929 to 1982. (From White and Morton, 1987.)

subdelta can be estimated from thickness of deltaic deposits presented by Donaldson and others (1970). The cumulative thicknesses of prodelta and overlying marsh deposits in the subdelta were approximately 1.56 m (5.2 ft) thick. This thickness would yield an average sedimentation rate of about 5 cm/yr (for the period 1935 to 1965). An isopach map of prodelta deposits at the Traylor Cut subdelta just offshore from delta-plain deposits (Donaldson and others, 1970) shows a thickness of about 1.26 m (4.2 ft), which indicates an aggradation rate of approximately 4 cm/yr (0.13 ft/yr). These rates of deltaic aggradation are an order of magnitude higher than the average rate of sedimentation (shoaling) in San Antonio Bay (0.37 m/100 yr [1.23 ft/100 yr]) determined by Shepard (1953).

Diversion of river discharge and sediment load through Traylor Cut apparently has hastened the retreat of the delta shoreline along Guadalupe Bay near the mouth of the North and South Guadalupe Rivers, which were the natural, active distributaries of the delta (Donaldson and others, 1970; White and Morton, 1987). Donaldson and others (1970, p. 111–112) stated that "relatively strong tidal flow through the narrow Guadalupe Bay and reduced discharge and load have transformed the North-South Guadalupe from a prograding to a deteriorating subdelta." The shoreline along most of the delta retreated (eroded) between 1930 and 1982, at an average rate of 1.5 m/yr (5 ft/yr); rates of erosion were locally as high as 2.7 m/yr (9 ft/yr) (White and Morton, 1987). Highest rates of erosion occurred on the gulfward half of the delta, where abandoned delta lobes do not receive sufficient fluvial sediment to counteract the effects of delta subsidence and wave and current action in San Antonio and Guadalupe Bays. A comparison of photographs taken in the mid-1950's with those taken in 1979 indicates an increase in water bodies and shallow subaqueous flats (Fig. 65). It appears that under current conditions, the Traylor Cut subdelta will continue to prograde into Mission Lake, producing additional wetland acreage, while inactive parts of the delta will continue to deteriorate through erosion and submergence. (Additional data on sedimentation in this area will be presented in a later section on San Antonio Bay.)

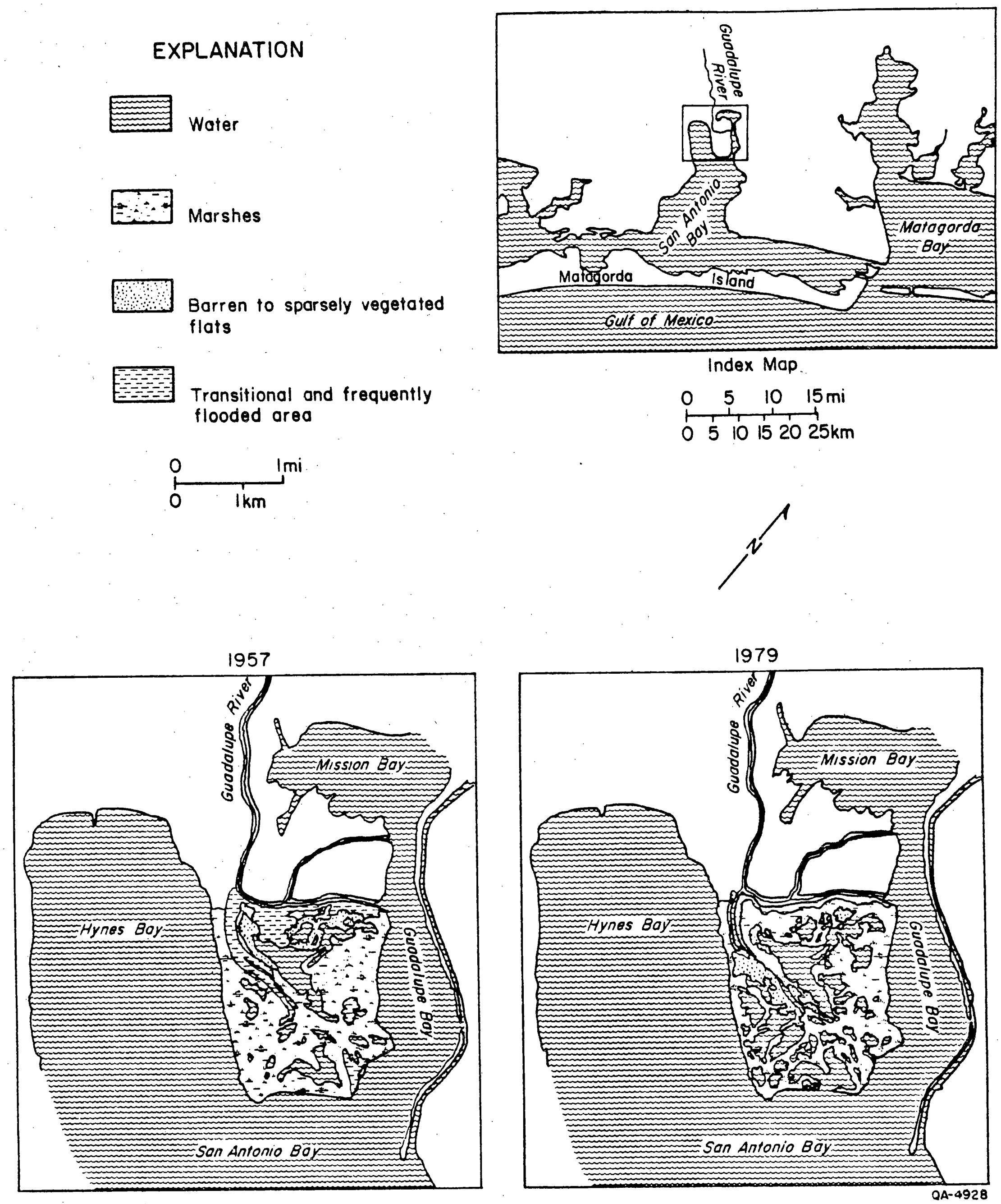


Figure 65. Changes in the distribution of wetlands between 1957 and 1979 on the seaward half of the Guadalupe River delta. (From White and others, 1989).

Trinity River Delta

The Trinity River, like the Colorado and Guadalupe Rivers, has constructed a delta at its mouth (Figs. 1 and 66). The salinity regime that characterizes this deltaic area is more similar to that of the Guadalupe River than the Colorado River, and thus, the Trinity River delta is the site of extensive brackish- to fresh-water marshes (Fisher and others, 1972; Rice Center for Community Design and Research, 1974; Adams and Tingley, 1977; Benton and others, 1979; White and others, 1985). The Trinity and Colorado Rivers are similar in that they are the only two Texas rivers (discharging into estuaries) that have appreciably extended their deltas since the mid-1800's (Shepard, 1953). Shepard reported that the Trinity delta had advanced a distance of about 0.48 km (0.3 mi) since 1855.

Sedimentary facies of the modern Trinity River delta have been investigated by McEwen (1969), from which most of the following discussion was derived. Similar to the other estuarine deltas discussed previously, the Trinity River delta has prograded into a relatively protected, shallow bay. McEwen (1969) defined the following depositional units in the Trinity River delta: (1) channel and associated deposits, (2) natural levee, (3) interdistributary bay and marsh, (4) delta-front churned sands, and (5) prodelta.

Composition of channel deposits, which include point bars, ranges from sand at the base of the channel to silty clay at the top. Sediment textures composing natural levee deposits vary from silty clay to clayey sand; plant debris and roots are common. Natural levees, which border the Trinity River and distributary channels, range from about 0.9 m (3 ft) in elevation in more landward parts of the delta, to sea level and below at the distributary mouths. According to McEwen (1969), the widths of natural levees range from a few feet along smaller distributaries to several hundred feet along the main river channel. Channel mouth bars are composed primarily of sand. Submarine levees are gradational with the channel mouth bars with sediment textures decreasing from channels toward the levee deposits. At the time of his study, McEwen reported that Jack's Pass (Fig. 66) was the only active distributary that was in a relatively natural state. Other distributaries, for example,

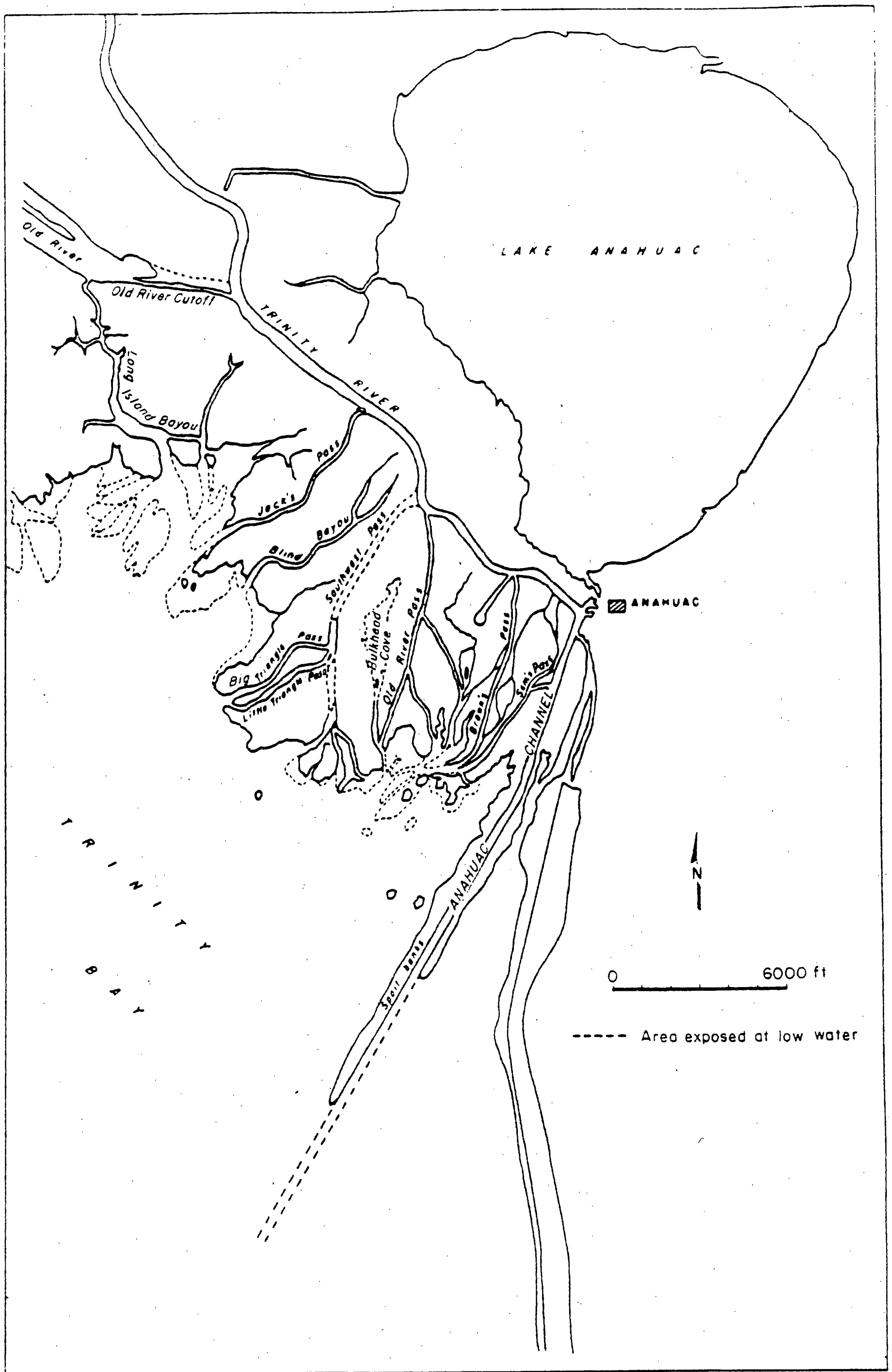


Figure 66. Trinity River delta. (From McEwen, 1969.)

Blind Bayou, Southwest Pass, Triangle Pass, and Old River Pass (Fig. 66), were abandoned and were being filled with sediments. Because the process of abandonment and filling is a gradual one, sediment characteristics are reflective of channel as well as lake and marsh deposits, and accordingly, range from sands and clays to muds.

In terms of areal distribution in the subaerial part of the delta, the marsh facies is the most important. Marsh sediment composition is predominantly organic-rich clays and sandy clays. The organics are composed of very fine grained material, as well as plant debris and rootlets. Marsh deposits with the highest sand content occur near distributaries and delta-front environments. McEwen examined sediments in Round Lake to define the characteristics of lake deposits, which he described as mostly gray to black clay with local silty burrows and lenses. The sediments could be distinguished from marsh sediments because of a complete lack of visible plant remains. Interdistributary bays are shallow environments that are bordered by distributary levees and marshes. Composed principally of clayey sand in central areas, these deposits grade laterally into submarine levee, marsh, and delta-front deposits.

The delta-front deposits (delta-front churned sands) are several feet in depth and are composed of sandy sediments that extend from the front of the delta to a distance of about 2.4 to 3.2 km (1.5 to 2 mi). The sediments decrease in grain size toward the bay, grading from well-sorted, fine sands to the almost pure clays that make up the open bay sediments. The bay or prodelta facies is predominantly black mud (McEwen, 1969).

The delta-front sands, which in most areas have been extensively churned by burrowing organisms, are volumetrically the most important depositional unit in the delta. In fact, McEwen (1969, p. 74) described the delta "as a mass of delta-front sands capped by a veneer of marsh deposits and transected by linear bodies of channel deposits." The base of the delta-front sands is about 4.5 to 4.8 m (15 to 16 ft) below sea level in the oldest part of the delta (northern part), where the sands reach a maximum thickness of between 4.2 and 4.5 m (14 and 15 ft). The unit thins to the east and southeast, apparently due to shoaling of the bay in that direction. Marsh deposits have a uniform thickness of 0.6 to 0.75 m (2 to 2.5 ft) in the older part of the delta. The base of marsh deposits, commonly about 0.3 m (1 ft) below sea level, is as much as 0.6 m (2 ft) below sea level in the central part of the interdistributary bays. These clayey deposits may reach 0.6 m (2 ft) above sea level near the Trinity River.

The oldest radiocarbon dates measured in shells (*Rangia flexuosa*) from cores in the delta-front sands were 810 and 750 yr B.P. If one assumes that approximately 5.5 m (17.5 ft) (thickness of the delta-front sands + marsh deposits) accumulated over this period, then the long-term average accumulation rate is about 7 mm/yr. Assuming that a thickness of 4.5 m (15 ft) accumulated during a period of 500 yr (Rice, 1969) will yield a rate of about 9 mm/yr (0.35 in/yr). These rates are close to the average 8 mm/yr (0.3 in/yr) sea-level rise estimated by Emery and Uchupi (1972) for the past 5,000 yr. It is presumed that sedimentation rates were high in the actively prograding parts of the delta and decreased substantially as delta surfaces became emergent and aggraded above sea level.

Rice (1967; 1969) calculated that sediment delivered by the Trinity River to the delta was made up of the following approximate textural percentages (in the suspended load): 25 percent sand, 30 percent silt, and 45 percent clay. In calculating the mass balance between the volume of sand delivered by the river and the volume composing the delta he found an imbalance. Considering the area of the modern delta to be about 6 mi^2 (9.5 km^2) (about 1,500 ha or 3,800 acres), and assuming that the delta was composed wholly of delta-front deposits with a thickness of 4.5 m (15 ft) and with a composition of 75 percent sand, Rice (1969) calculated that the delta contains a volume of 43,200 acre-ft of sand. Using the assumption that over the past 500 yr the Trinity River had a constant annual discharge of 3,000 acre-ft of sediment (based on suspended load records upstream at the Romayor station; Fig. 20) composed of 25 percent sand, Rice calculated a total volume of 375,000 acre-ft of sand. A comparison of these two volumes indicates that almost 10 times as much sand was transported by the river as is found in the delta. Rice noted that sand does not appear to be moving across the mouth of modern Trinity Bay and hypothesized that the lower portion of the Trinity River must be aggrading its valley through valley subsidence. In other words, the sand, assisted by the process of subsidence, was being deposited by the river upstream from the delta in floodplain and point-bar channel deposits.

Failing (1969), referring to the imbalance between sand delivered by the river and that found in the delta (Rice, 1969), noted that the Trinity River valley between Romayor and the delta had an area of about 800 km^2 (500 mi^2) (129,600 ha or 320,000 acres) and that sediment spread over this area to a thickness of 30 cm (1 ft) would have a volume of about 300,000 acre-ft (the approximate amount of the imbalance noted in the preceding paragraph). Failing (1969) concluded that because 30 cm (1 ft) of sediment equals the amount of subsidence that would have occurred over the past 500 yr at the rate of 6 cm/100 yr (0.2 ft/100 yr) (Rehkemper, 1969), then subsidence and sedimentation must be occurring at the same rate in the lower Trinity River valley.

McEwen (1969) made the observations that marsh deposits in the delta were thin and that no marsh or levee deposits were found at depths greater than 0.6 m (2 ft) below sea level. This led McEwen to conclude that no measurable subsidence had occurred during the growth of the delta, and there had been no relative change in elevation of the delta and sea level since delta formation began. This is somewhat puzzling, especially because of the hypothesis that subsidence was occurring along the river valley (Failing, 1969; Rice, 1969). Failing (1969) suggested, however, that if the delta was formed between 500 and 1,000 yr ago, as indicated by radiocarbon dates (McEwen, 1969), and the minimum subsidence rate is 6 cm/100 yr (0.2 ft/100 yr) (Rehkemper, 1969), then total subsidence of the delta would be on the order of 30 to 60 cm (1 to 2 ft), which may have gone undetected in McEwen's coring investigation.

For comparison, estimated compactional subsidence of parts of the Colorado River delta is on the order of 30 to 60 cm (1 to 2 ft), and locally as high as 1.2 m (4 ft) (Kanes, 1970). Donaldson and others (1970) estimated that at least 0.3 m (1 ft) of subsidence had occurred on the Guadalupe River delta as reflected in the transgression of marshes over relict levees.

Recent Changes in State of the Trinity River Delta

The Trinity River delta is one of only two Texas bay-head deltas identified by Shepard (1953) as having significantly prograded in recent history. Analysis of historical shoreline changes also indicates local delta shoreline progradation in which marshes have advanced bayward at rates of more than 1.8 m/yr (4.3 ft/yr) (and a high, locally, of 32 m/yr [103 ft/yr]) between 1930 and 1982 (Paine and Morton, 1986; Fig. 67). Older inactive parts of the delta (stations T46 to T51, Fig. 67) retreated at rates of 1.2 m/yr (3.8 ft/yr) to 3.1m/yr (10.2 ft/yr) between 1930 and 1974 or 1982. Shorelines at most stations on the active delta lobe (T45 to T51) prograded (accreted) into the bay from 1930 to 1982. The shoreline at delta station T42, however, eroded. In addition, although shoreline changes at station T43 indicated net accretion between 1930 and 1982, this shoreline began to retreat (erode) during the latter part of this period, as indicated by comparing marsh areas mapped on 1956 photographs (Fisher and others, 1972) and 1979 photographs (White and others, 1985). Comparison of marsh changes in interior deltaic areas indicates some encroachment of water into areas formerly mapped as marshes (White and others, 1985). Trends toward erosion and marsh submergence may be indicative of subsidence and reductions in sediment supplied to the delta.

Subsidence in the delta area between 1943 and 1978 was about 22.5 cm (0.75 ft) (Gabrysch, 1984). This rate, approximately 6 to 7 mm/yr (0.24 to 0.28 in/yr), is not exceptionally high. However, the quantity of fluvial sediment delivered to the delta by the Trinity River has apparently declined markedly since about 1969 when Lake Livingston was constructed (Fig. 35). Declines in sediment load of the river are treated more thoroughly in a later section (p. 179), but in general terms, recent suspended-load measurements (1969–1984) show that the average annual load of the Trinity River at Romayor (downstream from Lake Livingston) is only about 14 percent of that reported by Shepard (1953). As stated by the U.S. Corps of Engineers:

> At present, only about 2 percent of the sediment entering Lake Livingston passed through the spillway, and most sediment contributions from tributaries below Livingston dam will be inhibited by Wallisville dam. It seems likely that, in the future, aggradation and progradation of the delta will occur only during high flows, and that the primary agents in the shaping of the delta will be wind-generated wave erosion and deposition.
>
> (USACE, undated, p. A-10).

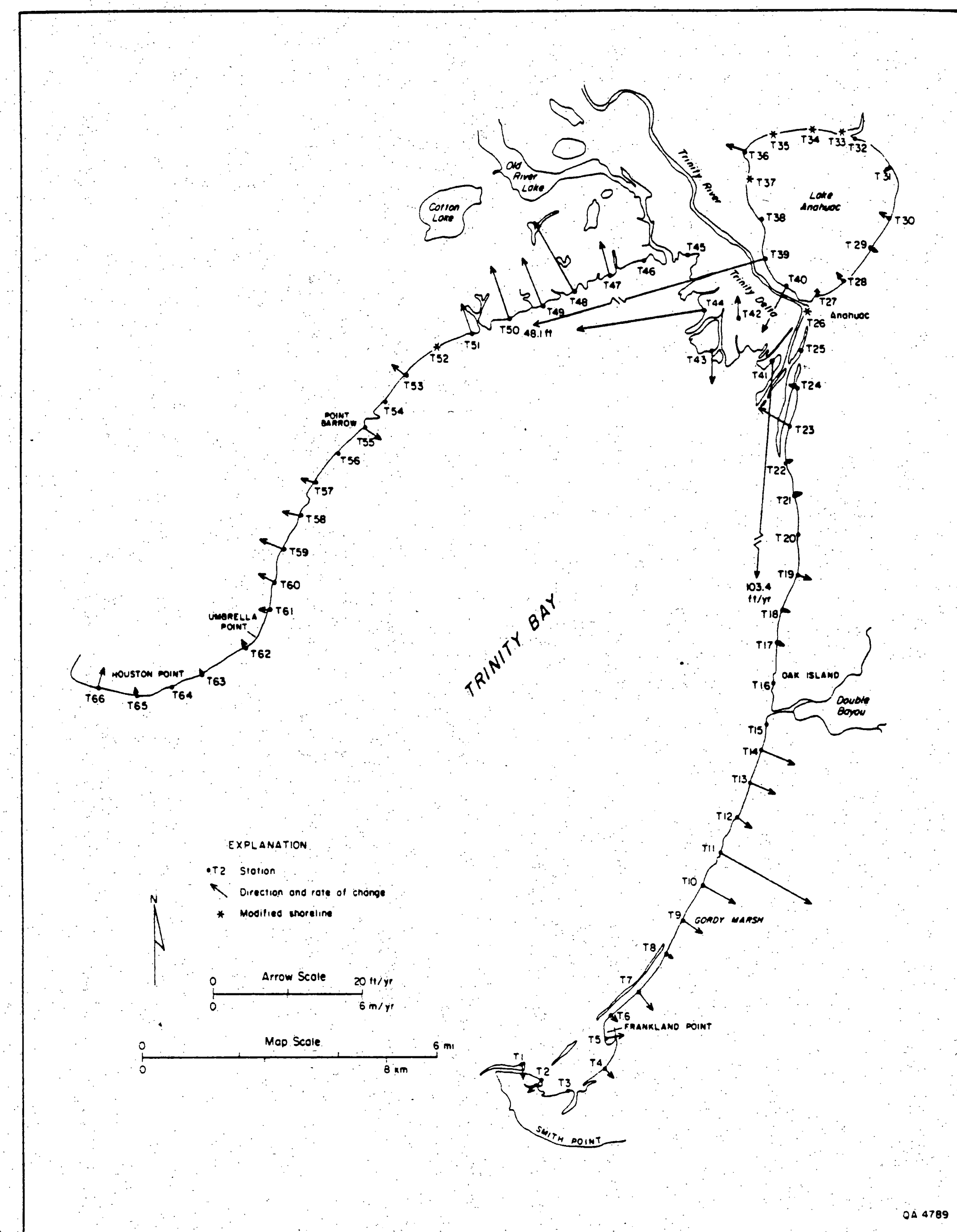


Figure 67. Rates of shoreline change along Trinity Bay and Trinity delta, 1930 to 1982. (From Paine and Morton, 1986.)

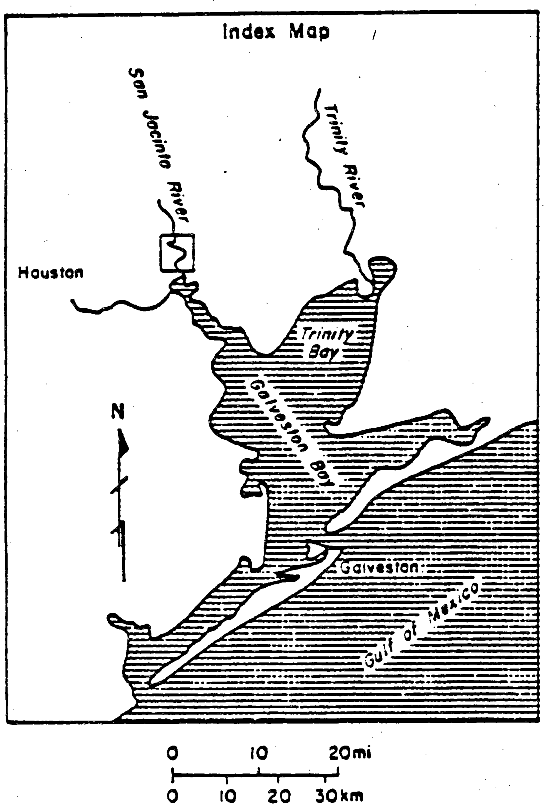
Recent Changes in State of Wetlands in Other Texas Fluvial-Deltaic Areas

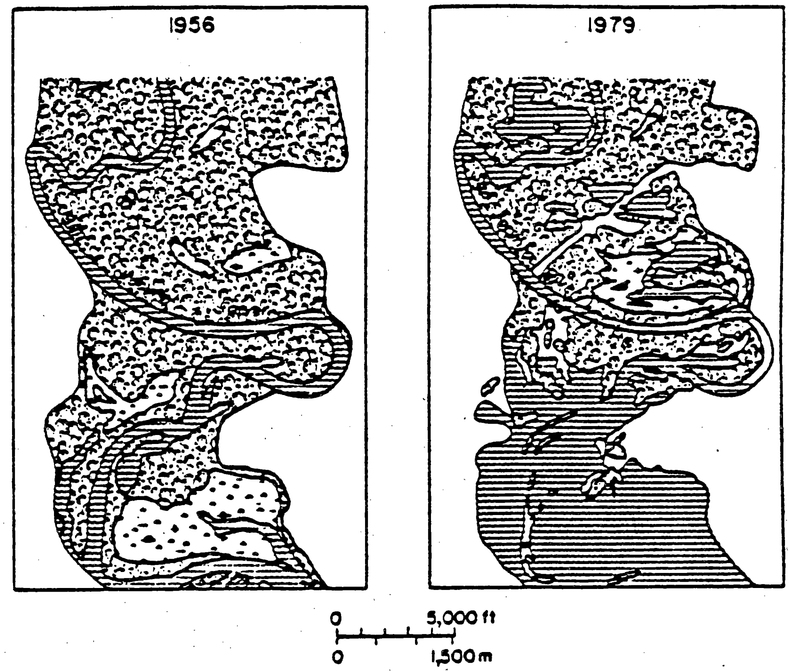
The preceding discussion on the depositional framework, development, and current state of the Colorado, Guadalupe, and Trinity River deltas sets the stage for examining recent conditions and trends in other bay-head delta and fluvial areas along the Texas coast. Two fluvial-deltaic areas, San Jacinto River and Neches River, stand out because of relatively recent extensive losses in wetland acreage (White and others, 1985; 1987).

San Jacinto River

A comparison of the distribution of wetlands as interpreted from photographs taken in 1956 and 1979 of the lower San Jacinto River valley at the head of Galveston Bay indicates a displacement of more than 560 ha (1,380 acres) of fluvial woodlands, swamps, and fresh- to brackish-water marshes by open water (Fig. 68). The principal reason for this loss of wetlands is that the lower reach of the San Jacinto River is in an area of subsidence (Fig. 9) caused primarily by ground-water withdrawal (Gabrysch, 1984). Between 1943 and 1978, 1.2 to 2.1 m (4 to 7 ft) of subsidence occurred in this part of the river valley (Gabrysch, 1984). The San Jacinto River lies within an entrenched valley similar to that of the Trinity River (Fig. 3), and as subsidence occurs, submergence and resulting changes in wetland environments progress inland along the axis of the valley (Fig. 69).

The change in wetlands along the San Jacinto River is pronounced because of the proximity of the valley to the center of maximum subsidence. Rates of subsidence, between 1943 and 1978, are as high as 60 mm/yr, which apparently greatly exceeds the rate of wetland aggradation in this area. The volume of sediments reaching the mouth of the San Jacinto River has diminished as a result of reservoir development in the drainage basin (Fig. 70). Lake Houston, which is located only a few miles (< 10 mi) upstream from the mouth of the San Jacinto River, has an estimated trap efficiency of about 87 percent (USDA, 1959), which suggests that only small quantities of sediment are





Wetland Map Unit	1956		1979		Net Change	
	Acres	Hectares	Acres	Hectares	Acres	Hectares
Water	875	355	2,295	930	+1,420	+575
Fluvial Woodlands & Swamps	3,480	1,410	2,090	845	−1,390	−565
Fresh to Brackish Marsh	650	265	300	120	−350	−145

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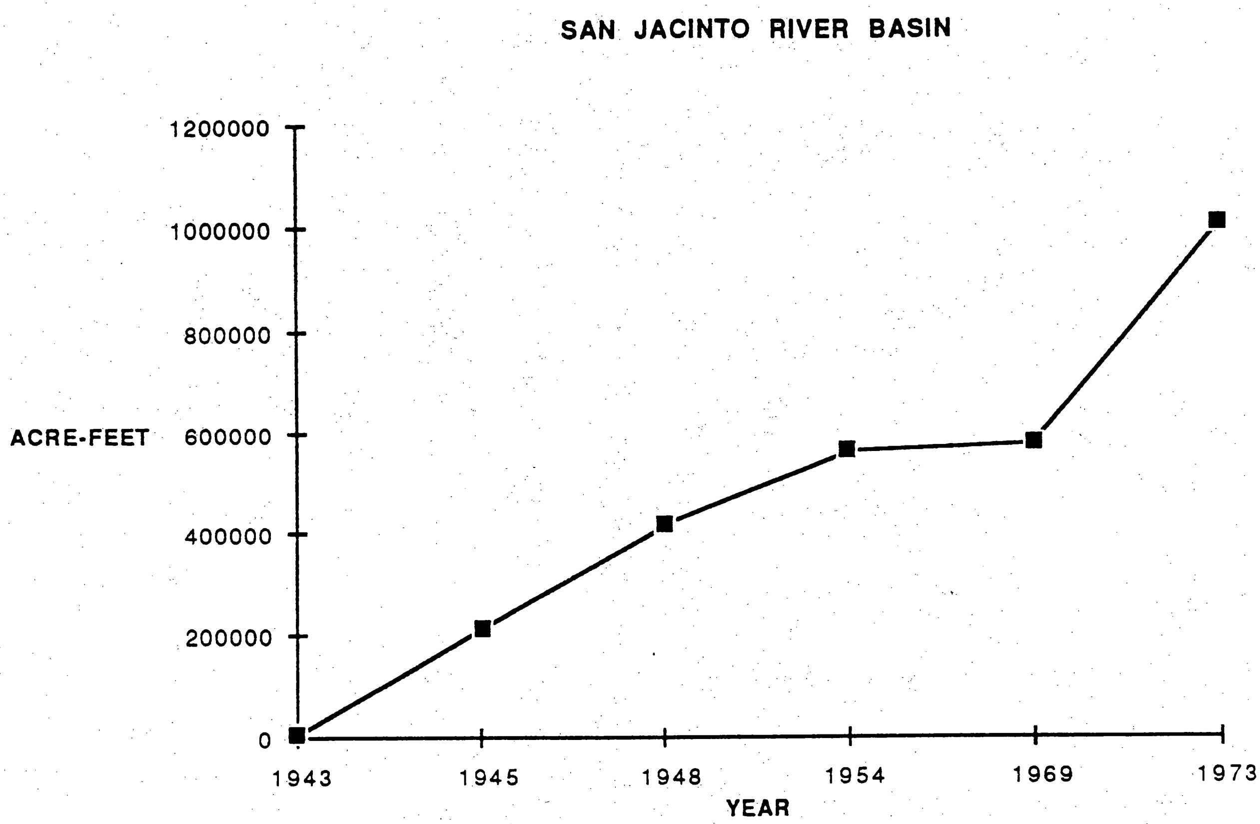
Figure 68. Changes in the distribution of wetlands between 1956 and 1979 of a subsiding segment of the San Jacinto River near Houston, Texas. (After White and others, 1985.)

Figure 69. Aerial photograph of the lower reaches of the San Jacinto River. Compare with figure 68.



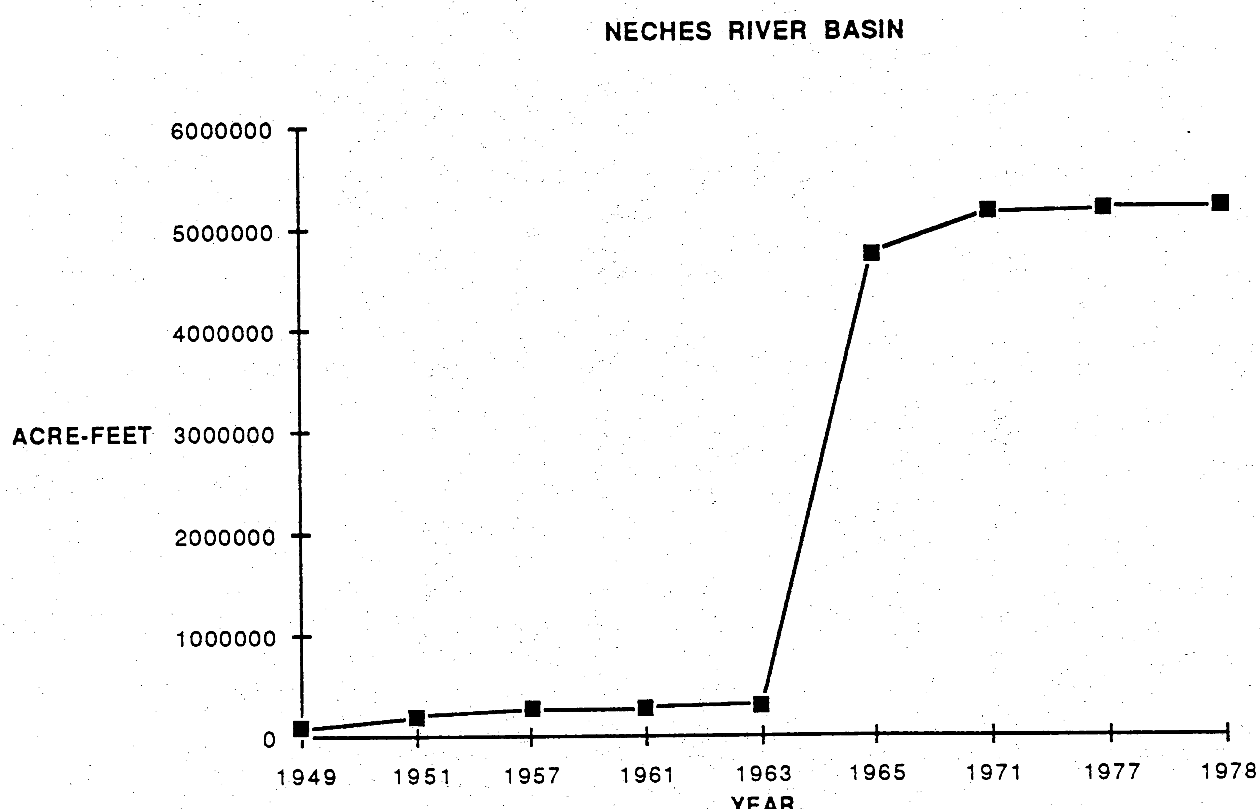
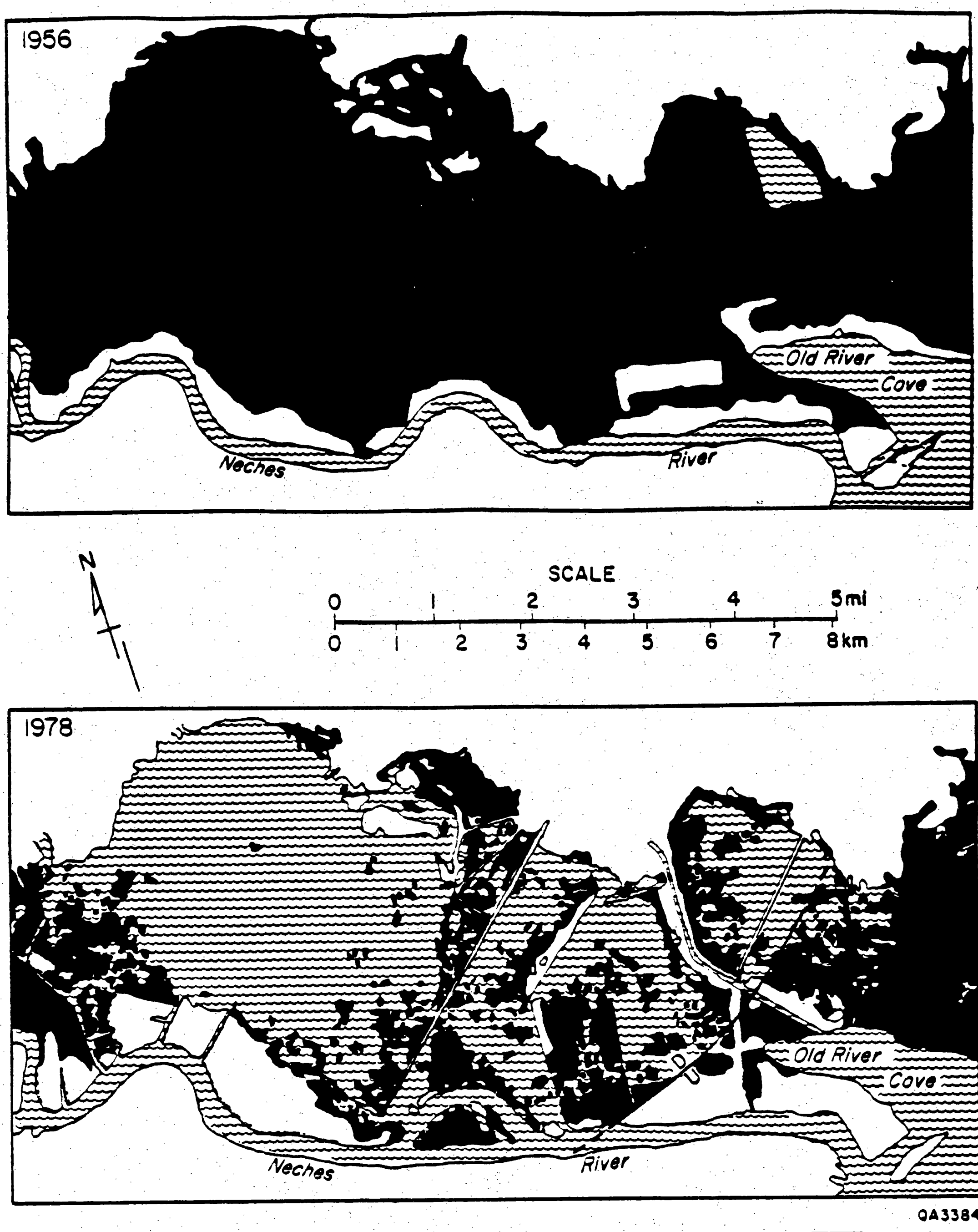


Figure 70. Cumulative reservoir storage capacity for selected rivers along the Texas coast. Data from Texas Water Development Board.

delivered to the area where wetlands are being submerged. Nevertheless, even without Lake Houston and other reservoirs in the San Jacinto River basin, land-surface subsidence is so pronounced that it is unlikely that aggradation rates could keep pace with subsidence rates except farther upstream where the rates decline. It appears that submergence will continue up the axis of the valley in the future, but rates of change should diminish somewhat as bay waters move into areas with slightly higher elevations and lower rates of subsidence.

Neches River

The most extensive loss of wetlands in a modern estuarine fluvial-deltaic system on the Texas coast has occurred along the lower Neches River (White and others, 1987). The Neches River discharges at the head of Sabine Lake; the lower part of the river valley is the site of relatively extensive marshes that have developed on fluvial-deltaic deposits (Fisher and others, 1973). Losses in wetlands in this area have been reported by Wiersema and others (1973), Gosselink and others (1979), and White and others (1987). Between the mid-1950's and 1978, about 3,800 ha (9,410 acres) of marshes were displaced primarily by open water along an approximately 16-km (10-mi) stretch of the lower Neches River valley (Fig. 71). The displacement of marshes by open water and shallow subaqueous flats in this area is apparently related to several factors (Wiersema and others, 1973; Gosselink and others, 1979; White and others, 1987), including: (1) relative sea-level rise resulting primarily from subsidence possibly due to both natural and human-induced causes (Swanson and Thurlow, 1973; Ratzlaff, 1980), (2) a decline in sediments supplied to this alluvial area as a result of (a) reservoir development in the Neches River basin (TDWR, 1981c; Fig. 70), (b) artificial levees (dredged spoil, Fisher and others, 1973) along the dredged portion of the river, and (c) changes in hydrology due to artificial channels (Wiersema and others, 1973), (3) active faulting, in which the downthrown side of one identified fault is subsiding at a more rapid rate than the upthrown side (White and others, 1987; Figs. 13 and 71), and (4) dredged canals, which can cause direct and indirect losses in marshes (Wiersema and others, 1973; Scaife and others, 1983).



Map Unit	1956		1978		Net Change*	
	Acres	Hectares	Acres	Hectares	Acres	Hectares
Water	2,560	1,037	11,070	4,483	+8,510	+3,446
Marsh	15,740	6,375	6,330	2,564	−9,410	−3,811

* Loss of additional 900 acres (365 hectares) of marsh primarily due to spoil disposal.

Figure 71. Changes in the distribution of wetlands between 1956 and 1978 in the lower Neches River valley near the head of Sabine Lake. The fault crossing this area has apparently contributed to the changes. (From White and others, 1987.)

In a brackish-water marsh about 50 km (30 mi) east of Sabine Lake on the Chenier Plain (gulfward of Calcasieu Lake, Louisiana), DeLaune and others (1983) reported that changes from marsh area to open water are occurring at ever-increasing rates, apparently because marsh aggradation rates (averaging 0.8 cm/yr [0.3 in/yr]) are not keeping pace with submergence rates (averaging 1.2 cm/yr [0.5 in/yr]). The change in marsh area to open water has been increasing by a factor of approximately 2 every 6 yr since 1963. DeLaune and others (1983) predict that this marsh area will complete its transformation to open water in less than 40 yr if the trends that have characterized the past 25 yr continue. Among the human activities that may be contributing to the transformation to open water are (1) ship channel construction (promoting salt intrusion and possibly sediment diversion) and (2) oil, gas, and groundwater withdrawals (accelerating subsidence). However, DeLaune and others (1983) concluded that it is difficult to document the human component precisely because of its pervasiveness in this area and because some of the observed trends contradict expected effects.

Similar to those of the marsh system near Lake Calcasieu, the factors contributing to marsh loss listed in the preceding discussion of the Neches River valley are complex and impossible to quantify adequately with existing data. But the conversion of marsh to open water indicates that marsh aggradation rates are not keeping pace with subsidence rates and relative sea-level rise. Subsidence rates in this area of the Neches River valley are not known, but the rate at Sabine Pass reported by Swanson and Thurlow (1973) for 1960 to 1969 is 1.25 cm/yr (0.5 in/yr), a rate equal to that reported by DeLaune and others (1983). This rate is among the highest reported along the Texas coast (Fig. 7). In the Neches River valley, the rate of subsidence may be similar or possibly even higher due to withdrawal of underground fluids (Ratzlaff, 1980) and faulting (White and others, 1987). Over a 22-yr period (1956–1978) the marsh has been replaced by open water at an average rate of about 160 ha/yr (400 acres/yr) (Fig. 71). It is possible that the rate of change has increased in more recent years as reported by DeLaune and others (1983) for the marsh near Lake Calcasieu. There is little evidence to suggest that there will be a reversal in trends, so it is likely that transformation of marsh to open water in the Neches River valley will continue in the future.

Lavaca River

The Lavaca River fluvial-deltaic area is small compared to the Colorado, Guadalupe, and Trinity River deltaic areas. The delta extends only a short distance into Lavaca Bay; fluvial-deltaic deposits capped by marshes are principally restricted to the alluvial valley stretching inland from the head of the bay (Fig. 1). An analysis of historical changes in deltaic shorelines and associated marsh areas in the Lavaca River delta indicates that the configuration of the delta margin did not change significantly between the mid-1850's and mid-1950's (McGowen and Brewton, 1975). Only about 39 ha (96 acres) of new wetlands formed as a result of deltation at the mouth of the Lavaca River; the active delta prograded about 4.3 km (2.7 mi). Inactive parts of the delta to the west eroded at rates of 0.1 to 1.5 m/yr (1 to 5 ft/yr) since 1934 (McGowen and Brewton, 1975). McGowen and Brewton reported that the Lavaca River delta was stable in terms of marsh surface-water level relationships, indicating a balance between sedimentation and subsidence rates.

Upstream from the delta along the Lavaca River valley, relative sea-level rise due principally to subsidence has apparently contributed to local wetlands submergence and the expansion of water bodies into areas formerly occupied by marshes and fluvial grassland (White and others, 1985; Fig. 72). Between 1918 and 1973, subsidence of approximately 15 cm (0.5 ft) has occurred over a relatively broad area and locally has exceeded 30 cm (1 ft) (Ratzlaff, 1980). Aggradation rates do not appear to be keeping pace with subsidence rates in this area (Fig. 72) of the Lavaca River valley. Not shown in figure 72 are extensive alterations of the natural hydrologic regime by artificial levees and canals. Reductions in fluvial sediments supplied to areas like this one will probably accelerate the transformation of wetlands to open water. Sediment yields to the mouths of the Lavaca River and Navidad River drainage basins are reportedly about 273 and 784 acre-ft/yr, respectively (Greiner, 1982). If Lake Texana, completed in 1980, traps a large percentage of the sediment delivered through the Navidad River system, it seems likely that fluvial-deltaic areas in the Lavaca River valley will fall farther behind aggradation rates, thus promoting more extensive submergence of valley wetlands in the future.

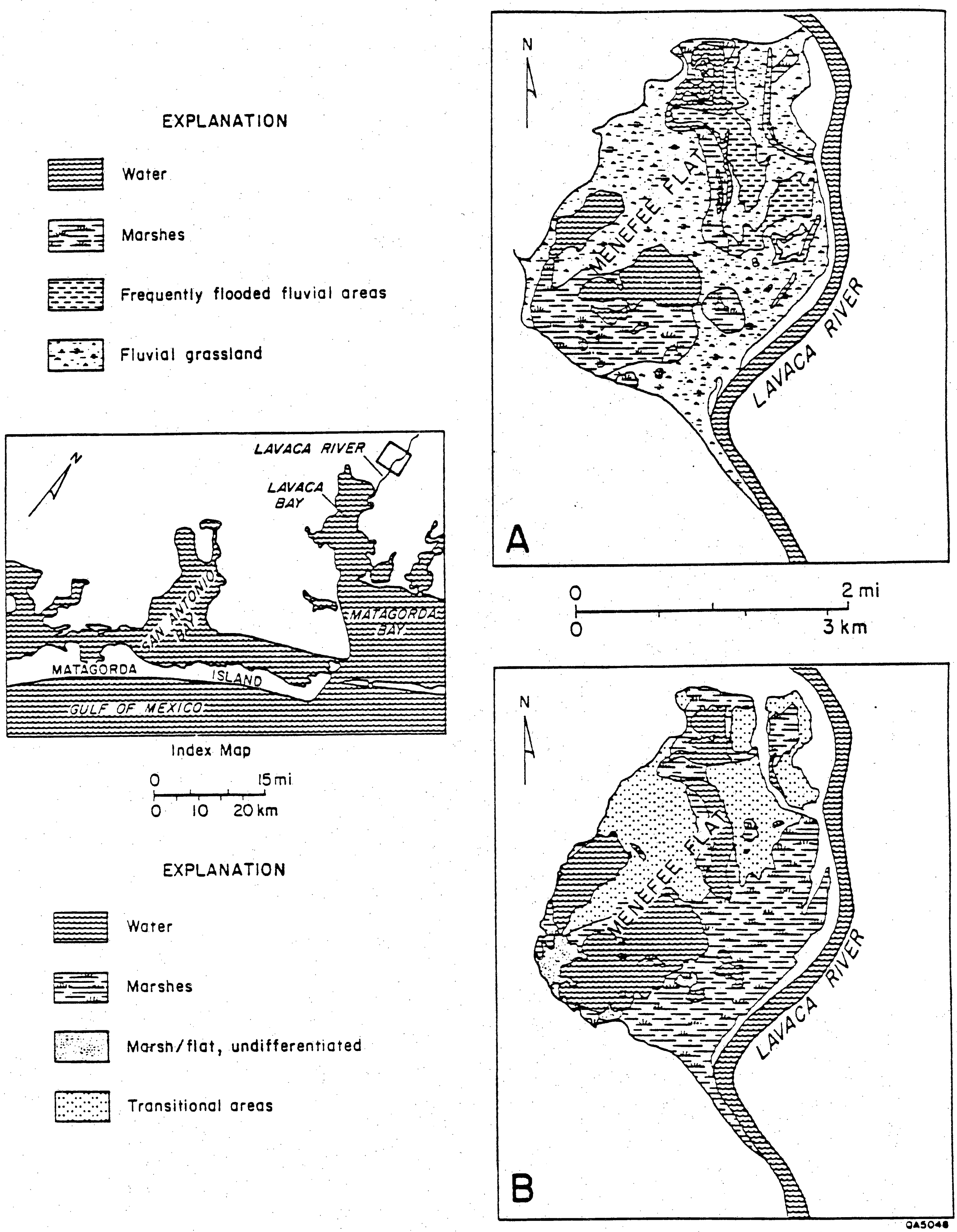


Figure 72. Changes in the distribution of wetlands between 1958 (A) and 1979 (B) at Menefee Flat in the Lavaca River valley. (From White and others, 1989.)

Nueces River

As pointed out by Morton and Paine (1984), Nueces Bay, into which the Nueces River discharges, is shallower, has a shorter wave fetch, and receives more fluvial sediment than nearby Corpus Christi Bay; these factors can promote shoreline accretion and reduce erosion. Monitoring of historical changes in shorelines from map surveys and photographs made in the late 1800's to 1982 indicates net accretion (advancement of the shoreline toward Nueces Bay) of marshes on the Nueces River fluvial-deltaic complex. Although net progradation was recorded for this period (1867 to 1982), aerial photographs taken in 1959, 1971, and 1974 indicate that marsh progradation ended sometime between 1930 and 1959 (Morton and Paine, 1984).

Comparison of interior marshes in the Nueces River fluvial-deltaic system on photographs taken in the mid 1950's and late 1970's suggests an increase in shallow subaqueous flats and a corresponding decrease in marsh area (White and others, 1983). Subsidence reported in the Nueces fluvial-deltaic area is interpreted to be on the order of 6 to 30 cm (0.2 to 1.0 ft) for the period 1942–1951 (Brown and others, 1976). These amounts of subsidence translate into annual rates of about 0.7 to 3 cm (0.28 to 1.2 in). It is possible that marsh sedimentation rates are falling behind subsidence rates. Fluvial sediments along the Nueces River have been reduced by Lake Corpus Christi (Fig. 32). In fact, more recent annual averages of suspended sediments measured on the Nueces River are only about 4 percent of those reported by Shepard (1953) (see Corpus Christi–Nueces Bay System, p. 179–181). Sediment yield at the delta for the drainage area below Lake Corpus Christi is approximately 190 acre-ft/yr (Greiner, 1982). If all the suspended load measured below Lake Corpus Christi at Mathis were to reach the delta, this would add another 19 acre-ft/yr. These data indicate that an average of about 210 acre-ft of fluvial sediment reaches the delta each year. This volume is about 40 percent of pre-1952 sediment loads reported by Shepard (1953) for the Nueces at Three Rivers.

BAY–ESTUARY–LAGOON SEDIMENTATION

Sediment Composition

Benthic sediment textures in the bay–estuary–lagoon system along the Texas coast have been mapped, most extensively, by McGowen and Morton, (1979) and White and others (1983, 1985, 1986, 1987, 1988, 1989, and in press). Earlier maps depicting sediment distribution in selected Texas bays include those by Shepard and Moore (1955), Fagg (1957), Shenton (1957), Elliot (1958), Rusnak (1960), Kane (1959, 1966), Alaniz and Goodwin (1974), Hall (1976), McGowen and Morton (1979), and Shideler and others (1981).

Bay sediment types are principally sand, mud (silt and clay), and gravel (composed primarily of shell material). The distribution of these different textural components follows a somewhat systematic trend for the larger bay systems, where muds characterize the deeper bay centers and sand the shallower bay margins (Figs. 73 to 79). The distribution of sediments is in part a reflection of the wave and current energy, which is related to water depth. Sands along the margins of larger bays reflect not only nearby sources of sand but also the relatively high energy of these shallow environments where breaking waves and littoral currents are common. Sand eroded from the bay margins remains in this environment because current energy decreases in deeper water. In a study of bay-margin sands in the Galveston–Trinity Bay system, Love and others (1985) found a correlation between prevailing and dominant wind directions and bay-margin sands, indicating a relationship between the wind-generated wave and current energy and water depth of sand deposition. Sands are concentrated (1) at the mouths of rivers where channel mouth sand bars are deposited and reworked in nearshore areas, (2) at the mouths of tidal inlets where current energy is sufficient to transport sand, and (3) along the margins of sand-rich barrier islands and peninsulas (Figs. 75 and 76). Shelly substrates are concentrated near oyster reefs and locally along bay margins where shell is deposited by waves and currents in beaches and storm berms.

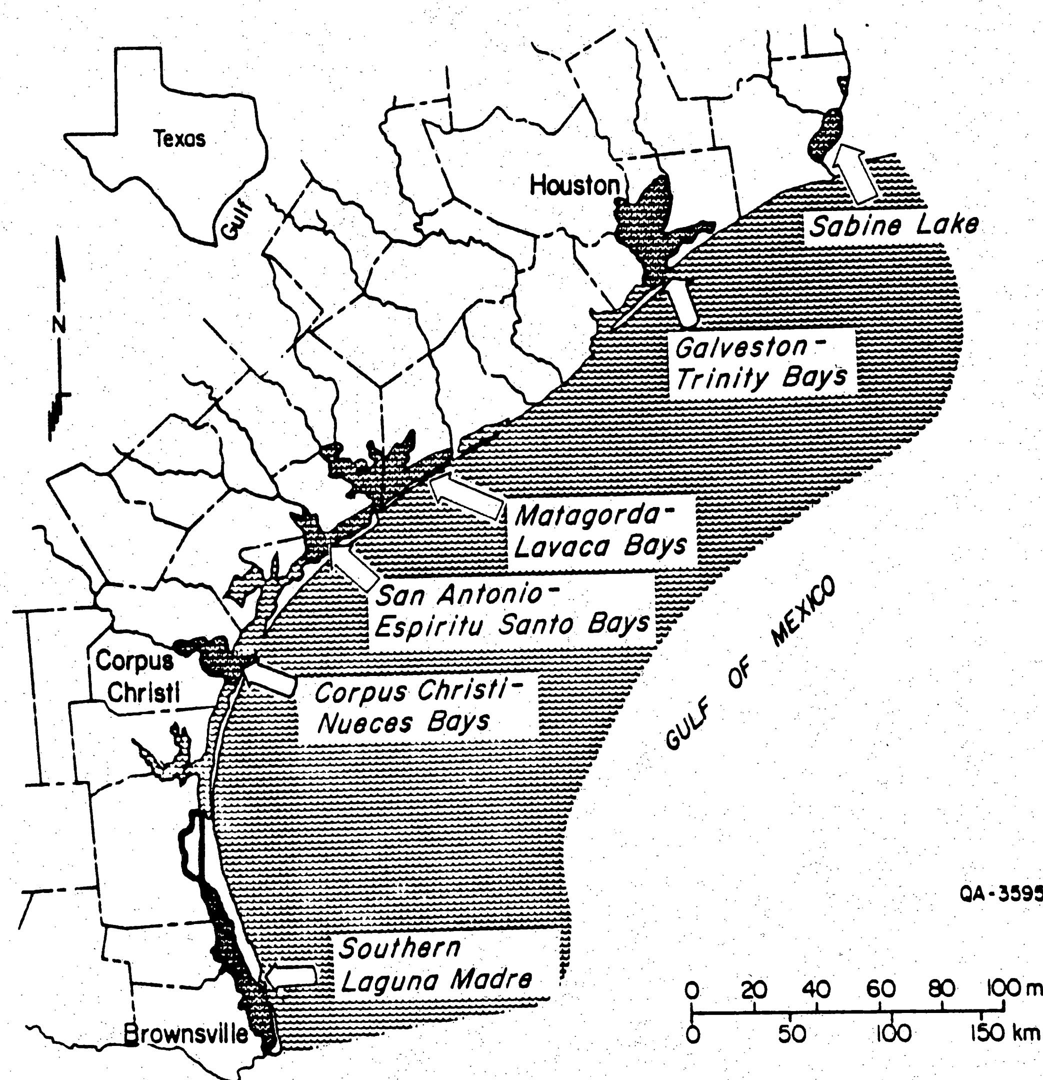


Figure 73. Index for sediment distribution maps shown in figures 74–79.

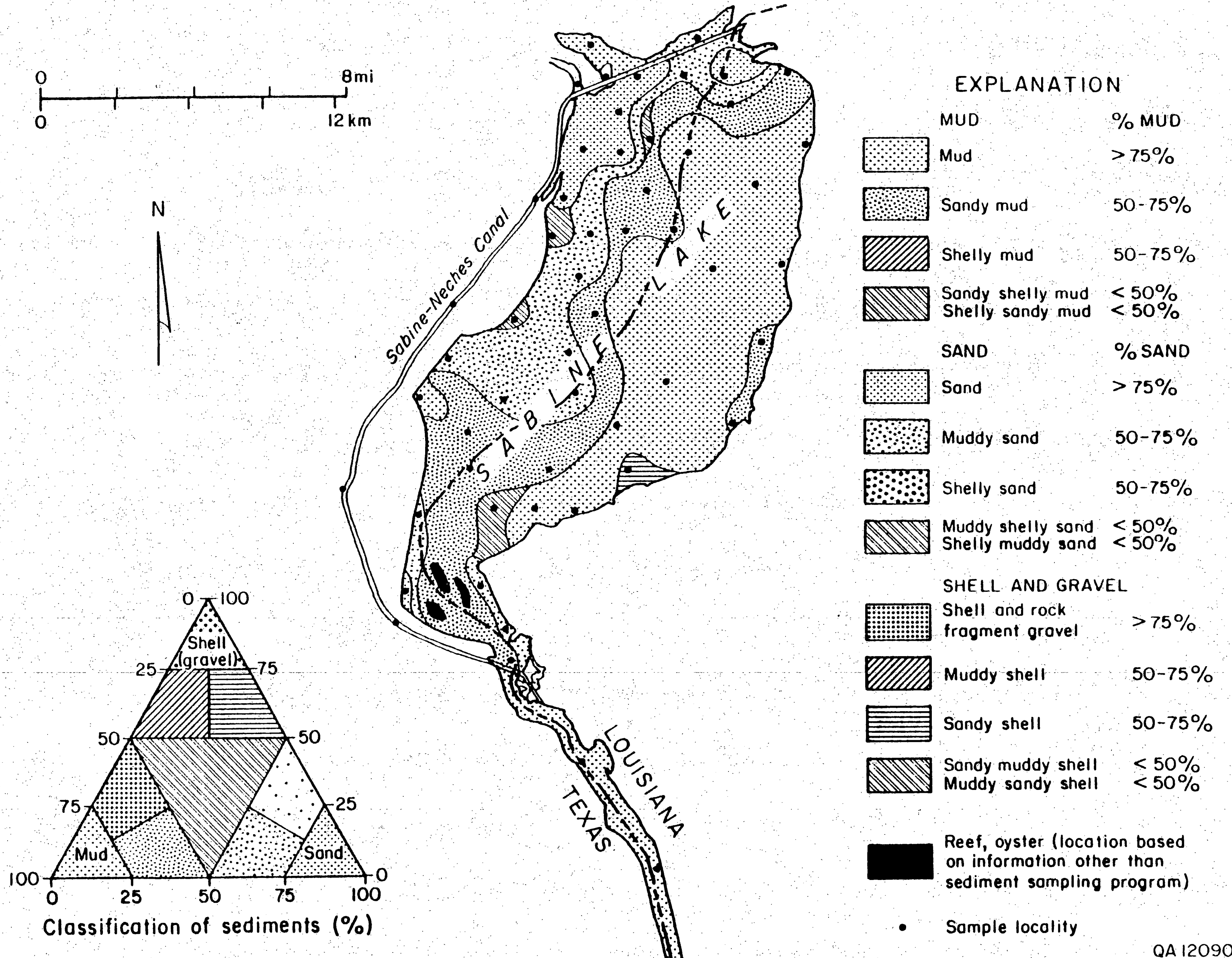


Figure 74. Distribution of sediments in Sabine Lake. (From White and others, 1987.)

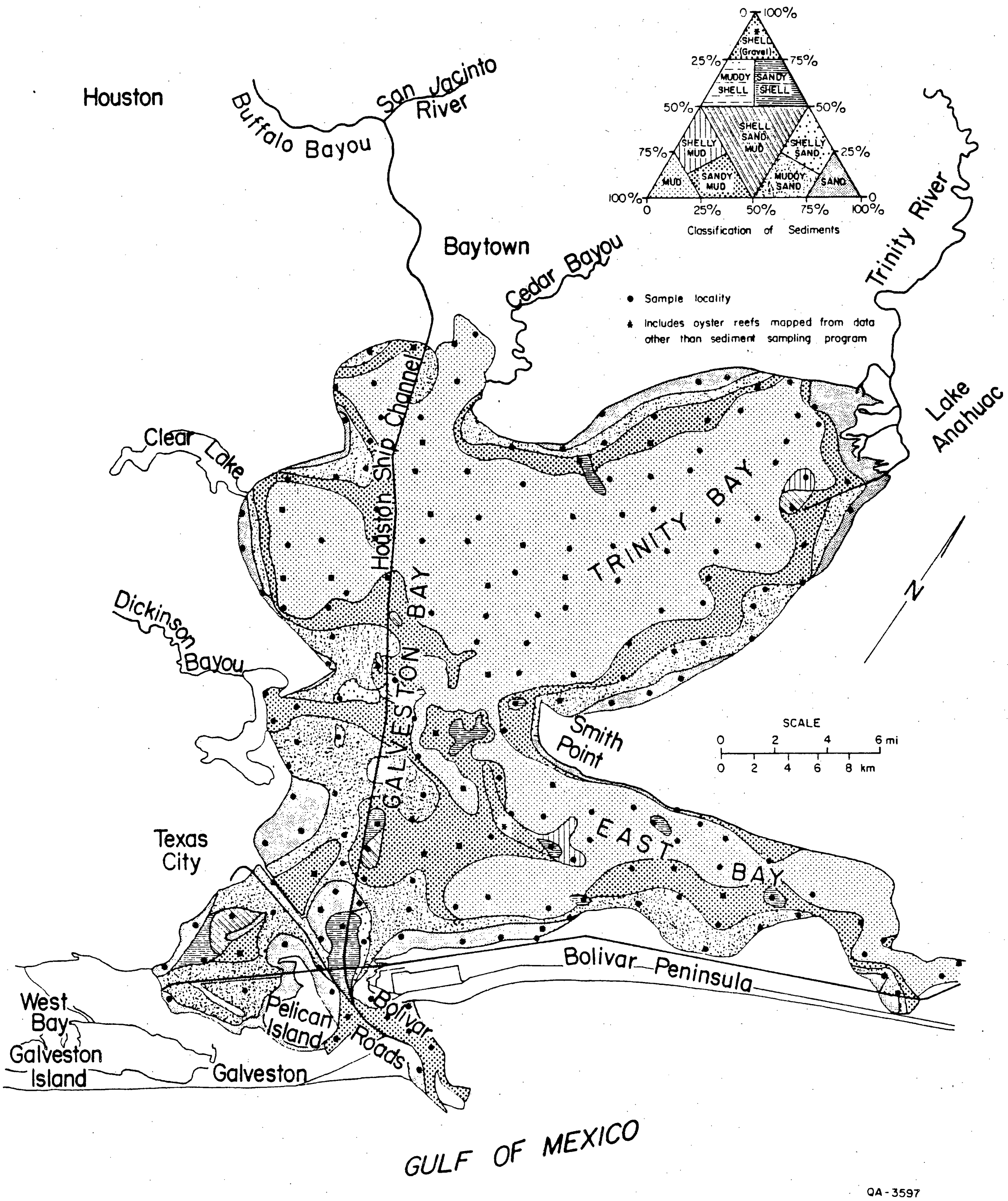


Figure 75. Distribution of sediments in Galveston Bay. (From White, 1985; modified from White and others, 1985.)

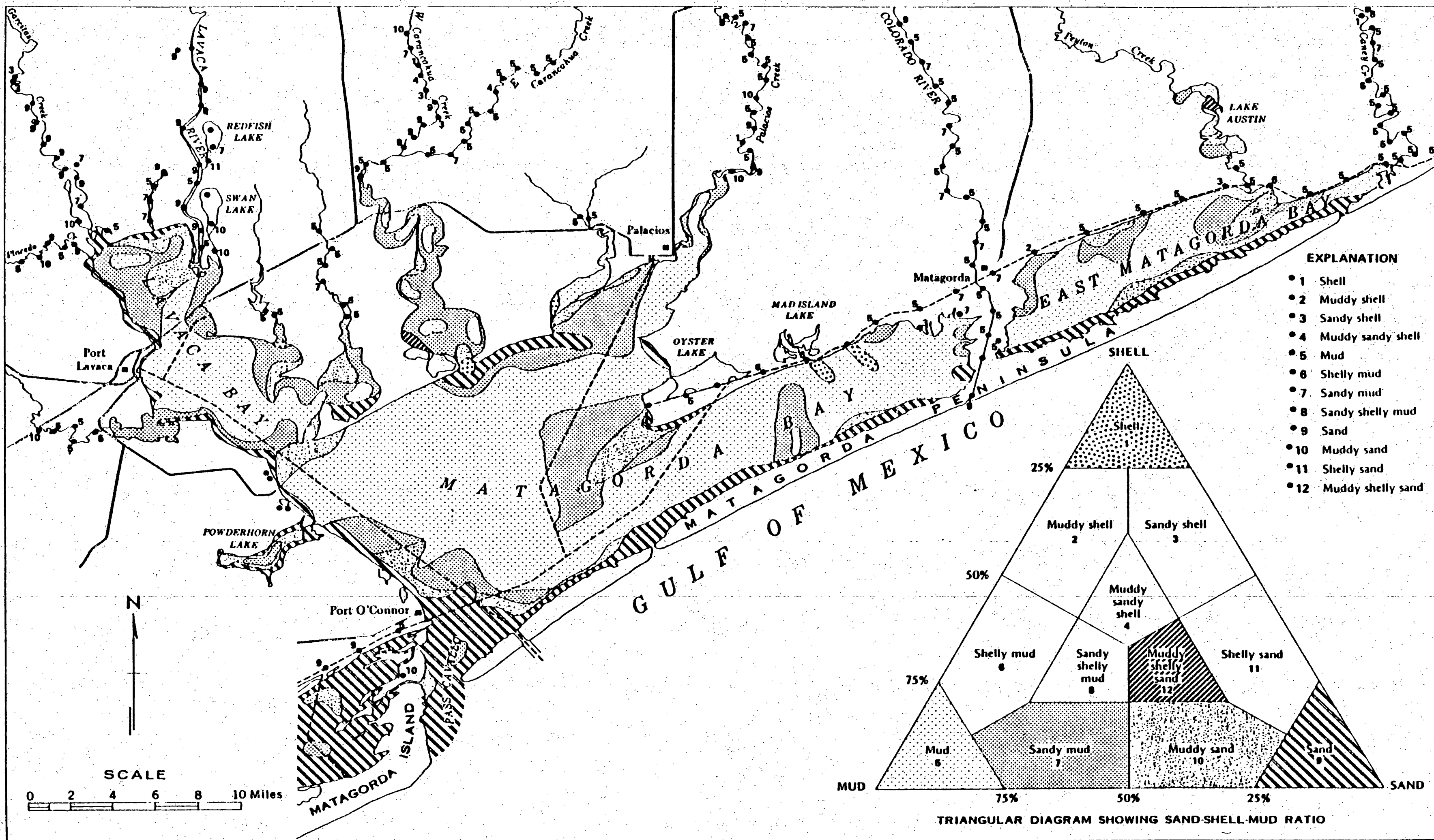


Figure 76. Distribution of sediments in Matagorda Bay. (From McGowen and others, 1979.)

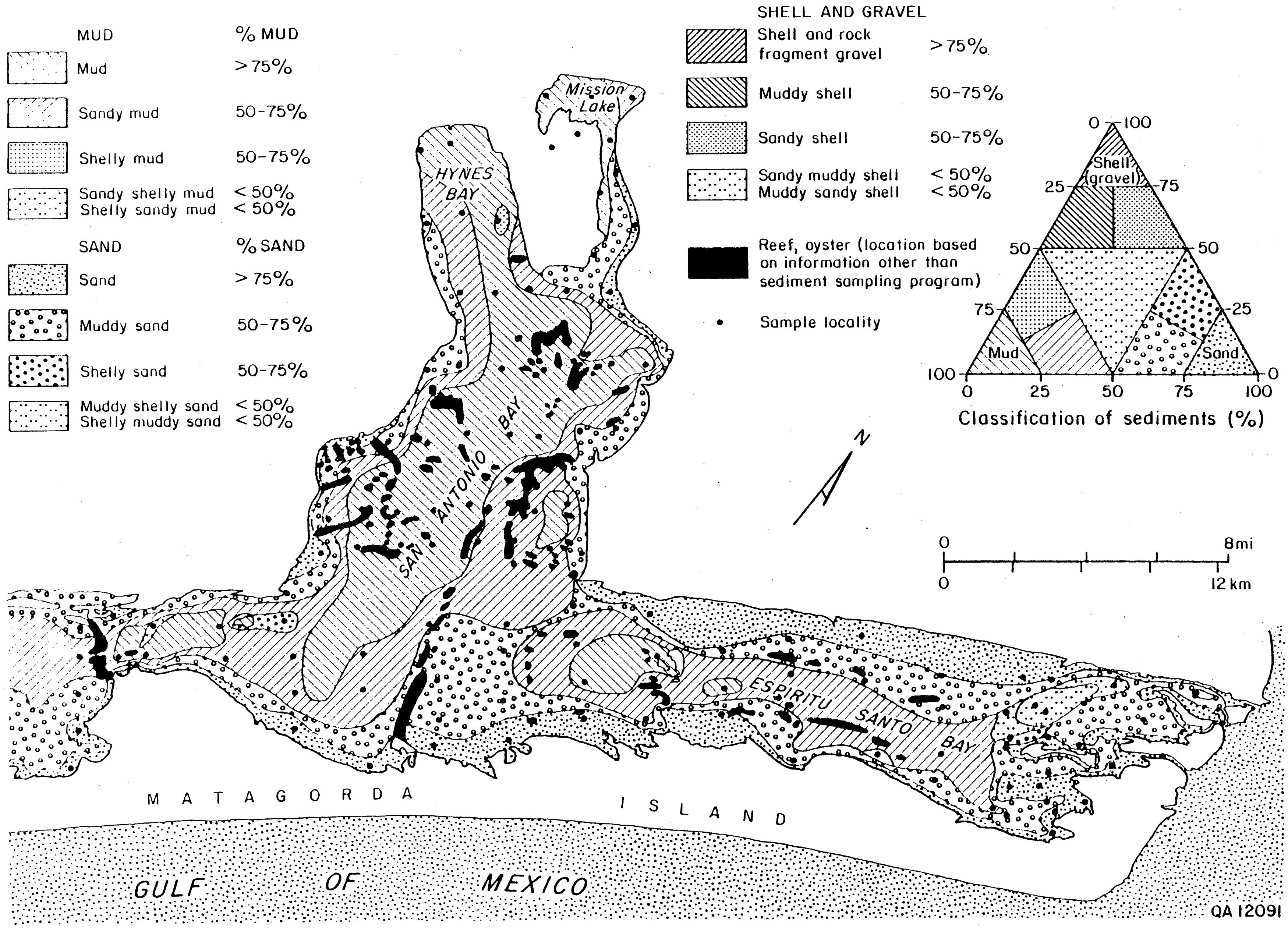


Figure 77. Distribution of sediments in San Antonio and Espiritu Santo bays. (From White and others, 1989.)

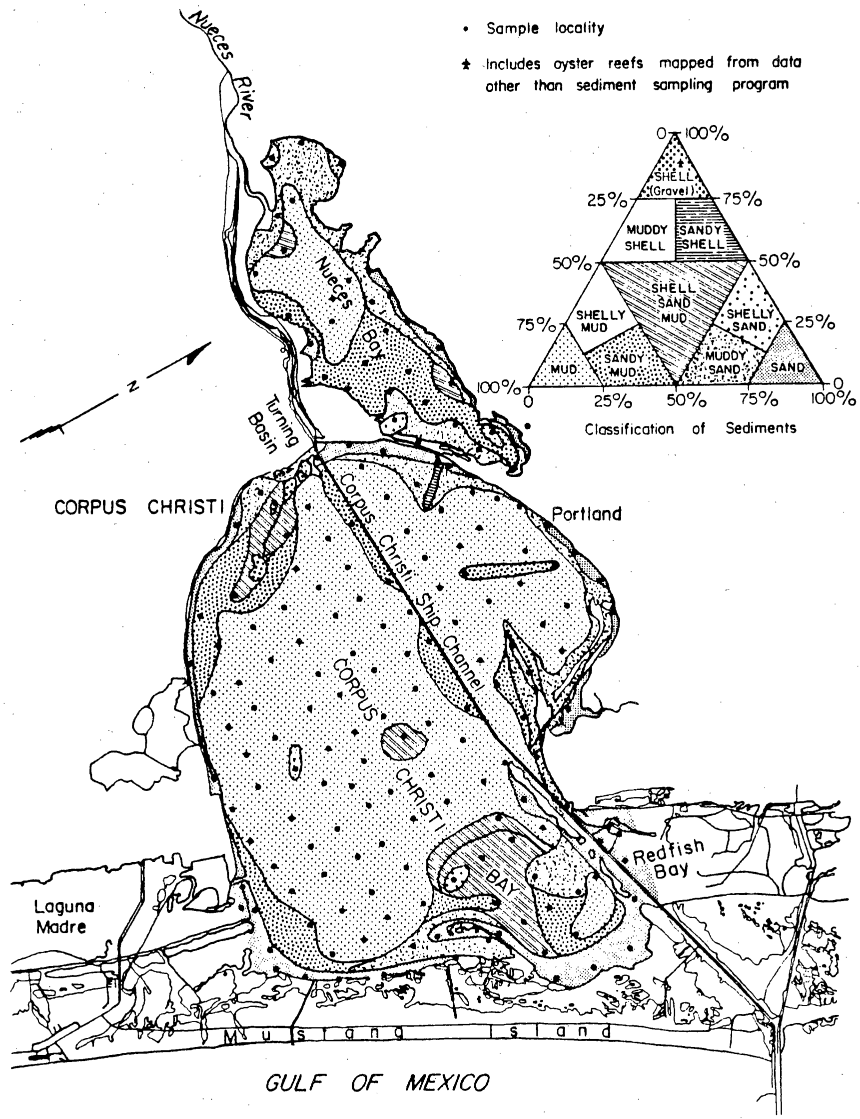


Figure 78. Distribution of sediments in Corpus Christi and Nueces Bays. (From White, 1985; modified from White and others, 1983.)

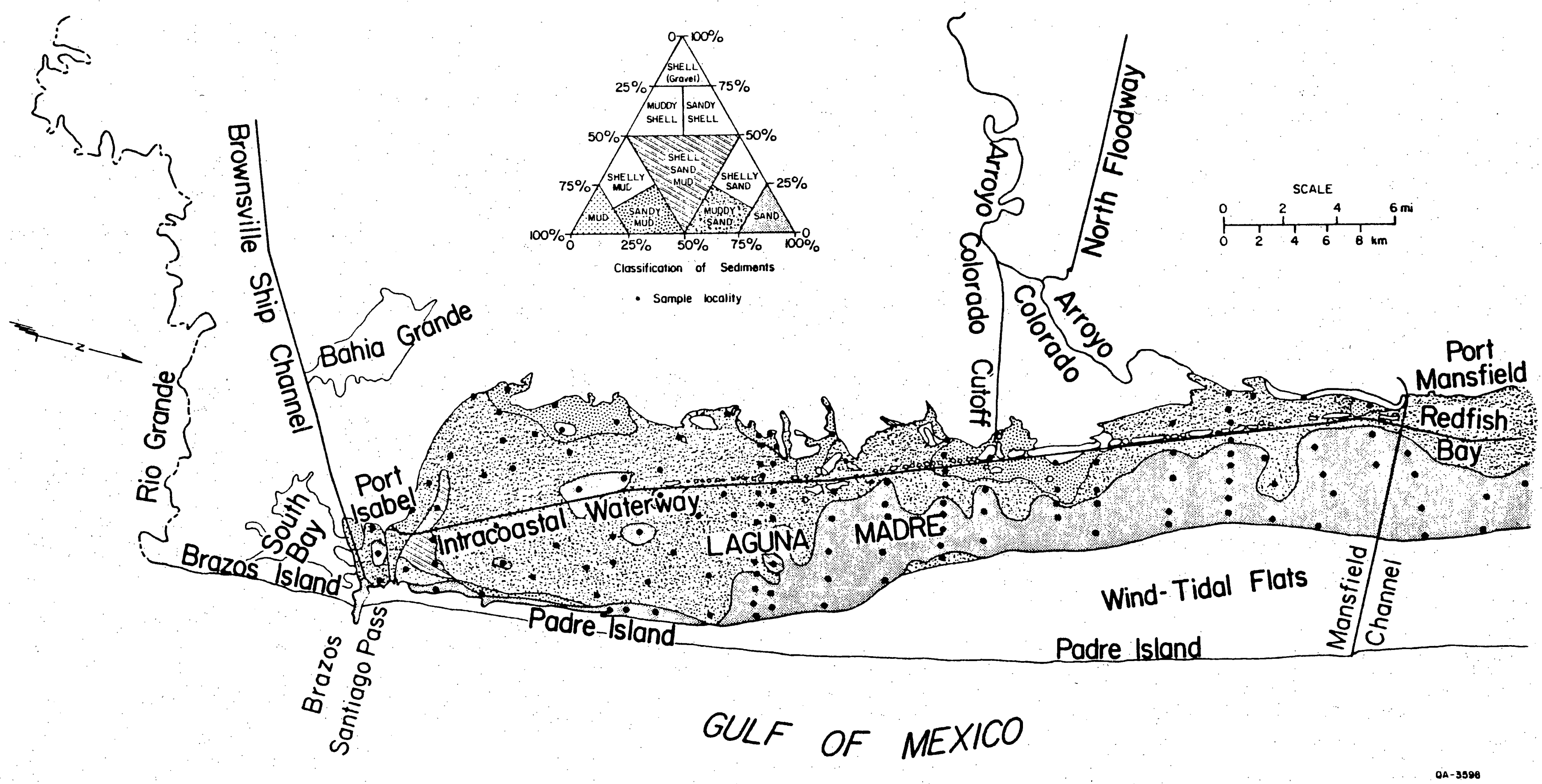


Figure 79. Distribution of sediments in southern Laguna Madre. (From White, 1985; modified from White and others, 1986.)

Inorganic bottom sediments in the bay–estuary–lagoon system are mostly composed of quartz, feldspar, and clay minerals. Oolites and coated grains are among the constituents of sediments in Baffin Bay along the south Texas coast (Behrens, 1963; Dalrymple, 1964; Frishman, 1969). In the suite of clay minerals, montmorillonite is more abundant than illite and kaolinite along the western coast of the Gulf of Mexico (Elliot, 1958; Folger, 1972; Morton, 1972; Byrne, 1975; Sorenson, 1975). Along the eastern Gulf and southeast Atlantic coast, kaolinite is predominant, and along the northeast Atlantic coast, illite and chlorite are most abundant.

Sources of Sediment

Kranck (1984), in citing Judson (1968) and Milliman and Meade (1983), noted that about 90 percent of the 15×10^{15} g/yr of terrestrial sediment delivered by the world's rivers is deposited near the continents, primarily in estuaries. Most of the estuaries along the Atlantic coast appear to be filling without much transfer of sediments to the shelf (Meade, 1982). Although landward sources of sediments are apparently the principal ones that supply sediments to estuaries (Meade, 1982), transport of sediments from the marine system into the estuaries, or from the mouth of the estuary landward, appears to be an important process (Stewart, 1958; Ritter, 1967; Windom and others, 1971; Emery and Uchupi, 1972; Pevear, 1972; Schubel and Carter, 1976; Renwick and Ashley, 1984; Summerhayes and others, 1985). Studies of sedimentation in Chesapeake Bay indicate that the importance of various sources of sediment changes from the head of the bay to its mouth (Schubel, 1968, 1972; Biggs, 1970; Schubel and Carter, 1976, 1984; Biggs and Howell, 1984; Officer and others, 1984). Approximately 70 percent or more of sediment from the Susquehanna River, which is the principal source of sediment to the main body of Chesapeake Bay, is trapped in the upper 30 km (19 mi) of the bay (Biggs, 1970; Schubel, 1972). Schubel (1968) estimated that in northern Chesapeake Bay the annual contribution of sediment from river runoff was about 5 times the amount from shoreline erosion. In the middle bay, bay shoreline erosion is the primary source, and near the mouth of the estuary, sediments are derived principally from a marine source. Yarbro and others

(1983), in a study of the sediment budget of a major estuary (Choptank River Estuary) on the eastern shore of Chesapeake Bay, found shoreline erosion to be the principal source of sediment, contributing 7 times the volume supplied by upland runoff. Factors influencing the predominance of erosional sources over river sources were low relief, rural character of the drainage basin, and poorly consolidated shoreline material, which contributed to high erosion rates (Yarbro and others, 1983).

Rusnak (1967) classified estuaries on the basis of the principal sediment source areas that were filling the estuary. Estuaries primarily filled from a land source (fluvial) were positive-filled estuaries, from a marine source, negative-filled, and from an internal source (redistribution of sediments), neutral-filled (or non-filling). Those Texas bays with relatively large fluvial sources, such as Sabine, Trinity, Lavaca, San Antonio, and Nueces, would be classified as positive-filled, whereas Laguna Madre, with a principal source from the Gulf, would be negative-filled. Some estuaries are being filled from both a marine and landward source (Fig. 80). Although most of the Texas estuaries along the central and northern Gulf coast are dominated by sediment deposition from fluvial sources, and secondarily from shoreline erosion, deposition also occurs at the mouths of tidal inlets (flood tidal deltas) indicating a marine source. The sources and major depositional features, as well as some of the hydrodynamic interactions in the estuarine basins are depicted in Figure 81.

Principal sources of sediments in the Texas bay–estuary–lagoon system include (1) suspended load and bed load of rivers and streams, (2) erosional products from bay margins, (3) Gulf or marine sediments transported through tidal passes and storm washover channels, (4) sediments transported across barrier islands and peninsulas by eolian processes, (5) nonterrestrial biogenic materials, primarily oyster shells (serpulid reefs in Baffin Bay), and (6) dredged material placed on the bay floor or bay bottom. Along the central and upper Texas coast, rivers that discharge into the bay–estuary–lagoon system are the principal source of sediments. For example, in the Lavaca Bay system, Wilkinson and Byrne (1977) concluded that rivers discharging into the bay account for roughly 73 percent of the sediment deposited there. Sediment from shoreline erosion was a significant, although secondary, source, supplying about 24 percent of the sediment supply. Estimates of

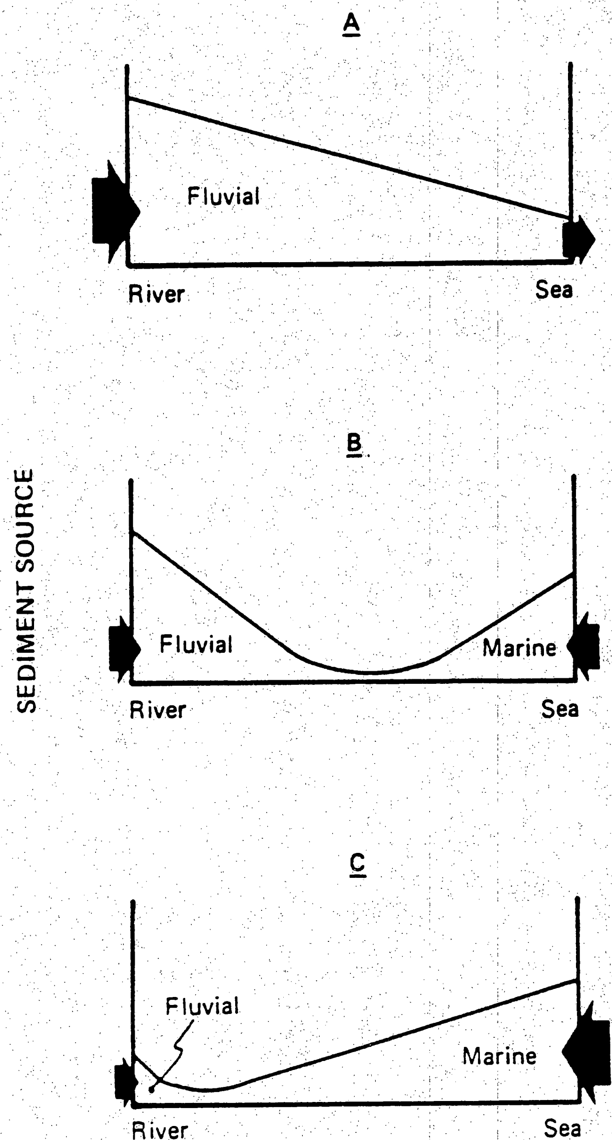
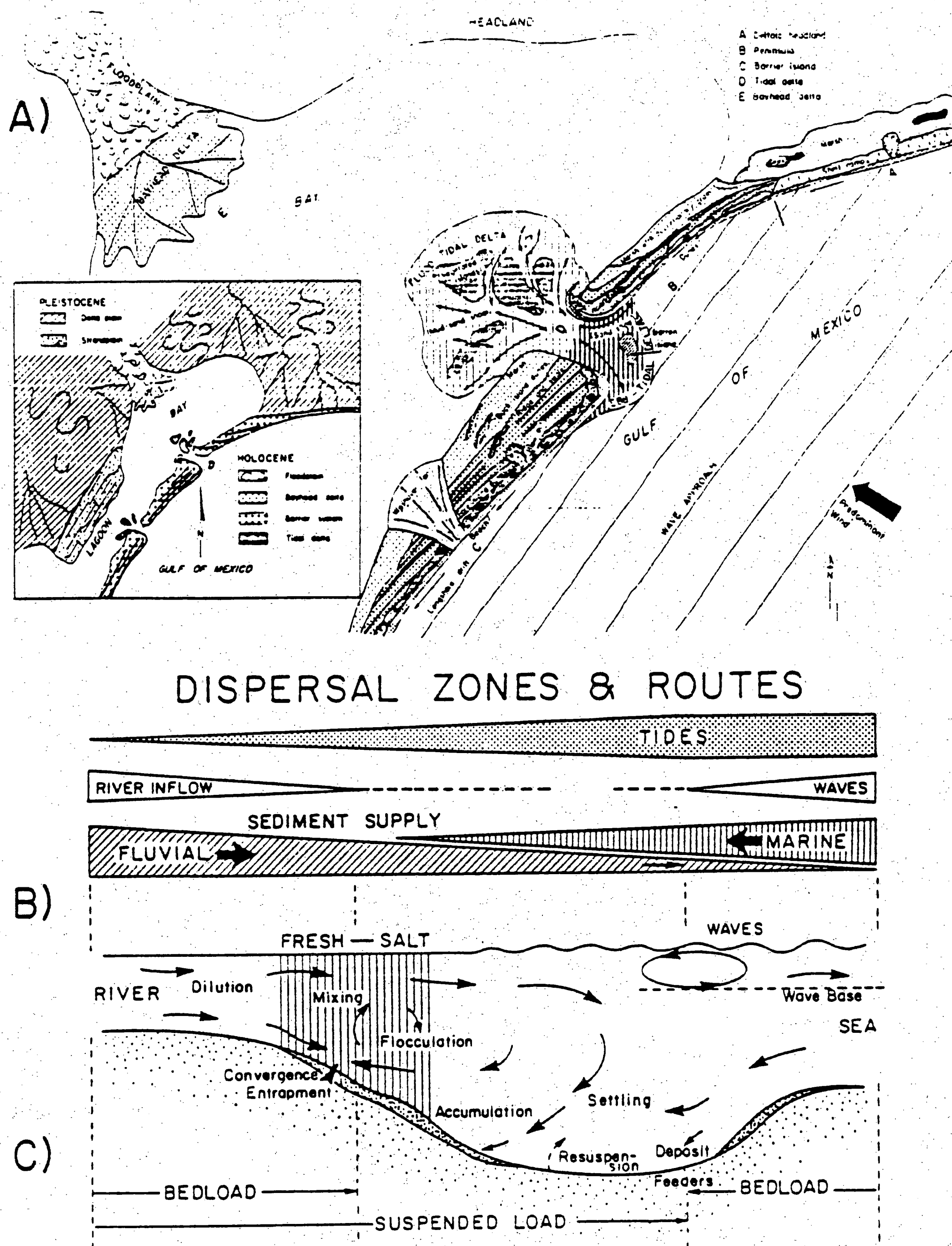


Figure 80. Conceptual model of sediment sources and associated deposition relative to the magnitude of the source. A system like the Mississippi, in which fluvial sediments escape seaward, is depicted in "A"; a system like Chesapeake Bay, in which fluvial and marine sediments are trapped in the estuary, is depicted in "B"; a system like Laguna Madre in Texas, which has relatively small fluvial input, is depicted in "C". (From Biggs and Howell, 1984; reprinted by permission.)



DYNAMIC CHARACTER

Tidal flow attenuated Seaward river flow	Tidal flow increasing seaward River flow influence during flood Estuarine circulation, two-way, two-layered dominant Wave activity near-surface	Tidal currents dominant with ebb-flood differentiation in channels Estuarine circulation diminished Wave activity intense in inlet

FACIES CHARACTER

Accumulation in meander bars dominate; cross-bedded sand and clay lenses Channel: interbedded sand and clay; gravelly Marsh and swamps: organic rich clay and silt	Shoals: rippled sand with clay pebbles and erosional contacts Flats: laminated clay or sand and clay, bioturbated Marsh: massive clay with plant fragments Channel: clay and silt laminated and interbedded with sand lenses and shells, massive silt and clay, bioturbated, seaward	Shoals: ripple cross-bedded medium to fine sand Channel: ripple cross-bedded coarse sand

Figure 81. Depositional features typical of (A) Texas bay-estuary systems (From McGowen and others, 1979), related to (B and C) sediment sources and processes (From Nichols and Biggs, 1985; reprinted by permission).

sediment contributions from various sources in the Galveston–Trinity Bay system by the U.S. Army Corps of Engineers (USACE, 1942; cited in Rehkemper, 1969b), indicate that fluvially derived sediments make up more than 90 percent of the total supply, with shoreline erosion contributing less than 5 percent. Along the southern Texas coast, where major rivers do not discharge into the bays and lagoons, fluvial sediment input is less important. The major sources of sediment deposited in Laguna Madre, for instance, are the Gulf and Padre Island, from which sand is transported by eolian and storm washover processes (Price and Gunter, 1942; Lohse, 1955; Price, 1958; Fisk, 1959; Hunter and Dickinson, 1970).

Turbidity

The characteristically turbid nature of waters in an estuary is a product of: (1) particulate matter from the watershed, off-inlet shores and bottoms, reworking and scouring of estuarine bottoms by tidal currents and waves, loosening of bottom sediments by burrowing animals, and decomposition of pelagic and benthic organisms; (2) the net two-layered opposing estuarine circulation pattern; (3) the mixing of fresh and salt water and consequent flocculation of finer particles; and (4) the presence of relatively quiet sedimentation areas provided by semi-enclosures and widening of the estuarine basin (Carriker, 1967). The release of large quantities of organic-laden silts and clays into the water column is a dominant feature of Texas bays and estuaries whether from beach erosion, river input, storm reagitation, or dredging (Odum and Wilson, 1962).

Turbidity generally varies widely throughout the year and reaches a maximum during floods of the rainy season. At all times of the year, there is a gradual decrease in turbidity with distance from the river mouth (Emery and Stevenson, 1957). As a consequence of estuarine tidal mixing, sedimentation, and reworking processes, turbidity is higher at a given point in the estuary at low than at high tide (Carriker, 1967). Turbidity can also change considerably over short distances and from hour to hour. For example, Mackin and Hopkins (1961) studied oyster mortality in Barataria Bay, Louisiana, and reported turbidity readings ranging from 18 to 198 ppm over a distance of less than 61 m (200 ft). Also, after a combination of high wind and rainfall, Mackin and Hopkins (1961) reported turbidities of up to 900 ppm in Barataria Bay. However, high turbidities were of short duration, normally lasting a matter of hours. Shideler (1980) observed that the response of Corpus Christi Bay's circulation system and associated turbidity patterns to changing wind conditions was rapid (less than a few hours), mainly because of the shallowness of the bay. Shideler (1980) concluded that for Corpus Christi and Nueces Bays wind was the dominant forcing agent influencing turbidity in the bayhead sector, where it both generates waves that resuspend bottom sediment and regulates river inflow. Turbidity in the baymouth sector was mainly influenced by tidal-forcing effects from Aransas Pass tidal inlet (Shideler, 1980).

Turbidity fluctuates diurnally in Texas bays and estuaries, especially in late spring and summer (Conover, 1964). Conover (1964) observed that on clear summer mornings, winds were often light until noon but would increase to 18 knots in the afternoon. Winds above 14 knots could raise fine sediments from the bottom and hold them in suspension. On a summer day, afternoon turbidities would reduce by more than one-third the amount of light penetrating 1 m below the surface compared with measurements taken in the morning. Turbidity was of little consequence in the winter except during stormy weather (Conover, 1964).

Spatial and temporal data on turbidities in Texas bays are generally recorded at only a few stations in a bay system and at variable times during the year. One of the best long-term data sets on turbidity is from measurements taken by the Texas Parks and Wildlife Department (Martinez, 1969 through 1975). Martinez (1969 through 1975) took monthly turbidity measurements in most Texas bays, estuaries, and lagoons, and even though spatial coverage in some bay systems was sometimes minimal, measurements were taken at most stations throughout the year. For example, turbidity data were taken from 1969 through 1975 at two to three stations in upper Trinity Bay and two to three stations in Matagorda Bay near the Colorado River delta (table 11). Mean turbidities were generally higher in upper Trinity Bay (66 to 169 ppm) than in Matagorda Bay near the Colorado River delta (35 to 73 ppm). Minimum turbidities in upper Trinity Bay generally occurred in the summer or fall and maximum turbidities occurred in most months of the year. In Matagorda Bay near the Colorado

Table 11. Minimum, maximum, and mean turbidity measurements (ppm) for upper Trinity Bay and Matagorda Bay near the Colorado River delta. Compiled from Martinez (1969 through 1975.)

	Upper Trinity Bay			Matagorda Bay near the Colorado River delta		
Year	Minimum	Maximum	Mean	Minimum	Maximum	Mean
1969	22 (Sep)	211 (Aug.)	66	5 (Jul)	300 (May)	73
1970	12 (Nov)	258 (May)	91	5 (Dec)	120 (Aug)	36
1971	44 (Feb)	114 (Jul, Nov)	76	0 (Jan)	250 (Mar)	40
1972	16 (Aug)	395 (Oct,Dec)	102	20 (Feb)	100 (Jul)	35
1973	5 (Apr)	395 (Mar)	169	8 (Feb)	90 (Jun)	45
1974	51 (Aug)	380 (Feb)	125	0 (Mar)	200 (Sep)	48
1975	22 (Jul)	390 (Jun)	154	4 (Nov)	75 (Oct)	38

River delta, minimum turbidities occurred generally in winter months and maximum turbidities in the summer or fall.

Turbidity and subsequent siltation are important environmental variables for both pelagic organisms and benthic species and seagrasses, because they can affect organism and plant distribution by: (1) decreasing the depth of the euphotic zone and thus decreasing primary production and productivity in seagrass beds; (2) increasing oxygen demand; (3) limiting suspension feeding; and (4) smothering benthic forms. Turbidity plays a principal role in regulating primary productivity rates (Flint and others, 1982). Odum and Wilson (1962) conclude that in order to produce maximum total photosynthesis in all the waters of Texas, measures for management should include reducing turbidity. Phytoplankton production per unit area is greater in shallow, clear bays, such as Redfish Bay and upper Laguna Madre, than in deep, turbid bays, such as Copano, Aransas, and Corpus Christi Bays (Odum and Hoskin, 1958). Increases in turbidity can reduce light penetration in seagrass beds and reduce productivity (Odum, 1963). Phytoplankton and seagrasses are dependent upon sunlight as the energy source for photosynthesis, and as the suspended sediment content of the water increases, the depth of light penetration decreases, resulting in decreased abundance and productivity of the phytoplankton and seagrasses. In turbid bays, much of the available light is being absorbed by clay particles before reaching phytoplankton plant cells (Odum and Hoskin, 1958). Increased turbidities can also reduce dissolved oxygen values (Sherk, 1971). The reductions have been attributed to an oxygen demand exerted by increased suspended sediment and the reduction of light penetration, which reduced photosynthetic oxygen production (Sherk, 1971).

Additions of inorganic nutrients into the water column may indirectly stimulate photosynthesis (Odum and Wilson, 1962). Where respiration is in excess of photosynthesis, inorganic nutrients may accumulate, stimulating photosynthesis. Thus a turbid mixture of organic and inorganic matter both interferes with photosynthesis by shielding light and stimulates photosynthesis by indirectly raising inorganic nutrient levels (Odum and Wilson, 1962).

Plankton, benthic, and nekton standing crops are dependent on turbidity levels. Flood events may resuspend and transport sediments, increase turbidity, and cause a rapid decrease in the standing crop of phytoplankton, zooplankton, benthos, and nekton (Texas Department of Water Resources [TDWR], 1980b). The time necessary for the recovery of the estuarine system after such an event is governed by variables such as season of year and temperature (TDWR, 1980b). Gilmore and others (1976) reported that benthic standing crop and species diversity in the Lavaca Bay estuary were significantly related to turbidity—species diversity was negatively correlated with turbidity and benthic standing crop was positively correlated with turbidity. The positive correlation between benthic standing crop and turbidity may be related to an increase in nutrients brought into the estuarine system by increased river inflows (Gilmore and others, 1976; Harper and Hopkins, 1976).

Turbidity is clearly an important factor in determining some molluscan communities. Suspension-feeding bivalves feed most effectively in relatively clear water (Loosanoff, 1962). Heavy concentrations of turbidity-creating substances can be lethal to bivalves, because they are not able to respire or feed normally (Loosanoff, 1962). The oyster, *Crassostrea virginica,* is able to tolerate turbid conditions with the assistance of well-developed mantle margins that mimic siphonal tubes, with discriminatory palps, with additional cleaning currents, and with more quick muscle fibres in the adductor muscle (McLusky and Elliott, 1981). Mackin (1961) reported that turbidities of up to 700 ppm produced no significant mortalities in experiments on *Crassostrea.* *Ostrea edulis* cannot tolerate conditions as turbid as those accepted by *Crassostrea* (McLusky and Elliott, 1981). Stora and Arnoux (1983) reported that when the ratio of freshwater to sediment volume discharged (the inverse of suspended sediment concentration) was lower than 1,500, mollusks can die. Even when the freshwater to sediment volume is less than 10,000, high concentrations of suspended matter may prove stressful to suspension-feeding mollusks and lower their resistance to other environmental factors (Stora and Arnoux, 1983).

Deposit-feeding organisms, by reworking muddy sediments, may create an unstable bottom and turbid water conditions that are unsuitable for suspension feeders. This kind of exclusion of one trophic group by another is termed trophic-group amensalism (Rhoads and Young, 1970). The feeding activities of deposit feeders may result in high biogenic reworking rates producing a fluid, fecal-rich surface easily resuspended by waves or by low-velocity tidal currents. Instability of the bottom, resulting in high turbidities, may inhibit the growth of most suspension feeders and reduce infaunal diversity (Rhoads and Young, 1970; Aller and Dodge, 1974). Unstable bottoms can also cause high larval mortality for settled, suspension-feeding larvae (Aller and Dodge, 1974).

Effects of Marine Grasses on Sedimentation

Marine grasses or seagrasses form the basis of many estuarine ecosystems. The grassflat environment exhibits high biologic productivity (Odum, 1963) and has long been recognized as an important source of food and shelter for benthic macroinvertebrates, attachment sites for epifauna and epiphytes, nursery grounds for fishes, and direct food sources for some animals, including migratory waterfowl, sea urchins, green sea turtles, manatees, and a variety of herbivorous and juvenile fish (Cornelius, 1975; Stoner, 1980; Ward and others, 1980). The infauna of seagrass communities is generally much more diverse and dense than that of surrounding unvegetated areas (O'Gower and Wacasey, 1967; Santos and Simon, 1974; Orth, 1977; see sections on Biomass/Density and Species Diversity and Species Richness in this report).

Seagrasses and their diverse faunal communities serve as major sources of detrital material, dissolved organic matter, and nitrogen (Tenore, 1977; Kenworthy and others, 1982). The plants have the ability to remove both nitrogen and phosphorus from the water surrounding their leaves and roots (Short and Short, 1984), and through decomposition and retention of organic matter from animal and plant detritus, they make substantial contributions to the sedimentary nitrogen pool (Kenworthy and others, 1982). Increasing the nitrogen supply increases the productivity of plants and animals in an estuary (Jordan and Valiela, 1982).

Five species of seagrasses occur on the Texas coast, including *Halodule wrightii* (= *H. beaudetti*) (shoalgrass), *Ruppia maritima* (widgeongrass), *Halophila engelmannii, Thalassia testudinum* (turtlegrass), and *Cymodocea filiformis* (manateegrass). Seagrasses occur in many areas along the bay margins of barrier islands and near tidal inlets but are much less abundant in the upper bays and near river deltas. Marine grasses are most abundant on the Texas coast in both upper and lower Laguna Madre (White and others, 1983). All five species found on the Texas coast occur in lower Laguna Madre (White and others, 1986). On the middle and upper coast, marine grasses are most abundant in the shallow waters of Redfish and Port Bays in the Corpus Christi area (White and others, 1983), Espiritu Santo Bay in the Port Lavaca area (White and others, 1989), and Christmas Bay in the Galveston area (W. Pulich, personal communication, 1988).

Species distribution is primarily a function of temperature, salinity, tidal regimes, water depth, and turbidity. Other factors affecting distribution include compactional and human-induced subsidence and eustatic sea-level rise. Hurricanes also greatly affect seagrass distribution and density.

Salinity is probably the most significant environmental factor controlling distribution. McMillan and Moseley (1967) and McMahan (1968) determined the general salinity tolerance of the various species. *Halodule wrightii* was successfully grown in salinities of 60 ppt and flourished best in a salinity of 44 ppt. The growth rate of *Cymodocea* indicated that it had the least salinity tolerance of the five seagrasses: growth terminated at a salinity of 45 ppt under controlled conditions. Chin (unpublished manuscript, 1977; citing McMillan and Moseley [1967]), ranked the various grasses from most to least salt tolerant: *Halodule, Thalassia, Ruppia,* and *Cymodocea*. Studies of *Halophila* were inconclusive, but its salinity tolerance is suspected to be between that of *Halodule* and *Cymodocea*. *Halodule* and *Ruppia* are most able to tolerate low salinities (Armstrong and Gordon, 1979); however, few marine grasses grow near areas of fluvial input.

Turbidity is a dominant feature in Texas bays and estuaries with organic-laden muds released in large quantity from the rivers, from dredging, from resuspension of the bottom sediment during storms, and from beach erosion (see Turbidity section in this report). Increases in turbidity can impair light penetration thus limiting primary productivity and reducing photosynthetic oxygen production. During dredging activities in Redfish Bay, Texas, Odum (1963) noted that primary production in turtlegrass beds decreased, and there was an imbalance of respiration over photosynthesis. This may have been caused by silts that were resuspended during dredging. However, diminished productivity was not permanent, since Odum (1963) reported that growth was exceptional the year after dredging, perhaps because dredging released nutrients.

Clay particles in suspension tend to filter out the shorter wave lengths of light (Conover, 1964). There is evidence that terrestrial plants fail to develop normal reproductive organs if grown in sunlight without the higher frequency portion of the visible spectrum (Conover, 1964). The reproductive sterility of many seagrasses on the Texas coast may be related to the filtering effect of clay particles that absorb the shorter wavelengths of light (Conover, 1964).

Most seagrasses on the Texas coast grow best at depths of less than 2 m (6.6 ft) (McMillan, 1984). Conover (1964) reported that *Halodule wrightii* and *Thalassia testudinum* grew best in water no deeper than 1.5 m (4.9 ft) in the Aransas and Redfish Bay area. *Cymodocea filiformis* grew well to depths of 2 m (6.6 ft) in lower Laguna Madre from Port Mansfield to Port Isabel (Conover, 1964). Light appeared to be the only factor that diminished with depth and may have limited seasonal growth in the deeper water. Species that extend into shallow water must also be tolerant of high temperatures. Of the five species that occur on the Texas coast, *Halodule wrightii* has the greatest heat resistance and can extend into shallow, in-shore sites that may be exposed to air at low tides (McMillan, 1984). *Thalassia* occurs at intermediate depths, and *Cymodocea* occurs in areas that are most likely to remain submerged. *Halophila* may occur in shallow areas as well as in areas in the depth range of *Thalassia* (McMillan, 1984).

Recent aerial photos show that the areal extent of seagrasses is increasing in some areas along the Texas coast (for example, along the bay margin of Mustang Island and the flood-tidal delta near Pass Cavallo) probably as a result of relative sea-level rise caused by both natural compactional and human-induced subsidence and eustatic sea-level rise (White and others, 1983; 1989). These factors tend to raise water levels, thereby decreasing the width of wind-tidal flats and leading to the more extensive and relatively constant inundation that favors the establishment of seagrasses (White and others, 1989).

In contrast to factors causing a gradual increase in the areal extent of seagrasses, hurricanes have contributed to their sudden decline. For example, grassflats that were once present along Matagorda Peninsula, just southwest of the Colorado River delta, have virtually disappeared since Hurricane Carla in 1961 (White and others, 1988). The loss of marine grasses may be related to hurricane washover processes and the resulting deposition of sediments along the bayward margin of Matagorda Peninsula (White and others, 1989).

Marine grasses alter the sedimentation processes by increasing sedimentation rates, by concentrating preferentially the finer particle sizes, and by stabilizing the deposited sediments (Burrell and Schubel, 1977). Studies by Ward and others (1984) and Almasi and others (1987) indicate that sedimentation rates are substantially higher in seagrass communities than in unvegetated areas. Leaf and root structures effectively attenuate waves and baffle tidal currents, leading to increased deposition and consolidation of sediments and reduced resuspension (Ginsburg and Lowenstam, 1958; Wayne, 1974; Kemp and others, 1984; Ward and others, 1984). The reduction in resuspension is a direct function of plant biomass (Kemp and others, 1984). The efficiency of seagrasses in baffling the current flow and removing fine suspended particles depends primarily upon the leaf structure of the species and upon plant density (Burrell and Schubel, 1977; Harlin and Thorne-Miller, 1982). Ward and others (1984) estimated that sediment accumulated in seagrass beds at 2-3 mm/mo over a 6-month growing season. However, the fate of trapped sediment during plant senescence in the fall and winter is uncertain, especially in view of intense winter storms. The trapped sediments are probably held in place for at least 6 to 8 months, thus increasing water clarity during the most productive part of the year (Kemp and others, 1984).

Almasi and others (1987) reported that during the summer in the Indian River Lagoon in southeastern Florida, the mean weight percentages of mud trapped in artificial and natural seagrass (*Thalassia*) areas were always significantly greater than the amount of mud trapped in grass-free areas. In the summer, the average depositional rate was 4.96 g of mud per week trapped in the *Thalassia* bed and 3.04 g of mud per week trapped in the sandy, grass-free area; during the winter, these rates were 1.60 and 1.50 g /week, respectively. The amount of mud retained in grass-free and grassy areas was always greater in summer than winter as a result of the effects of runoff, bioturbation, wind speed, and direction: (1) Rain, surface runoff, and river discharge decrease during the winter, causing a reduction in the amount of suspended particles. (2) Increased biological activity during the summer causes destabilization of the bottom sediment, reduces erosion velocities, and thereby enhances resuspension. (3) Even though winter winds are strong, their direction has a short fetch, and resuspension and settlement of sediment decrease during this time. Alamasi and others (1987) argue that reduction of sediment accumulation in seagrass beds during the winter is not a result of winter die-off that subsequently reduces the baffling effect of the grass blades. Partial winter die-offs do occur but not in their artificial seagrass plot. Therefore, the decreased winter mud flow in artificial grass beds must be due to decreased particle supply, not from a decreased baffle effect (Almasi and others, 1987).

Damming of rivers may result in a decrease in streamflow, which could lead to inadequate flushing of pollutants from seagrass beds. Since the harnessing of the Rhone River in France has been completed, floods are rare and weaker than before (Peres and Picard, 1975). The accumulating clay minerals adsorb some anionic detergents when suspended in sea water and later desorb them in interstitial water (Peres and Picard, 1975). Comparison of *Posidonia* beds submitted to high sedimentation rates of clay in unpolluted and polluted areas on the French Mediterranean coast reveals that the disappearance of the beds occurs only in areas that were polluted by domestic sewage containing detergents, and this does not happen in areas where bottom currents are sufficient to prevent high sedimentation rates.

Seagrass sediments generally contain a higher percentage of silt and clay and organic carbon and are more poorly sorted than nonseagrass sediments (Lynts, 1966; Orth, 1973; Orth, 1977; Grady, 1981; Hoskin, 1983). Lynts (1966) found a close correlation between sediment size and density of turtle grass (*Thalassia testudinum*). Finer-grained sediments were found in areas of densest growth, about equal ratios of sand and mud in regions of dense to moderate growth, while the coarsest sediments were found in regions of patchy growth (Lynts, 1966). Hoskin (1983) found more silt in *Thalassia* sediments than in nearby sediments with bare sand. Orth (1973; 1977) reported that particle size and degree of sorting decreased and organic carbon increased from the edges of an eelgrass (*Zostera marina*) bed to an area where the eelgrass was most dense. Organic carbon content in sediments with *Halodule* and *Thalassia* within the intertidal zone was 1.9 times higher than that of adjacent sand flats but much less than that in the subtidal seagrass meadows (Grady, 1981).

Trap Efficiencies of Bays

Much has been written with regard to various sources of sediments and whether bays and estuaries are permanent sinks for most sediments. There is evidence that many estuaries and bays may be relatively efficient in trapping sediments delivered to them. For example, Conomos and Peterson (1976) concluded that only about 6 percent of the annual riverborne load delivered under normal discharge conditions to San Francisco Bay was lost to the ocean. Biggs and Howell (1984) examined various estuaries using a modified version of a ratio formulated by Brune (1953) to measure the trapping efficiency of reservoirs. Brune (1953) compared the volume of a reservoir with the volume of incoming river water to estimate the amount of sediment that would be trapped, and therefore, the expected life of the reservoir. A reservoir large enough to hold one-hundredth of the annual volume of inflowing water could trap about half of the incoming sediment, and one large enough to hold one-tenth of the annual inflow could trap 80 to 90 percent of the sediment influx (Brune, 1953; Meade 1982). By applying this concept to estuaries, Biggs and Howell (1984) found that the sediment trapping ability of estuaries could be estimated (Fig. 82). The ratio C/I is a comparison of the volume (capacity; C) of the estuary to the total potential freshwater inflow (I), which is the product of the watershed area and the annual precipitation. The C/I ratios of various estuaries outside Texas (Biggs and Howell, 1984) and of some Texas bays and estuaries are shown in table 12. These data suggest that Sabine Lake, which has a ratio similar to that of Mobile Bay, may have the lowest trapping efficiency of any Texas bay, around 60 to 70 percent. However, a complicating factor in Sabine Lake is the Port Arthur and Sabine–Neches Canal network, through which as much as 80 percent of the fresh-water discharges of the Sabine and Neches River flow to

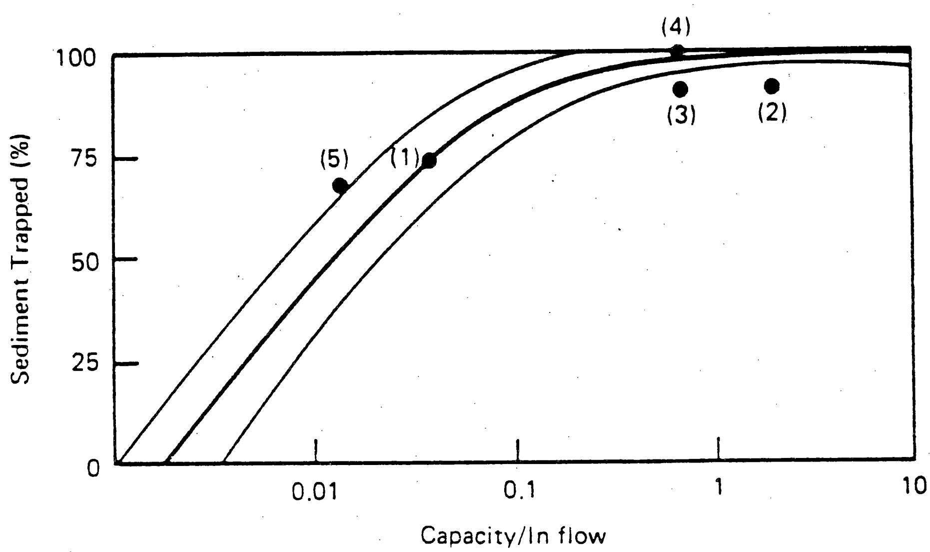


Figure 82. Sediment trap efficiency curves of Brune (1953) (see figure 37) applied to estuaries. Measured trapping efficiencies for selected estuaries are plotted in the figure: (1) Chesapeake Bay; (2) Rappahannock River; (3) Choptank River; (4) James River; and (5) Mobile Bay. (From Biggs and Howell, 1984; reprinted by permission.)

Table 12. Comparisons of predicted trapping efficiencies of other estuaries with those in Texas. Data for estuaries outside of Texas from Biggs and Howell (1984); data on estuarine volume and inflows for Texas from Armstrong (1982), based on data from Texas Department of Water Resources (1980 a, b, and 1981 a, b, c), and Currington and others (1966). Percent sediment trapped estimated from Brune (1953).

Estuary	Volume (km^3) (C)	Inflow (km^3/yr) (I)	C/I	Predicted Trapping Efficiency (percent)
Narragansett Bay, RI	2.4	6.24	0.4	95 ± 5
Long Island Sound, NY	60.8	46.2	1.3	97 ± 3
New York Bay	2.3	41.8	0.05	77 ± 10
Delaware Bay	19.4	36.3	0.5	98 ± 3
Chesapeake Bay	80.5	121	0.7	98 ± 3
Patuxent River	6.1	2.42	2.5	> 99
Pamlico Sound, NC	0.9	14.3	0.06	80 ± 10
Apalachicola Bay, FL	0.4	61.6	0.006	30 ± 15
Mobile Bay, AL	3.2	150	0.02	61 ± 12
Barataria Bay, LA	0.3	6	0.05	76 ± 10
San Francisco Bay, CA	2.5	112	0.02	61 ± 12
Texas Estuaries				
Sabine-Neches	0.33	16.1	0.02	61 ± 12
Trinity-San Jacinto	2.9	12.1	0.24	95 ± 5
Lavaca-Tres Palacios	2.1	3.6	0.59	98 ± 3
Guadalupe (San Antonio)	0.75	2.8	0.27	95 ± 5
Nueces (Corpus Christi)	1.15	0.84	1.37	97 ± 3

the Gulf of Mexico and thus bypass Sabine Lake (Ward, 1973, cited in Gosselink and others, 1979). Other Texas bays, based on C/I comparisons characterizing other estuaries, appear to have trapping efficiencies exceeding 90 percent. Maps of shoaling rates in Texas estuaries (Shepard, 1953) indicate that most of the river-derived suspended sediments may be deposited in the upper half of the bays. However, estimates by the Corps of Engineers (USACE, 1942) of the disposition of fluvial sediments in the Trinity–Galveston Bay system indicate that about 40 percent of the riverine sediments pass through the bay and into the Gulf of Mexico. Wilkinson and Byrne (1977) concluded that much of the fluvial sediment delivered into Lavaca Bay was transported through the bay and into Matagorda Bay. The USDA (1972) estimated that 60 to 70 percent of the total sediment delivered to Texas bays and estuaries is retained, while the remainder is transported into the Gulf.

In a study of the expected delta progradation at the mouth of the Colorado River when it is diverted into Matagorda Bay (USACE, 1981), van Beek and others (1980) concluded that about 7 to 8 percent of the incoming suspended load (there is little or no bed load because of a sediment trap upstream) would be retained in the area of the delta and the remainder would be carried out of the bay (eastern arm of Matagorda Bay).

Holmes (1982) postulated that during frontal passage a substantial amount of suspended sediment from the bays and estuaries along the central and upper Texas coast is transported out of the bays through tidal inlets and is deposited on the Continental Shelf (Holmes' model is discussed more fully on p. 166). White and others (1983; 1985) found some supporting evidence from trace metal concentrations in sediments on the shelf, suggesting that the sediments were derived from adjacent bays. Major flood events that apparently transport the largest volumes of sediments may push substantial amounts of sediment out to sea, at least in Gulf Coast estuaries.

Importance of Episodic Events to Fluvial-Estuarine Sedimentation

The importance of extreme events such as hurricanes and other storms in transporting fluvial sediments into estuaries has been reported by several investigators (for example, Schubel and Meade, 1977; Gross and others, 1978; Milliman and Meade, 1983). According to Schubel and Meade

> Only a few river sediment stations have been in operation long enough to have documented the extreme events that are so important in the introduction of sediment: events such as the hurricane flood of August 1955 when the Delaware River carried more sediment past Trenton in two days than in all five years combined in the mid-1960's drought; or the three days in December 1964 when the Eel River in northern California transported more sediment than in the preceding eight years; or the week following Tropical Storm Agnes in June 1972 when the Susquehanna discharged 20-25 times as much sediment as during the previous year. Events of this magnitude occur only rarely—a few times a century at most—but their importance to estuarine sedimentation is so great that programs should be designed to record their effects when and where they do occur.
>
> (Schubel and Meade, 1977; p. 205)

In a study of sediment loads of the Susquehanna River delivered to Chesapeake Bay, Gross and others (1978) found that of 50 million tons of suspended sediment discharged by the river during a 10-yr period (1966–1976), approximately 80 percent was discharged in only 10 days. Two extreme events, Tropical Storm Agnes (1972) and Hurricane Eloise (1975), were responsible. Apparently the sources of much of the contributed sediment were reservoirs along the Susquehanna River that were flushed during the events (Gross and others, 1978; Meade, 1982). The sediment from the reservoirs represented about 20 years of storage (Meade, 1982). Meade (1982) cautioned about extrapolating these findings to other areas due to the fact that the reservoirs are narrow, and an independent check needed to be made of the observations.

Van Heerden and Roberts (1980) and van Heerden and others (1981) reported evidence of the importance of annual flood events in the subaerial growth of the Atchafalaya Delta in Louisiana. Larger increases in area occurred following floods during the 1970's.

The importance of episodic flood events has also been reported in Texas coastal fluvial-estuarine areas. In a study of the suspended load of the Trinity River in 1965, Rice (1967) determined that approximately 80 percent of the year's sediment discharge occurred during two major floods; the larger flood, which carried 1,770 acre-ft of suspended sediment, accounted for 59 percent of the annual total. The sediment load carried by the Colorado River near Matagorda, Texas, during two floods in 1979 totaled 970 acre-ft, of which about 2 percent was bed load (Welborn and Andrews, 1980). Although the total sediment load of the river near Matagorda for 1979 is not known, estimates by the Corps of Engineers indicate an annual average of about 1,350 acre-ft (USACE, 1977), which is supported by van Beek and others (1981). The volume of sediment discharged by the two floods monitored by Welborn and Andrews (1980) represents about 70 percent of this annual average. The relationships between discharge and sediment load along the Colorado at Columbus and Matagorda were plotted by van Beek and others (1980) (Fig. 83 A). In analyzing sediment transported by the Colorado River, van Beek and others (1980) found that sediment load frequency graphs indicated that 81 percent of the total annual sediment load was transported by the five percentile classes representing highest river discharge (Figs. 83 B and 83 C).

Scott and others (1969) and McGowen (1971) reported on large contributions of sand to a fan delta on the margin of Nueces Bay that resulted from extensive rains accompanying Hurricane Beulah. The layer of sand deposited was more than 1 m thick in many places and composed a volume of approximately 2.5 × 105 m^3 (5.5 × 105 yd^3) (Scott and others, 1969). Apparently the sediment was deposited in less than a week (possibly a day) while 60 cm (2 ft) of rain fell within the 80 km^2 (30 mi^2) drainage basin of Gum Hollow Creek, the stream that supplies the fan delta with sediment.

The amount of suspended sediment that passes through an estuary and out onto the continental shelf during extreme flood events is not easily determined because of the difficulty of monitoring estuarine systems during such events. In examining the effects of Tropical Storm Agnes on the Rappahannock Estuary (in the Chesapeake Bay system), Nichols (1977b) concluded that 90 percent of the total sediment influx was trapped within the estuary; deposition of sediment occurred mainly in the zone of the turbidity maximum (at the interface of the marine salt wedge and the inflowing fresh water). He further postulated that in large estuaries such as Chesapeake Bay, most flood events are not able to push the salt intrusion out to sea and that floods might actually enhance the ability of the hydraulic regime to trap sediments by intensifying the stratification and convergence

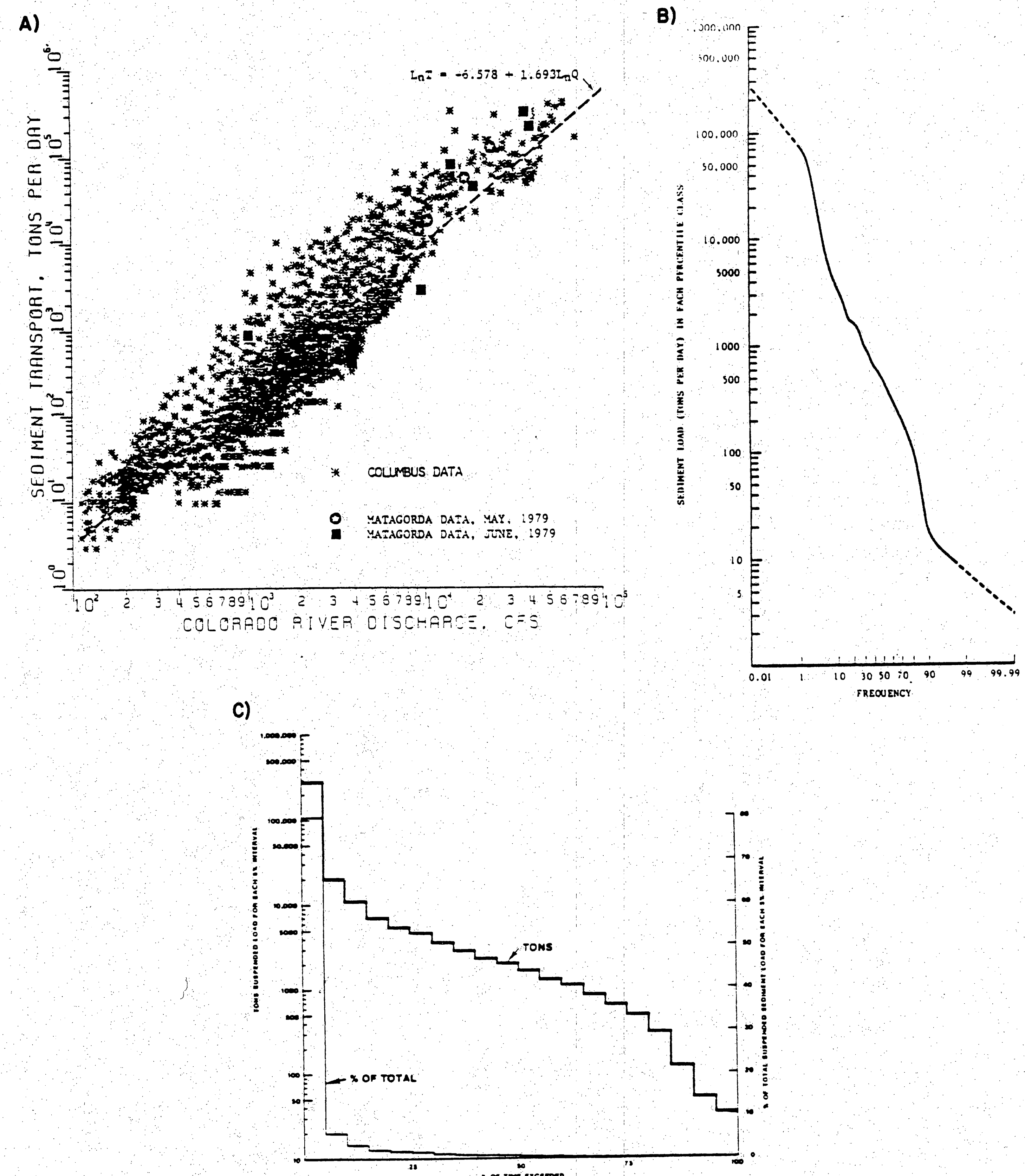


Figure 83. Sediment transport by the Colorado River. (A) Comparison of discharge and sediment load at Columbus (for 1957–1967) with that at Matagorda (May and June 1979); (B) Sediment load frequency of occurrence; (C) Tons of suspended load per class and percent of total suspended load per class by frequency of occurrence. (From van Beek and others, 1980.)

between the fresh-water and salt-water zones. He further concluded that where water depths are increased by dredged channels, entrapment is compounded. Sediment deposition in the estuary as a result of the tropical storm decreased seaward from a thickness of 7.5 mm to 0.5 mm (0.3 to 0.02 in); deposition in a channel in the area was extreme, totaling about 40 percent of the average annual deposition (Nichols, 1977b).

Texas bays are considerably different (shallower, lower tides, more homogeneous with respect to vertical salinities, more protected by barrier islands at their mouths) from Chesapeake Bay (Nichols, 1977b). The trapping efficiencies of bays and estuaries along the Gulf coast may be significantly less during storms. As mentioned briefly on p. 161, Holmes (1982) has proposed that bay systems along the northern Texas coast, including Matagorda Bay, are staging areas for the transport of fine-grained sediments out of the bays and southward onto the Outer Continental Shelf. The proposed delivery events are frontal systems that are common occurrences in the late fall, winter, and early spring. As noted by Holmes (backed by observations of other researchers), as a frontal system approaches the coast, southeasterly (onshore) winds increase in strength, pushing Gulf water into the bays and estuaries; associated waves and currents stir up sediments and increase bay turbidities. As the front passes, usually accompanied by significant amounts of precipitation, the winds shift to the north. The combination of high freshwater inflows and offshore winds acts to flush the swollen bay and estuarine waters loaded with sediments, out of the tidal passes and onto the continental shelf (Fig. 84). The outflow may last for days as a result of the hydraulic head and flooding (Holmes, 1982).

Hurricanes and tropical storms can have even more impact with respect to flushing sediments from the bay–estuary–lagoon system. Isphording and others (1987) reported that more than 83 million tons of sediment, scoured from the bottom of the Apalachicola Bay system during the passage of two hurricanes in 1985, were carried into the Gulf of Mexico by high-velocity currents. According to the authors, the removal of this large amount of sediment significantly extended the life of the bay, which is being filled with sediment. Hayes (1967) mapped a blanket of sediment more than 9 cm (3.5 inches) thick, locally, on the inner shelf, which he attributed to deposition by

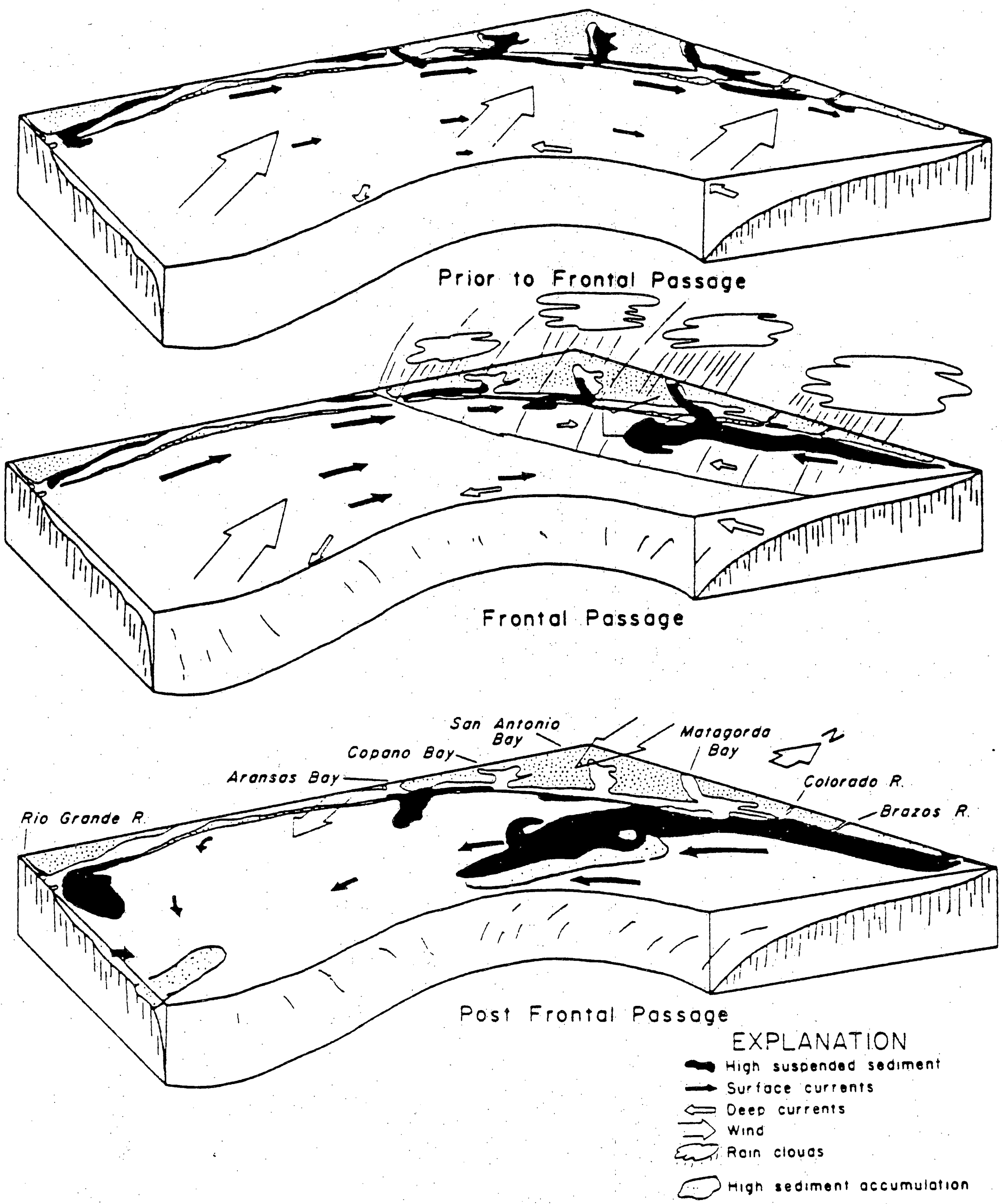


Figure 84. Model depicting possible effect of frontal passage on the transportation and distribution of fine-sediment from the bay-estuary-lagoon system to the South Texas shelf. (From Holmes, 1982; reprinted by permission.)

Hurricane Carla surge waters discharging back to the Gulf (there is disagreement as to the source of this sediment, however; R. A. Morton, personal communication, 1988). Hurricanes also dump considerable amounts of sediment (as manifested in coarse-grained washover-fan deposits at the termini of washover channels) into the bays and estuaries. The net movement of fine-grained sediments is difficult to quantify because of the extreme conditions in which measurements have to be made. However, one would expect that extensive aftermath rainfall that often accompanies hurricanes (for example Hurricane Beulah; Scott and others, 1969) adds considerable amounts of fine-grained sediment to the estuarine system and contributes to the expulsion of sediment-laden water out of the bay. Again, the quantity of suspended sediment brought into the bays during storm surge flooding compared with how much is taken out during ebb flows has not been determined.

In summary, the sand (bed load) delivered into the bays and estuaries by rivers is deposited near the river mouths and trapped because of diminishing current velocities. Sand may be reworked along the margin of the bay and even transported along shore by littoral currents, but it is effectively trapped in the system unless the river fills the estuary with sediment and discharges directly into the Gulf. The fate of suspended sediments (silt and clay) is less clear. Apparently the largest amount of the fluvial sediment is deposited in the upper reaches of the estuary, however, the volume (or percentage of the total supplied) that is transported out of the estuary and into the Gulf cannot be confidently quantified.

Sedimentation in Texas Bays

Bay–estuary–lagoon systems have been characterized by many researchers as efficient sediment traps (Kennedy, 1984). As noted by Biggs and Howell (1984), Emery and Uchupi (1972) concluded on the basis of studies of suspended sediment discharged by rivers (excluding the Mississippi) along the Gulf and Atlantic coasts, that on average, if sea level remained unchanged, the bays and estuaries should be filled in approximately 9,500 yr (or faster if other sources of sediments are included). Average rates of deposition in the bay–estuary–lagoon system, according to Emery and

Uchupi (1972), are about 70 cm/100 yr, or 7 mm/yr (2.3 ft/100 yr, or 0.28 inch/ yr). But this rate is close to the average rate of sea-level rise (80 cm/100 yr [2.6 ft/100 yr]) over the past 5,000 yr, which suggests a balance between shoaling and deepening (Emery and Uchupi, 1972).

As noted in the introductory section on the origin of bays along the Texas coast, the rate of global sea-level rise over the past few thousand years has varied. The average rate of rise of 8 mm/yr (0.3 inch/yr) for the past 5,000 yr (Emery and Uchupi, 1972) is more than the current global sea-level rise of between 1 and 1.5 mm/yr (0.04 and 0.06 inch/yr) (Gornitz and others, 1982). However, when subsidence (Swanson and Thurlow, 1973) is considered in conjunction with global rise, relative sea-level rise averages about 11 mm/yr (0.43 inch/yr) along the central and upper Texas coast; the rates are considerably higher in areas of human-induced subsidence in the Houston area (Fig. 9).

If the average rate of relative sea-level rise is 11 mm/yr (0.43 inch/yr) in the different bays (assuming the bay rates are similar to those documented by tide gauges near the Gulf shoreline), are bays along the Texas coast shoaling (becoming shallower from sediment deposition) or are they deepening (from subsidence and scouring at rates higher than sediment deposition)?

Using unpublished bathymetric data collected by the Coast and Geodetic Survey in the bay–estuary–lagoon system of the Texas coast, Shepard (1953) compared more than 20,500 points where soundings were made during the last century (1852–1875) with those made during the present century (1934–1935). Although there is much variation from bay to bay and within a bay (Fig. 85), he detected, overall, a trend toward bay shoaling at an average rate of about 3.8 mm/yr (1.26 ft/100 yr). Because the silt load of Texas streams is too high for this rate of shoaling, he concluded that the bays must be submerging at a rate of about 5.2 mm/yr (1.7 ft/100 yr) based on the trend set by tide-gauge records along the Gulf coast including Galveston (Marmer, 1954). Thus, Shepard (1953) estimated the total rate of average sedimentation (submergence + shoaling) in the Texas bays to be 9 mm/yr, or 3 ft/100 yr.

In a comparison of the silt load of Texas streams (from river stations as close to the coast as allowed by the sampling network), Shepard (1953) found a correlation between stream sediment

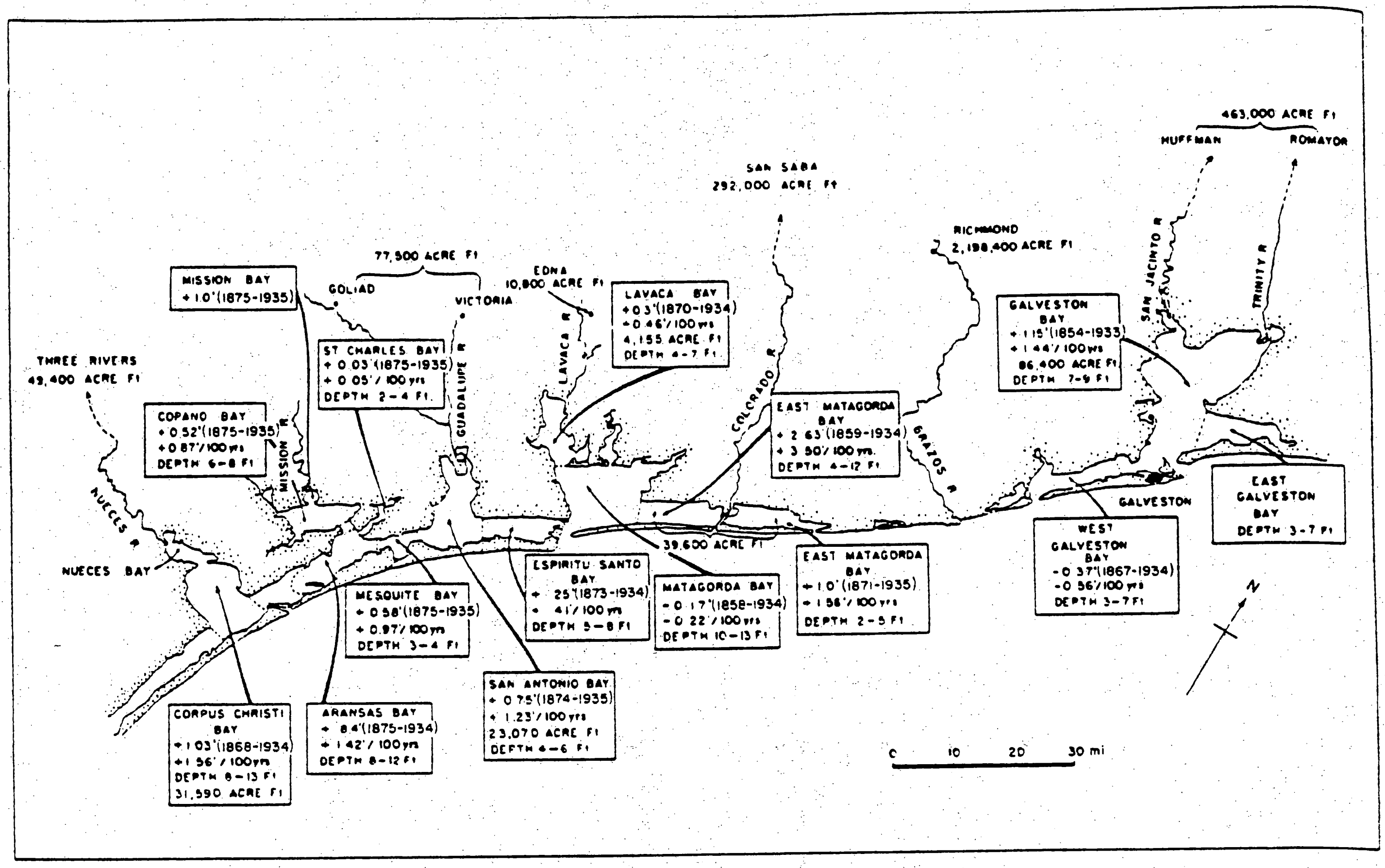


Figure 85. Depth changes in Texas bays. Boxes for the different bays include shoaling (+) or deepening (-), shoaling rate per century (ft), approximate volume (acre-ft) of sediment per century represented by shoaling rate, and depth of main parts of bay (ft). Values for rivers (acre-ft) represent 100 yr at rates for 1950–1951. (From Shepard, 1953; reprinted by permission.)

load and bay shoaling rates (or total bay fill) projected over a century. For example, the Trinity and San Jacinto Rivers, together, transport the largest amount of sediment, and the Galveston–Trinity Bay system, into which they discharge, has the largest amount of bay fill (table 13). The total silt load transported by Texas streams, however, was too high for the amount of bay fill (table 13). This disparity was attributed in part to deposition of sediment in stream floodplains below the stream gauging stations, or behind dams, although Shepard noted that all the large dams were constructed near the end of period between bathymetric surveys. However, he postulated that other sources of sediments probably equalize the stream sediment load not reaching the bays, and suggested that the difference between the silt load carried by the streams and bay fill was an indication of submergence in all bays but Corpus Christi, where bay shoaling and stream sediment load were more balanced. As noted previously, this theory of submergence along the Texas Gulf coast was supported later by Swanson and Thurlow (1973). The Corpus Christi Bay system is apparently undergoing subsidence also, as reflected by tide-gauge records at Aransas Pass (Swanson and Thurlow, 1973).

Comparing the average bay depth of 2.1 m (7 ft) with his average shoaling rate of 3.8 mm/yr (1.26 ft/100 yr), Shepard concluded that most of the Texas bays will be filled in less than 1,000 yr. However, evidence from cores and borings in bay-floor sediments suggests that conditions similar to the present have persisted for thousands of years and that sea-level rise and subsidence could maintain the bays far longer than present shoaling rates indicate. Shepard (1953) also recognized the high water content of recently deposited sediments and the fact that compaction after burial would produce a considerably thinner sedimentary sequence. Shoaling rates of Texas bays and estuaries are not significantly different from sedimentation rates reported in other bay–estuary–systems; however, they are lower than rates reported for Apalachicola and Mobile Bays along the Gulf coast (table 14).

Table 13. Relation of bay fill to suspended load of entering river. (From Shepard, 1953.)

Bay	Solid fill (acre-ft/100 yr)[1]	River and station[2]	"Silt" load (acre-ft/100 yr)	Ratio river load and bay fill[3]
Corpus Christi	31,590	Nueces, Three Rivers	49,400	1.56
San Antonio	23,070	Guadalupe, Victoria and San Antonio, Goliad	77,500	3.36
Lavaca	4,155	Lavaca, Edna	10,800	2.61
East Matagorda	39,600	Colorado San Saba	292,000	7.37
Galveston	86,400	Trinity, Romayor and San Jacinto, Huffmann	463,000	5.36

1 Reduced to 30% of volume because of estimated water content.

2 From Texas Board of Water Engineers (1952).

3 The river "rafts" between 1865 and 1926 (Price and Gunter, 1942) probably led to the diversion of an unusually large amount of sediment to the flood plain, thus increasing this ratio.

Table 14. Bay–estuary sedimentation and net deposition rates.

Bay–Estuary System	Rate (mm/yr)	Source
Chesapeake Bay	3.7	Biggs (1970)
Chesapeake Bay (head of bay)	3–4	Schubel and Carter (1984)
Great Sound	1–5	Thorbjarnarson and others (1985)
Raritan	1.5–3.4	Renwick and Ashley (1984)
Apalachicola Bay	6.7	Isphording (1986)
Mobile Bay	5.6	Ryan and Goodell (1972)
Texas bays:		
Galveston–Trinity	4.4	Shepard (1953)
	3.7	Rehkemper (1969a)
Lavaca	1.4	Shepard (1953)
Matagorda	-0.7	Shepard (1953)
San Antonio	3.8	Shepard (1953)
Corpus Christi	4.7	Shepard (1953)

Contribution of Fluvial Sediments to Bay Sedimentation

Rivers are the primary source of sediments deposited in the bay–estuary–lagoon system, at least from Corpus Christi Bay northward. The importance, in terms of volume contributed, of fluvial sediments is reflected (in addition to table 13) by Shepard's maps of depth changes for the various bay systems. The map of the Galveston–Trinity Bay system is a good example (Fig. 86). The decrease in water depth in the upper half of Trinity Bay compared to other parts of Trinity Bay and Galveston Bay reflects sediment deposition from the Trinity River. Maps of other Texas bays show similar trends, although of lesser relative magnitude (except for the Colorado River).

The Filling of Lavaca Bay—An Example

Byrne (1975) and Wilkinson and Byrne (1977) reconstructed the geologic depositional history of Lavaca Bay. As noted in the section on the origin of Texas bays, valleys that were eroded during the most recent sea-level lowstand were flooded by sea-level rise during the present interglacial stage. The depth of the axis of the incised Lavaca Bay valley (near the mouth of Lavaca Bay) is about 30 m (90 ft) below current mean sea level (Wilkinson and Byrne, 1977; Wright, 1980) (Fig. 87). Deposition of sediment in the valley has taken place over the past 10,000 yr, and consists of (from the base of the valley upward) (1) fluvial sand, (2) deltaic sand and muddy sand, and (3) bay-estuarine mud (with local oyster shells) (Fig. 88). At the head of the bay near the mouth of the Lavaca River, fluvial-deltaic sand has prograded over the estuarine muds (Fig. 88). This sequence is generally similar to that recorded by Rehkemper (1969a) for the Trinity–Galveston Bay valley-fill sequence with a few exceptions (including local, thin peat beds at the top of the fluvial-deltaic sequence in the Trinity–Galveston Bay depositional sequence).

Despite a 25-m (82-ft) rise in sea level, Lavaca Bay has remained a relatively shallow estuary with rates of sedimentation only slightly exceeding flooding rates throughout its depositional history (Wilkinson and Byrne, 1977). Wilkinson and Byrne (1977) constructed a paleobathymetric

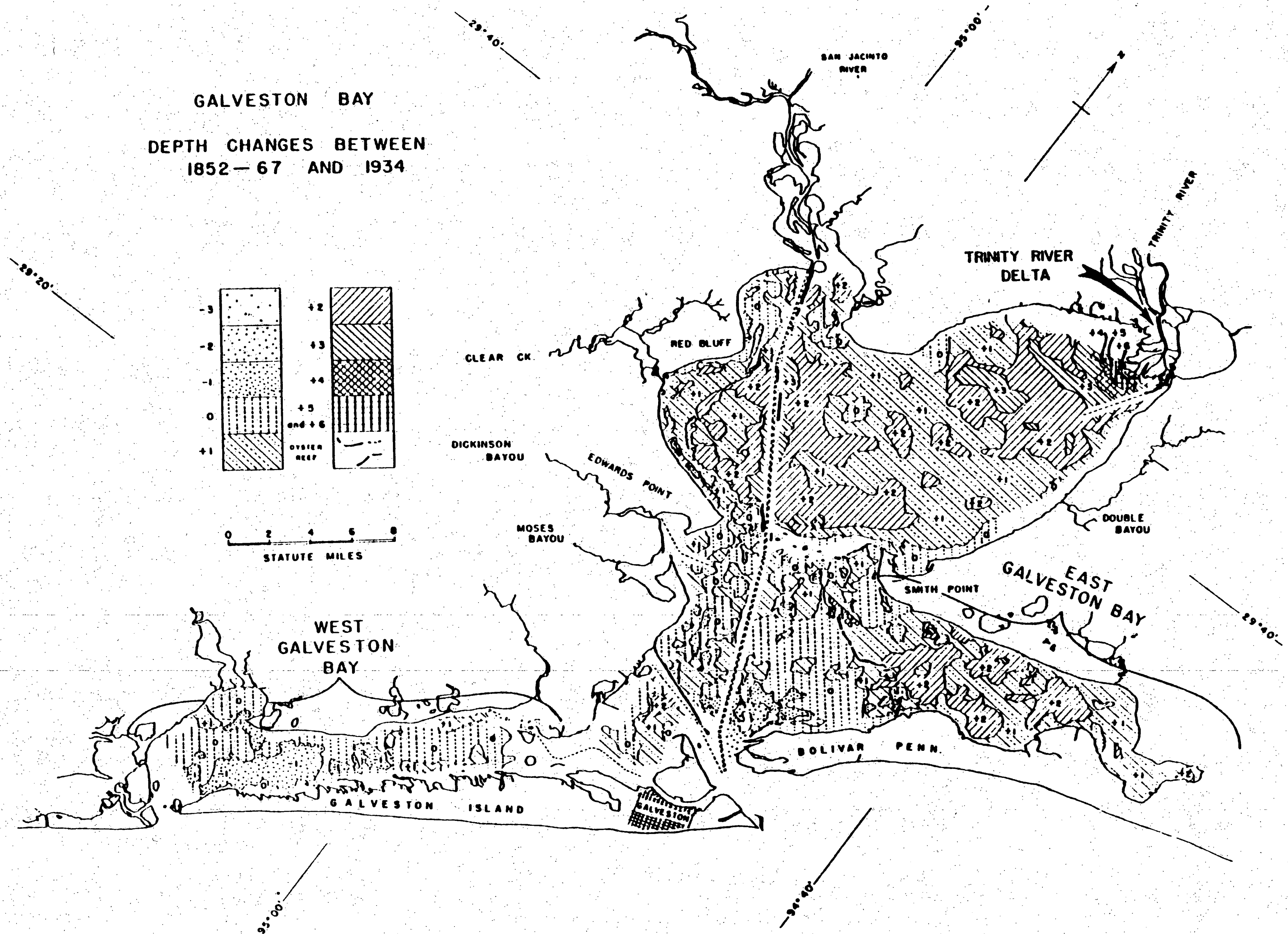


Figure 86. Galveston Bay depth changes determined by comparing bathymetric surveys ranging from 1852 to 1867 with surveys in 1934. Highest rate of fill is near the Trinity River delta. (From Shepard, 1953; reprinted by permission.)

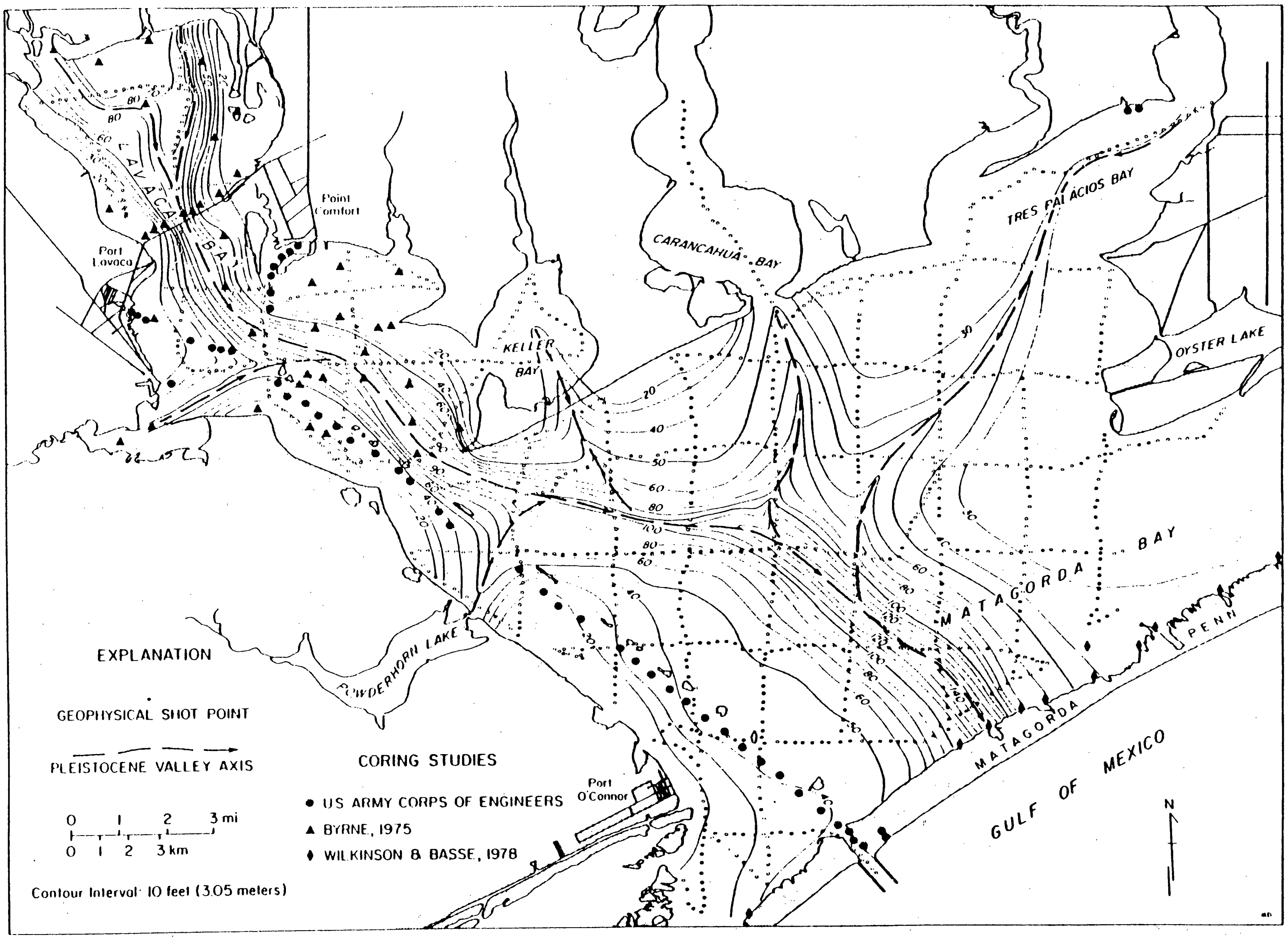


Figure 87. Map showing depth to Pleistocene surface beneath Lavaca and Matagorda Bays. (From Wright, 1980.)

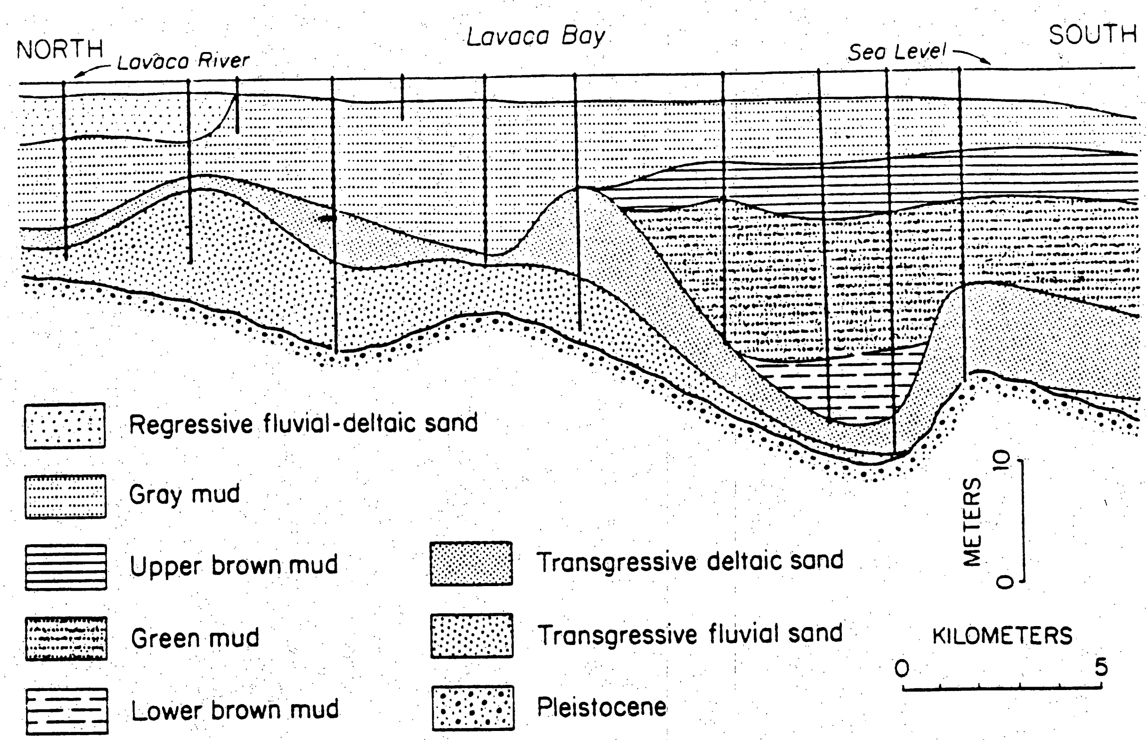


Figure 88. Cross section down Lavaca River valley showing sediments that have accumulated in Lavaca Bay above the Pleistocene surface. (From Wilkinson and Byrne, 1977; reprinted by permission.)

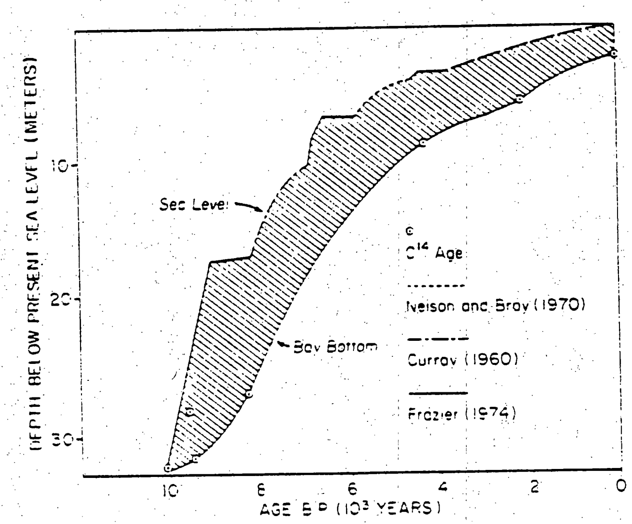


Figure 89. Paleobathymetric curve over the past 10,000 yr for Lavaca Bay. Sea-level positions are maximum values taken from various authors. Positions of the bay bottom are based on radiocarbon dates. Past maximum water depths of Lavaca Bay are represented by the vertical distance between the two curves. Radiocarbon data from Byrne (1975). (From Wilkinson and Byrne, 1977; reprinted by permission.)

curve (of the past 10,000 yr) representing the depth of the axis of the Lavaca Bay valley and its changes through time (Fig. 89). The curve indicates that initial sedimentation (aggradation) rates were high but gradually slowed. This change was attributed to an ever-widening valley, which required larger volumes of sediments to deposit a given vertical unit of sediment (Wilkinson and Byrne, 1977). Water depths were relatively stable, but gradually decreased until about 3,000 yr ago when maximum depths were about 3 m (10 ft). This gradual shoaling indicates that rates of sedimentation exceeded sea-level rise (Wilkinson and Byrne, 1977).

To determine an argillaceous-sediment (predominantly clay-sized material) budget for Lavaca Bay, Wilkinson and Byrne (1977) analyzed the contributions of the various sources of sediments. As mentioned previously, they found the rivers to be the most important source of sediment, supplying about 73 percent of the total. Erosion of bay shorelines, characterized by Pleistocene bluffs, was a secondary but significant source, supplying 24 percent. The remainder was attributed to sediment contributions from Matagorda Bay.

Comparison of geologic sedimentation rates with the historic rate estimated by Shepard (1953) for the Lavaca Bay system indicates that historic rates are higher. Wilkinson and Byrne (1977) suggested that the higher rates over the past several decades may be due to historic agricultural practices (rice farming) in the drainage basin. Price and Gunter (1942) also reported increases in rates of bay–lagoon siltation in some areas along the Texas coast as a result of human activities.

Trends in Relative Sea-Level Rise and Fluvial-Estuarine Sedimentation

Data from Shepard (1953) and Wilkinson and Byrne (1977) suggest that Texas bays are becoming shallower due to infilling, but what are the more recent trends along the Texas coast with regard to subsidence rates and reductions in sediment supplied by major rivers?

It should be emphasized that subsidence rates derived from tide-gauge records vary from location to location along the Texas coast (Fig. 7 A), and the most reliable available data are from

National Oceanic and Atmospheric Administration gauges located near the Gulf shoreline. Longer-term records are most reliable and less likely to be influenced by meterological conditions, such as wind, precipitation and river discharge, atmospheric pressure, and temperature, and other shorter-term factors that produce variations in water levels. The longest continuous tide-gauge records are from Galveston (Pier 21) (Fig. 7 B). This records show an acceleration in relative sea-level rise from 1962–1982 compared to 1942–1962 (Turner, 1987; Penland and others, 1988). Again, Turner (1987) suggests that this acceleration may be temporary and reflective of an oscillation around a longer-term constant rate of rise (Fig. 7 B). For purposes of discussion in the following sections, the recent increases in relative sea-level rise are compared to Shepard's (1953) sedimentation rates, which are based on a period (1850's–1870's to 1930's) when fluvial-sediment contributions to the bay–estuary–lagoon system were at higher rates than at present.

Corpus Christi–Nueces Bay System

The bathymetric surveys used by Shepard (1953) generally encompass a period of time that preceded placement of dams along the major rivers. Shepard recognized that after construction of a dam along the lower Nueces River (Corpus Christi Lake), contribution of fluvial sediment to Corpus Christi Bay ceased to be important in 1930. Using data on suspended sediment loads of the Nueces River (at Three Rivers, upstream from Lake Corpus Christi), Shepard determined that the river transported about 494 acre-ft/yr. More recent measurements of suspended sediment loads along the Nueces River (from a station downstream from Lake Corpus Christi) show an annual rate of less than 19 acre-ft/yr (Texas Water Development Board and Water Commission, published and unpublished data, 1962–1984), or about 4 percent of the amount (translated into an annual mean) reported by Shepard (1953) (table 13).

Shepard noted that of all the bays he investigated, the Corpus Christi Bay system had the closest balance between sediment supply and bay fill (table 13). The stream silt load value used by Shepard (1953) is based on data from the late 1920's to the early 1950's (Fig. 31) and probably reflects larger than natural sediment loads due to human development in the drainage basin. On the other hand, Price and Gunter (1942) reported that increases in bay siltation in some areas of south Texas were probably due to a combination of natural (droughts) and human events (clearing of the land for agricultural purposes and overgrazing) beginning in the 1880's. Soil conservation measures implemented by the Soil Conservation Service before the 1950's may have caused the average annual stream silt-load used by Shepard (1953) to be less than that carried by the Nueces River between 1868 and 1934 (the period between the bathymetric surveys) when land use practices and droughts probably caused higher sediment yields.

Considering the large decrease in sediment supplied by the Nueces River, it seems likely that bay shoaling rates have also decreased substantially from the average of 4.7 mm/yr (0.18 inch/yr) derived from Shepard's bathymetric study (table 14). Even though local increases in bay shoreline erosion may be contributing additional sediment to the bay floor (Morton and Paine, 1984), it is possible that sedimentation rates have fallen behind relative sea-level rise and that a larger percentage of Corpus Christi Bay is becoming deeper through time (about 3 percent of the bathymetric stations surveyed by Shepard showed an increase in depth).

Much of the sediment delivered by the Nueces River is apparently deposited in shallow Nueces Bay (Shideler, 1980). Bay bathymetry, however, has also been affected by excavation of oyster shell as an economic resource. According to Texas Parks and Wildlife Department Shell Dredger's Reports, more than 1.9 million m^3 (2.5 million yd^3) of shell was dredged from Nueces Bay from 1969 to 1975 (Kier and White, 1978). This is equivalent to about 1,550 acre-ft of material, or about a 47-yr supply of suspended sediment from the Nueces River (based on an average annual supply of 32.7 acre-ft/yr from 1969 to 1975, measured at the Mathis station below Lake Corpus Christi; Mirabal, 1974, and Dougherty, 1979). Dredging of shell material apparently increased the depth of Nueces Bay, at least locally.

San Antonio–Espiritu Santo Bay System

Comparison of depth changes between 1873–74 and 1934–35 for San Antonio and Espiritu Santo Bays indicates shoaling rates of 37.5 cm/100 yr (1.23 ft/100 yr) and 12.5 cm/100 yr (0.41 ft/100 yr), respectively (Fig. 85; Shepard, 1953). The considerably higher rate in San Antonio Bay is reflective of the fluvial sediment delivered by the Guadalupe and San Antonio Rivers. Shepard (1953) reported that, together, these rivers supplied about 775 acre-ft/yr of suspended sediment, based on measurements at Victoria and Goliad (Fig. 85). More recent data indicate that the annual volume of suspended sediments has decreased slightly (table 15), but not at the magnitude documented for other Texas rivers such as the Trinity, Brazos, Colorado, and Nueces rivers (Figs. 21–26 and 31–32).

Table 15. Annual average suspended load of the San Antonio and Guadalupe Rivers combined (Goliad and Victoria measuring stations; the period of record for the Guadalupe River station at Victoria began in 1946).

Water years	Average annual load (acre-ft)[1]
1943-1951	751.9
1951-1961	657.5
1961-1971	523.5
1971-1984	669.4

[1]Data compiled from Stout and others (1961), Adey and Cook (1964), Cook (1967, 1970), Mirabal (1974), Dougherty (1979), and unpublished records available from the Texas Water Development Board (1976-1984).

In San Antonio Bay, Shepard (1953) noticed that the greatest shoaling had occurred in the upper bay. In the central bay, less shoaling had occurred on the eastern half compared to the western half, but this trend was reversed in the lower part of the bay (gulfward of the Intracoastal Waterway) where less shoaling had occurred on the western half relative to the eastern half.

In contrast to San Antonio Bay, Espiritu Santo Bay has no important fluvial source of sediments, and the depths remained relatively constant with the exception of shoals at each end of the bay (Shepard, 1953). Although, overall, there was a slight decrease in depth in Espiritu Santo Bay (12.5 cm/100 yr [0.41 ft/100 yr], which is equivalent to a shoaling rate of 1.3 mm/yr [0.05

in/yr]—considerably lower than the 3.8 mm/yr [0.15 in/yr] for San Antonio Bay), Shepard (1953) reported that at about 28 percent of the 905 notations where depths were compared, a slight deepening had occurred. This percentage is high compared to most other Texas bays.

Since the analysis by Shepard, depths of San Antonio Bay have been affected locally by the excavation of oyster shell. Records of the Texas Parks and Wildlife Department show that more than 28 million m^3 (36 million yd^3) of shell material was mined between 1969 and 1983 (White and Morton, 1987). This is equivalent to about 22,336 acre-ft of material, which is a volume equal to about 33 yr of suspended load supplied by the Guadalupe and San Antonio Rivers at the average rate of 670 acre-ft/yr (based on the period 1971–1984). This volume is also very close to the amount of bay fill per century (23,070 acre-ft) estimated by Shepard (1953) for San Antonio Bay. Depths in the bay have undoubtedly been increased locally by shell removal.

Comparison of bay shoaling to sediment supplied by the San Antonio and Guadalupe Rivers by Shepard indicates that the ratio of river load to bay fill is about 3.4 to 1 (table 13). Differences in these volumes are attributed to deposition of sediments up stream in channels and floodplains, and transportation of sediments out of the bays into the Gulf; however the principal reason for the difference is attributed to subsidence or submergence (Shepard, 1953).

Much of the fluvial sediment delivered by the rivers is apparently being deposited in Mission Lake (Fig. 64); when the lake is filled, the Guadalupe River delta could resume its progradation down the axis of San Antonio Bay (it should be noted, however, that with the exception of progradation of the Traylor Cut subdelta, and shoreline erosion, little change has been noted in the configuration of the Guadalupe delta since the mid-1800's) (Shepard, 1953; Donaldson and others, 1970; White and Morton, 1987). Donaldson and others (1970) reported that when the Traylor Cut subdelta began to prograde into Mission Lake in 1935, the lake was about 1.5 m (5 ft) deep, and by 1965 it was only 0.6 m (2 ft) deep. Citing the shoaling data presented by Shepard and Moore (1960), Donaldson and others (1970, p. 130) state that "the lower part of San Antonio Bay most likely will fill up and become a tidal flat before the Guadalupe delta reaches the barrier island." These researchers predict that if the many variables controlling sedimentation continue without drastic change, the bay will eventually (probably with reference to a geologic timeframe) be filled and the river will transgress the barrier island and discharge into the Gulf.

According to the Bureau of Reclamation (USDI, 1978) on the San Antonio–Guadalupe River basins, Childress and Bradley (1975) hypothesized that the amount of fluvial sediments being delivered to San Antonio Bay estuary is in equilibrium with the bay system and that changes in the historic load would create an imbalance, which would reduce bay productivity and nutrients and increase erosion.

The shoaling rate determined by Shepard (1953) for San Antonio Bay (3.8 mm/yr [0.15 inch/yr]) is equal to the average (3.8 mm/yr [0.15 inch/yr]) for all the bays he analyzed. Assuming that the average subsidence rate was 5.2 mm/yr (0.2 inch/yr) (based on Shepard's conservative estimate of Marmer's [1954] data), then the sedimentation rate was about 9 mm/yr (0.35 inch/yr), which is equivalent to the average sedimentation rates for all bays (Shepard, 1953). If the more recent average rate of relative sea-level rise in San Antonio Bay is about equal to subsidence rates presented by Swanson and Thurlow (1973) (Fig. 7 A) for Port Aransas (12.8 mm/yr [0.50 inch/yr]) and Freeport (11.2 mm/yr [0.44 inch/yr]) (these rates are about equal to the rate of 11.7 mm/yr [0.46 inch/yr] for Galveston from 1962 to 1982 [Penland and others, 1988]), then the rate of sedimentation may not be keeping pace with submergence, and the bay, overall, could be getting deeper at a possible rate of about 2 to 4 mm/yr (0.08 to 0.16 inch/yr). As noted previously, if Mission Lake fills with sediment, then fluvial sediment loads may be redirected into San Antonio Bay and thus, increase sedimentation rates.

Lavaca–Matagorda Bay System

Wilkinson and Byrne (1977) suggest that sedimentation rates in Lavaca Bay have possibly increased in historic time. This conclusion was reached by comparing Lavaca Bay sedimentation rates during geologic time (based on the total volume of argillaceous sediment in the bay and its accumulation over the past 8,000 to 10,000 yr) with more recent historic sedimentation rates (net rate of 1.4 mm/yr [0.06 inch/yr]) reported by Shepard (1953). The geologic rate is about half the historic rate (Wilkinson and Byrne, 1977). Although several reasons may account for the difference, including rounding of the historic sounding data to the nearest one-half and one-fourth of a foot, and the fact that the recent rates reported by Shepard are for uncompacted sediments, Wilkinson and Byrne (1977) concluded that compaction alone could not account for the difference and that the historic rate is higher possibly because of agricultural development in the drainage basin.

Comparing the silt load volumes transported along the Lavaca River (Edna station) with the volume of fill in Lavaca Bay, Shepard (1953) found that the stream silt load exceeded bay fill by a factor of about 2.5 (table 13); he postulated that the excess stream silt load, which characterized most Texas bays, could be accounted for principally through the process of submergence, or subsidence. Over the geologic timeframe, Wilkinson and Byrne (1977) decided that much of the sediment supplied by the streams discharging into Lavaca Bay was transported into and deposited in Matagorda Bay. In fact, they estimated that the total amount of fill in Lavaca Bay represented only 30 percent of all the sediment available from different sources; the remainder was transported into Matagorda Bay. It is probable that a certain amount reached the Gulf through processes described by Holmes (1982). Shepard (1953) found that during historic time, Matagorda Bay, with the exception of its eastern arm, which was supplied with sediments from the Colorado River, actually became slightly deeper, overall, at an average rate of about 0.7 mm/yr (0.03 inch/yr) (Fig. 85). Recent stream silt-load volumes for the Lavaca River (Edna station) are similar to those reported by Shepard (1953) (108 acre-ft/yr based on records prior to 1952; more recent rates are approximately 100 acre-ft/yr—based on silt loads from published and unpublished records of the Texas Water Development Board and Texas Water Commission, for 1950–1984). The Navidad River and Garcitas Creek are additional sources of fluvial sediments entering Lavaca Bay. The Navidad River source (quantified below) has undoubtedly declined with the completion of Lake Texana in 1980.

Although the upper half of Lavaca Bay decreased in depth (indicating net sedimentation) (Shepard, 1953), the lower half and most of Matagorda Bay had no change (Fig. 90), which suggests an equilibrium in rate of sedimentation and submergence-erosion. Average contribution of the

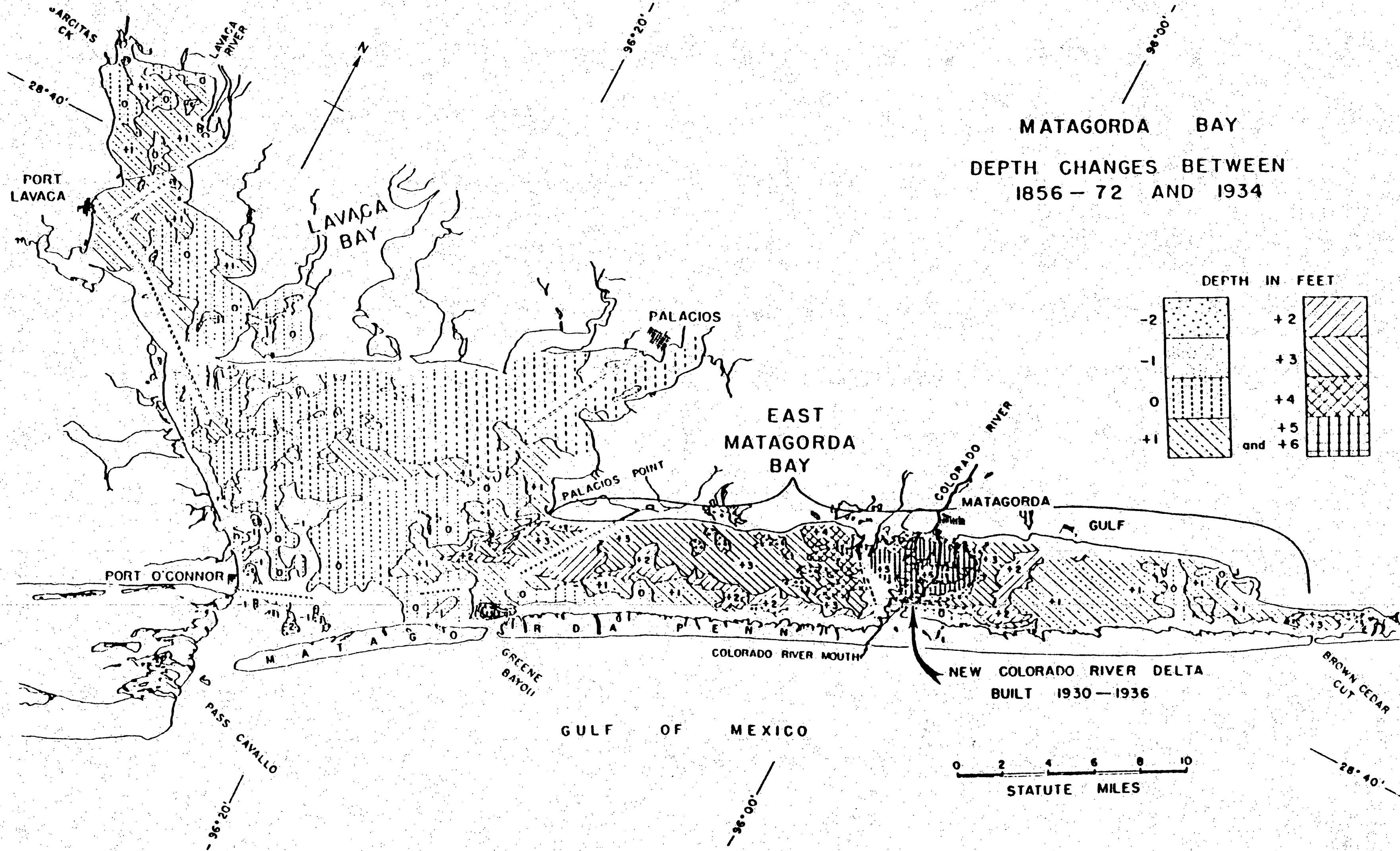


Figure 90. Lavaca–Matagorda bay system depth changes determined by comparing bathymetric surveys ranging from 1856 to 1872 with surveys in 1934. Highest rate of fill is near the Colorado River delta. (From Shepard, 1953; reprinted by permission.)

Navidad River, based on data from 1963 to 1984 collected at the Hallettsville station, was about 26.4 acre-ft/yr (3.26×10^{10} cm^3/yr). This rate of sediment delivery is approximately 20 percent of the geologic rate of sedimentation (annually) in Lavaca Bay, and about 10 percent of the historic rate (Shepard, 1953; Wilkinson and Byrne, 1977). However, Greiner's (1982) data on sediment yield indicate that the Navidad River basin is a much larger source of sediment than measurements at the Hallettsville station indicate (about 75 percent of the drainage basin apparently lies below this station). Calculations of sediment volumes based on Greiner's sediment yield factors show that the respective yields of the Lavaca River and Navidad River drainage basins are 272.6 acre-ft/yr (3.36×10^{11} cm^3/yr) and 783.7 acre-ft/yr (9.60×10^{11} cm^3/yr). These volumes suggest that the Navidad River basin produces about 3 times as much sediment as the Lavaca River basin. The Garcitas Creek area adds another 77.9 acre-ft/yr (9.61×10^{10} cm^3/yr) (Greiner, 1982).

It is not possible to predict the exact effect on future depth changes in the Lavaca–Matagorda Bay system as a result of loss in sediment supplied by the Navidad River, but the sedimentation rate will likely decline and possibly fall behind relative sea-level rise in parts of Lavaca Bay. Using data from Shepard (1953), Wilkinson and Byrne (1977), and Greiner (1982), rough approximations of possible changes in the sedimentation rate can be estimated.

The net sedimentation rate presented by Shepard (1953) for Lavaca Bay is 0.14 cm/yr (0.06 inch/yr). Considering that the area of Lavaca Bay is 2.3×10^8 m^2, Wilkinson and Byrne (1977) translated the vertical rate (0.14 cm/yr [0.06 inch/yr]) into a net rate of sedimentation of 3.2×10^{11} cm^3/yr (1.9×10^{10} $inch^3$/yr) for historic time. Combining rates of sediment yield for the Lavaca and Navidad Rivers and Garcitas Creek, as presented by Greiner, provides a total yield of 1.398×10^{12} cm^3/yr (8.5×10^{10} $inch^3$/yr). Assuming that 30 percent of the fluvial sediment delivered to the bay is trapped in Lavaca Bay (Wilkinson and Byrne, 1977), the total fluvial sediment deposited is 4.20×10^{11} cm^3/yr (2.6×10^{10} $inch^3$/yr). This volume is about 1.3 times the net rate of sedimentation noted above. If this volume is spread over the area of the bay, it yields a vertical sedimentation rate of 0.18 cm/yr (0.07 inch/yr), which is slightly higher than Shepard's net rate of 0.14 cm/yr (0.06 inch/yr). If Lake Texana traps as much as 95 percent of the sediment from the Navidad River Basin (as indicated by the Brune curve, Fig. 37), the total fluvial sediment load (including the Lavaca River and Garcitas Creek) delivered to the bay would be reduced to 4.82×10^{11} cm^3/yr (2.9×10^{10} $inch^3$/yr). Again, assuming a trapping rate of 30 percent, this translates into a sedimentation rate from fluvial sources of 0.063 cm/yr (0.02 inch/yr), or only about 35 percent of the estimated pre-Lake Texana rate. If Lake Texana traps only 32 percent of the river-sediment load (Texas Water Development Board, 1974), this would yield a sedimentation rate of about 78 percent of the estimated pre-Lake Texana rate. The actual reduction in the sedimentation rate in Lavaca Bay as a result of sediment trapped by the lake will presumably fall somewhere between those two extremes.

Planned diversion of the Colorado River into Matagorda Bay (U.S. Army Corps of Engineers, 1981) is expected to increase the sedimentation rate in that bay. Estimated sediment loads that will be contributed to the eastern arm of Matagorda Bay are approximately 1,370 acre-ft/yr (1.69×10^{12} cm^3/yr) (van Beek and others, 1980) The eventual contribution of sediment from the diverted Colorado River to the central part of Matagorda Bay and the mouth of Lavaca Bay, however, is difficult to assess.

Galveston–Trinity Bay System

Subsidence in the Galveston–Trinity Bay system has had a definite impact on bay bathymetry. Shepard (1953) indicated that overall the Galveston–Trinity Bay system (excluding East and West Bays) had undergone shoaling between 1854 and 1933 at an average rate of 4.4 mm/yr (1.44 ft/100 yr). The rate of sedimentation in Galveston and Trinity Bays calculated by Rehkemper (1969a) on the basis of Carbon-14 dates (without considering sediment compaction) was 3.7 mm/yr (1.2 ft/100 yr), which is relatively close to that calculated by Shepard (1953). If the average rate of sea-level rise of about 4.2 mm/yr reported by Marmer (1954) for Galveston (for the period 1909–1937) is assumed to be primarily from subsidence for the bay, then Shepard's rate of sedimentation becomes about 8.6 mm/yr (0.34 inch/yr). The rate of sedimentation near the delta at the mouth of the Trinity River is obviously much higher.

However, more recent subsidence rates in the Galveston–Houston area are much higher than those in the past (Fig. 9). As an example, the rate of subsidence near Morgan's Point at the head of Galveston Bay was approximately 47 mm/yr (1.85 inches/yr) between 1943 and 1978 (Gabrysch, 1984). Of course the rate decreases away from the area of maximum subsidence. Near the center of Trinity Bay the rate is closer to 15 mm/yr (0.59 inch/yr) (for the period 1973 to 1978; Gabrysch, 1984). Still, this rate is higher than the average sedimentation rate of 8.6 mm/yr (0.34 inch/yr) based on Shepard's (1953) data.

Morton and McGowen (1980) compared recent sounding data from the Submerged Lands of Texas project (McGowen and Morton, 1979; White and others, 1985) collected during 1977 with 1968 National Ocean Survey bathymetric data (Fig. 91). This comparison indicates considerable deepening of the Galveston–Trinity Bay system. The actual magnitudes of the depth increases, which are as high as 1.5 m (5 ft) in the upper reaches of Galveston Bay, are not reliable because the soundings in 1977 were not adjusted to a mean sea-level datum. Nevertheless, a comparison of the changes in depths with subsidence (Fig. 9) indicates that the trends or directions of deepening of the bay floor correlate with trends of increasing subsidence.

Of all the bay systems considered by Shepard (1953), the Galveston–Trinity Bay system had one of the largest differences between stream silt load and bay fill (this bay system was surpassed only by East Matagorda Bay, which was the receiving basin for the Colorado River). The combined silt load for the Trinity and San Jacinto Rivers was more than 5 times the calculated bay fill (table 13). Again, factors such as deposition of sediment in stream floodplains, but principally subsidence, probably accounted for the discrepancy, according to Shepard (1953). In the upper half of Trinity Bay near the mouth of the Trinity River, deposition rates were high, as indicated by changes in depth shown in Figure 86. Net sedimentation rates near the margin of the Trinity River delta, in fact, exceeded 20 mm/yr (0.79 inch/yr) and were more than 8 mm/yr (0.31 inch/yr) for much

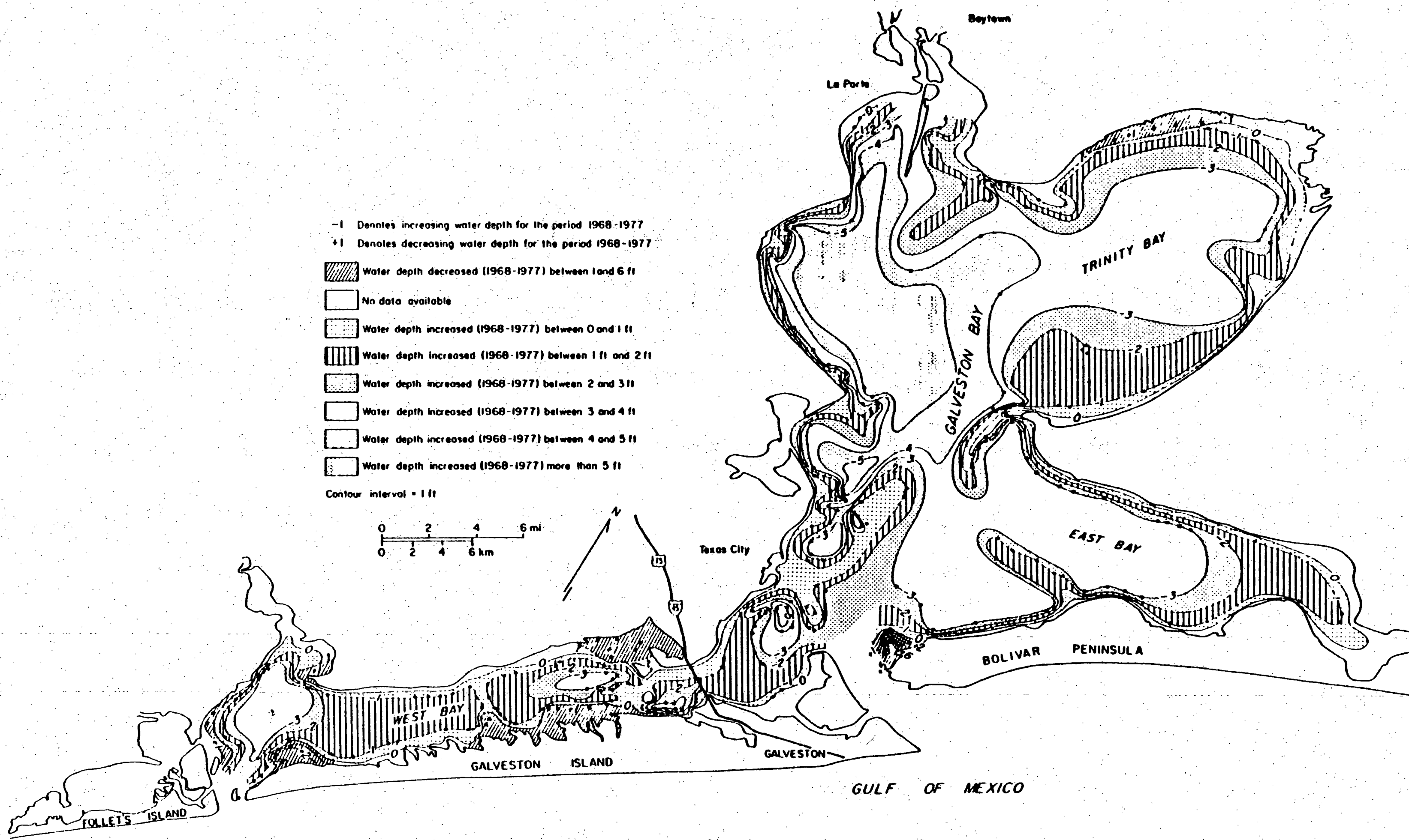


Figure 91. Bathymetric changes for the Trinity–Galveston bay system for the period 1968–1977. Bathymetry for 1968 is from published navigation charts. Bathymetry for 1977 is based on sounding taken at sampling stations, on 1-mi centers, in the fall and winter of 1977. Because the soundings in 1977 were not adjusted to mean sea level, the depth changes shown in this figure must not be taken as absolute. (From Morton and McGowen, 1980.)

of the upper bay. The Trinity River prograded into Trinity Bay about 473 m (0.3 mi) between 1854 and 1933 (Shepard, 1953).

Sedimentation in East Bay was high enough to indicate that sediment from the Trinity River was transported around Smith Point and into the bay (Shepard, 1953). White and others (1985) suggested that trace metals concentrated in East Bay were tied to sediment movement from Trinity Bay (and Trinity River).

In contrast to sedimentation in East Bay, West Bay actually became deeper during the 67-yr period represented by Shepard's data. The rate of deepening was about 1.7 mm/yr (0.07 inch/yr), the highest rate of all the bays measured. Shepard (1953) concluded that the deepening was the result of slow deposition (far removed from a fluvial sediment source) and subsidence.

Recent measurements (1969–1984) indicate that the silt load of the Trinity River (Fig. 21) is only about 14 percent of that reported by Shepard (1953). The U.S. Army Corps of Engineers (USACE, 1942; cited in Rehkemper, 1969a) estimated annual average sediment contributions by the Trinity River to be 7,260 acre-ft (which is higher than Shepard's estimate for the combined load of the Trinity and San Jacinto Rivers); the more recent measurements of Trinity River sediment load (528 acre-ft/yr) are only about 7 percent of this value. Contributions of sediment by the Trinity and San Jacinto Rivers (before 1940) were more than 90 percent of the total sediment contributions to the bay system (USACE, 1942; Rehkemper, 1969a). Rehkemper (1969a) noted that the USACE (1942) estimate for the Trinity River exceeded both Texas Board of Water Engineers and USGS data (by a factor of approximately 2), but concluded that this higher value could be due either to a much shorter sampling period used by USACE, or to reservoir construction on the Trinity River after 1940, in which case the USACE data would be more representative of unaltered sediment loads.

USDA (1959) and Greiner (1982) estimated (based on suspended load measurements at Romayor) that the annual volume of sediment being deposited in Galveston Bay from the Trinity River was about 4,500 acre-ft, of which about 800 to 900 acre-ft, or approximately 20 percent, was estimated to be bed load (unmeasured). Greiner (1982) acknowledged that more recent measurements indicated that this rate of sediment delivery had diminished and that with continued implementation of soil conservation measures and construction of reservoirs up stream, this decline is expected to continue. Recent measurements reflect this decline. Average annual suspended load of the Trinity River at Romayor from 1969 to 1984 was approximately 530 acre-ft, or about 15 percent of the suspended load reported by Greiner (1982). If one assumes that about 20 percent of this amount (530 acre-ft) provides an estimate of bed load transported by the Trinity River to the head of the bay, then the resulting bed-load volume is about 105 acre-ft/yr.

Dredging records (U.S. Army Corps of Engineers, unpublished maintenance dredging records) for Anahuac Channel, along which the Trinity River discharges into Trinity Bay, indicate that the amount of material dredged from the channel during this period (1969–1984) averaged about 90 acre–ft per year, which is a volume close to the 105 acre-ft/yr approximated above. It is acknowledged that the volume of material deposited in Anahuac Channel is not necessarily a reflection of the total bed load carried by the Trinity River for several reasons. Among them are that some bed load is deposited upstream, some is removed by maintenance dredging along the channel to Liberty, and some is transported into Trinity Bay at the mouth of the channel. Nevertheless, a closer examination of the dredging records for Anahuac Channel indicates some trends. For example, dredging records for selected periods show that the annual average volume of sediment removed from the channel during the period 1965 to 1975 exceeds earlier and later periods by a factor of approximately 1.5 (Fig. 92). Comparisons of streamflow and river suspended load with volume of sediments dredged on a year-by-year basis provide some possible reasons for at least some variations in dredged material (Fig. 93).

Contribution of sediment to Trinity Bay by shoreline erosion was estimated as less than 5 percent of the total sediment contribution to the bay (USACE, 1942). Although erosion rate of the bay shoreline may have increased in recent years (Paine and Morton, 1986), which could increase sediment supply from this source, it seems likely that the reduction in stream sediment supply plus subsidence will yield increasingly deeper bay waters, particularly in more rapidly subsiding areas, during future years.

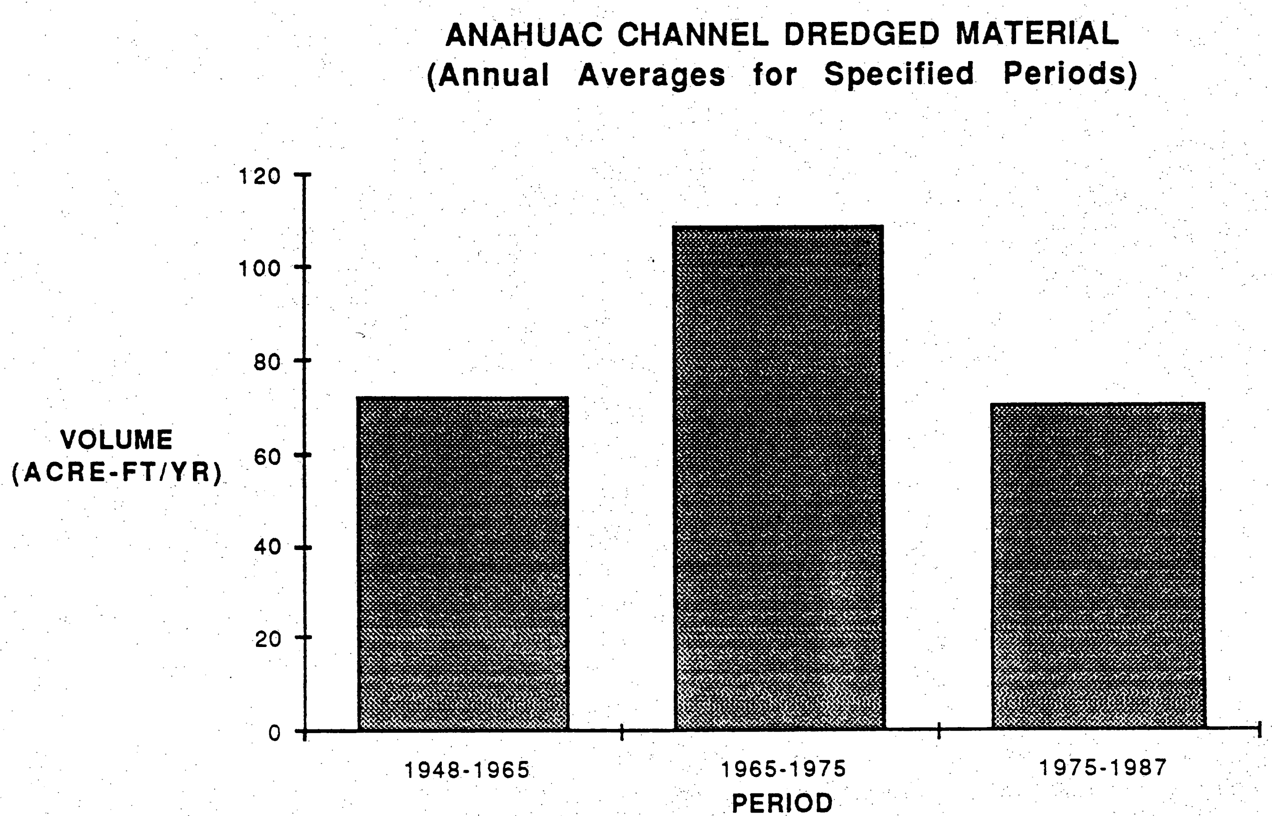


Figure 92. Annual-average volume of sediments dredged from Anahuac Channel along the Trinity River. (Based on unpublished dredging records provided by the U.S. Army Corps of Engineers.)

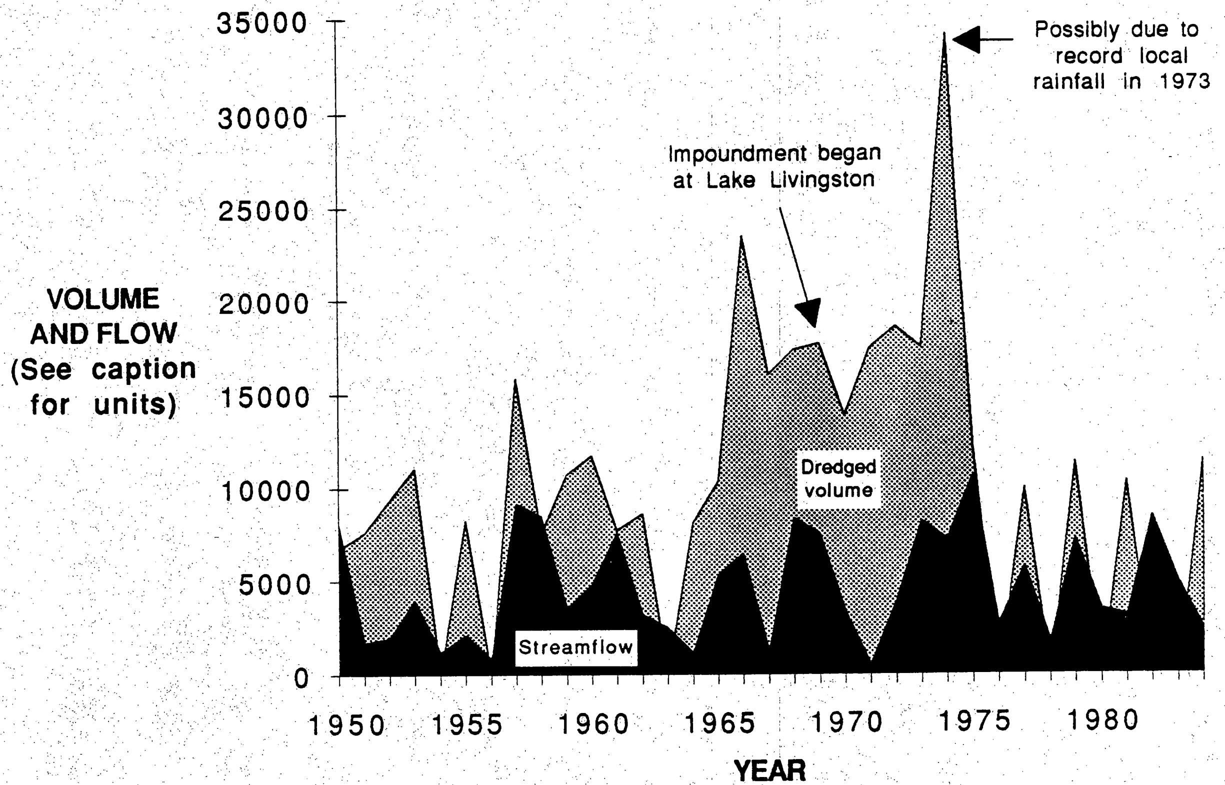


Figure 93. Relationship of sediment volume dredged from Anahuac Channel with annual streamflow along the Trinity River. Streamflow is in acre-ft divided by 1,000, and volume of dredged material is in yd^3 divided by number of months between dredgings. Two events that possibly affected rates of sediment accumulation in the channel are also shown. (Based on unpublished dredging records provided by the U.S. Army Corps of Engineers, and streamflow data from sources listed in figure 21.)

MACROBENTHOS–SEDIMENT RELATIONSHIPS

Introduction

Bottom sediments contain epifaunal and infaunal invertebrates that live on or in the substrate (benthos) and that obtain their food from the water column or from the sediment. Benthic organisms are important in estuarine ecosystems for several reasons: (1) they provide support for primary producers by recycling nutrients, such as nitrogen, in the sediments (Flint and Kalke, 1985). This source of nutrients might be especially important in stabilizing ecosystems largely dominated by nutrient input from river runoff, which may be subject to climatic variations (Rowe and Smith, 1977); (2) they consume bacteria and meiobenthos and serve as both primary and secondary consumers in the detritus based food chain (Armstrong, 1987); (3) they are food for many bottom-feeding fish; (4) they have limited mobility compared to plankton or fishes, and their abundance and diversity have often been studied in order to demonstrate changes in the health or productivity of an estuary; (5) they accumulate trace metals above concentrations found in surrounding waters or sediment (Scrudato and others, 1976); (6) they pellitize fine-grained sediment into agglomerated fecal pellets, which have greater settling rates than their composite particles; and (7) they rework the sediment and influence the transport and fate of sediment (Schaffner and others, 1987).

Biologically mediated sedimentation processes may be as important as the mechanical or physical processes that lead to deposition of fine sediments (Biggs and Howell, 1984). The conceptual model (Armstrong, 1987) in Figure 94 illustrates the role and importance of benthos in cycling nutrients, in organic sedimention, and in organic matter decomposition. The heavier lines show the flow of organic matter to the sediments and the return of nutrients to the water column to be cycled again by phytoplankton. This is the primary flow of organic matter and nutrients in the estuary contrasted to the lesser influence of riverine nutrients and oceanic inwelling (Armstrong, 1987).

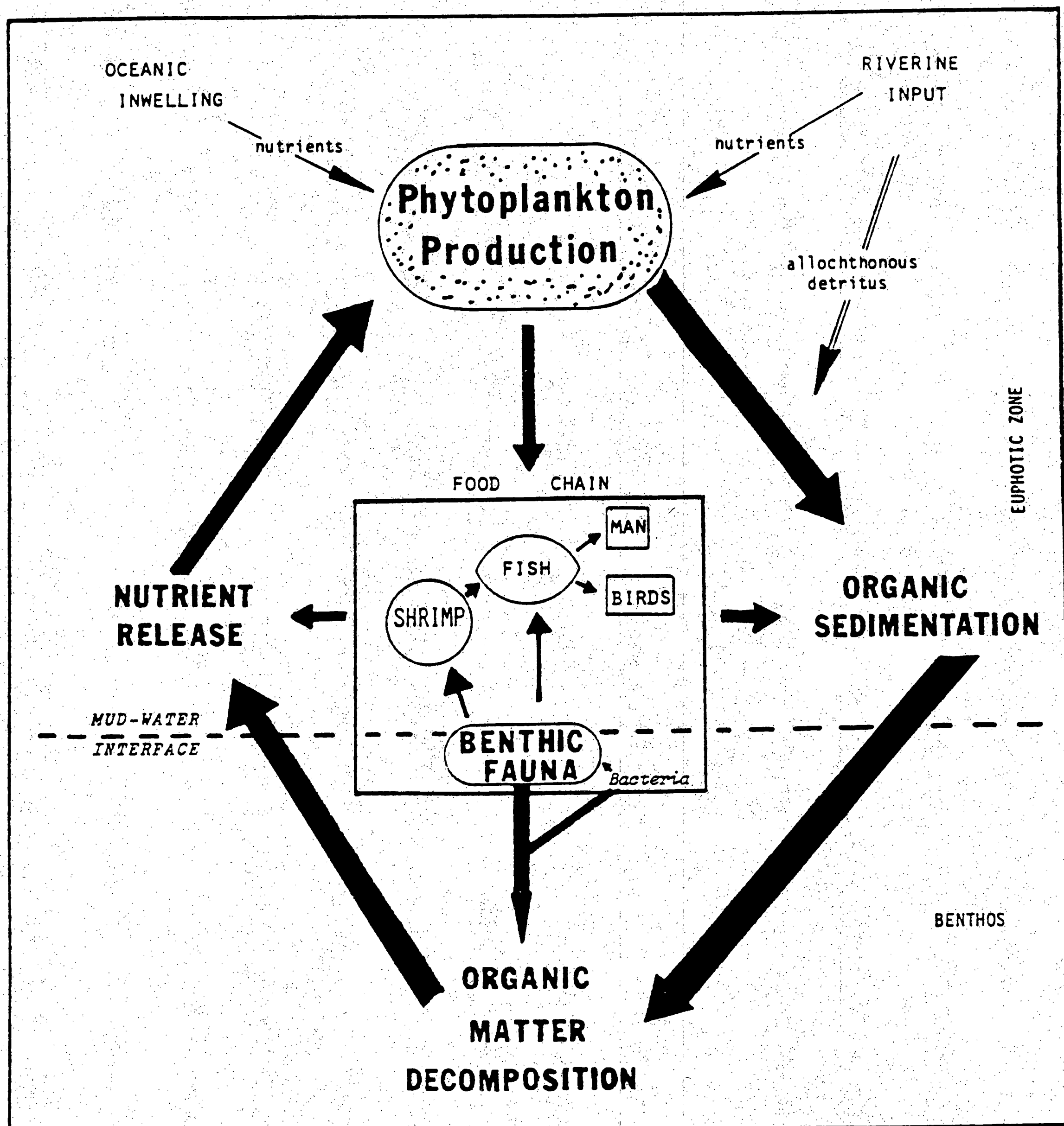


Figure 94. Conceptual model illustrating the role of the estuarine benthos in food chain dynamics and nutrient recycling. (From Armstrong, 1987)

Many ecologic factors affect the distribution, diversity, and biomass of benthic invertebrates in estuaries, including substrate, salinity, temperature, organic content, seagrass distribution, interspecific competition, predation, vagility, and others. Early workers in the field of benthic marine ecology, such as Bader (1954), Thorson (1957), and Sanders (1958), and more recent workers, such as Boesch (1971), Johnson (1971), Bacescu (1972), Gray (1974), Rhoads (1974), Holland and others (1975), Holland and others (1977), Loi and Wilson (1979), and Flint and Kalke (1985), concluded that macrobenthic species are greatly influenced by sedimentary parameters and closely associated ecologic variables, such as organic content, depth, turbidity, or sediment stability. However, simultaneous responses to other environmental factors, such as temperature, salinity, currents, light, and many others, make it difficult to determine the ranges of tolerances of organisms to various sediment parameters. Biological interactions between macrofauna (Commito, 1982) and predation by large, motile predators (Virnstein, 1977) are also important processes controlling community structure. Although this report concentrates on sediment–faunal relationships and their importance in the estuarine ecosystem, the estuarine environment is complex and no single ecologic factor governs the population dynamics of the benthic community.

The close association between substrate and benthos begins with pelagic larvae. Many larval species are restricted to certain types of sediment, and settlement is far from random (Thorson, 1957; Gray, 1974). Larvae of many polychaetes and echinoderms are able to test the substrate, and those ready to metamorphose are able to delay metamorphosis until they find a suitable substrate (Thorson, 1957). Behavioral responses to light, pressure, gravity, salinity, and water currents play a significant part in restricting the range of substrates available to the larvae (Gray, 1974). Physical properties of the substrate that are of great importance to larval settlement are structure and contours of the surface, grain size (Gray, 1974), and sediment stability (Orth, 1977). Chemical and biological factors of the substrate that help determine larval settlement are (1) inorganic or organic compounds, (2) the presence of live microorganisms on the surface of the substrate, (3) the presence of populations of the same species, and (4) other factors (Gray, 1974).

Trophic Structure

Another important aspect in the relation between the benthos and sediment is the manner in which an organism feeds. Most benthic species feed in one or more of the following ways: (1) by consuming plant tissue (herbivores); (2) by feeding on living or recently dead animal tissue (carnivores-scavengers); (3) by feeding on the fluids of living tissues (parasites); (4) by feeding on deposited detritus (deposit feeders); or (5) by feeding on organic particles and inorganic detritus in the water column (suspension feeders) (Rhoads, 1974). The latter two methods are the most common; however, the distinction between these feeding types is not always clear because some species of estuarine organisms have shown flexibility in their mode of feeding. For example, McLusky and Elliott (1981) have shown that the bivalve *Macoma balthica* can spend 10 to 40 percent of its life suspension feeding and 60 to 90 percent of its life deposit feeding.

Organisms that feed exclusively on deposited food would be expected to reach maximium diversity on muds containing an abundant food supply (Rhoads and Young, 1970). Weak currents favor the deposition of silt- and clay-sized particles and prevent rapid removal of organic detritus. Consequently, the proportion of deposit feeders comprising the fauna will increase as organic content of the sediment increases. High organic carbon in the sediment is related to grain size. Particulate organic carbon may be absorbed to fine-grained sediment, particularly to clay-sized particles (McGowen and others, 1979). It is the organic content of the sediment, not silt-clay content, that is causally related to the proportion of deposit feeders in the sediment (Sanders and others, 1962; Purdy, 1964).

The density (numbers of organisms/m^2) of deposit feeders is, in part, controlled by the abundance of microorganisms (Driscoll, 1975; Levinton, 1977). Bioturbation and fecal formation by deposit feeders result in increased surface area for colonization by microorganisms. Driscoll (1975) suggested that a feedback relationship exists, increasing microorganisms resulting in increasing deposit-feeder abundance. The latter, through bioturbation and biodeposition, can produce an environment suitable for more microorganisms. The feedback rate is temperature dependent (Driscoll, 1975).

The optimal mean grain size for suspension feeders is in the fine- to medium-sand range (0.125 to 0.50 mm) (Sanders, 1958; Bloom and others, 1972; Whitlatch, 1977), although many suspension feeders also occur on gravel/shell-sized sediments (>2.0 mm) (Craig and Jones, 1966). Unlike deposit feeders, suspension feeders are unaffected by organic content of the sediment because, by definition, suspension feeders consume organic detritus in the water column, and it is the amount of food in the water that is of primary importance. Other factors being equal, the higher the current velocity, the greater the amount of organic matter brought to the suspension feeders per unit time and consequently the larger the proportion of suspension feeders in the benthos (Purdy, 1964). However, high current velocities can cause considerable substrate mobility or instability and this can reduce the benthic population (Purdy, 1964). Whitlatch (1977) found that coarse sand stations at Barnstable Harbor, Massachussetts, had reduced abundances of both deposit and suspension feeders. These stations also had pronounced surficial ripple marks, evidence of sediment instability.

Sediment instability can also be produced by deposit feeders in fine-grained sediments. Deposit feeders can modify silts and clays by (1) high turnover due to resuspension in tidal currents, (2) high turbidity at the sediment–water interface, and (3) production of textural and compositional grading (Rhoads and Young, 1970). Such instability inhibits suspension feeders and sessile epifauna by clogging filtering mechanisms, resuspending and burying larvae, and discouraging the settlement of larvae of suspension feeders and adults of sessile epifauna. This process, the exclusion of one trophic group in an area occupied by another, is termed trophic-group amensalism (Rhoads and Young, 1970). Conversely, polychaete and crustacean tubes can help stabilize the sediment and increase species diversity (Orth, 1977). Seagrasses are also known for their ability to stabilize sediments by baffling currents and damping wave action (Ginsburg and Lowenstam, 1958; Orth, 1977).

In Texas bays and estuaries, deposit feeders are abundant in the river-influenced assemblage (table 16). This assemblage generally occurs in the upper bays, near areas of fluvial input and where

Table 16. Feeding type of benthic macroinvertebrate species characterisitic of river-influenced, oyster-reef, and grassflat assemblages in the Galveston Bay system.

Location and assemblage	Feeding type
Galveston-Trinity-East Bays	
River-influenced assemblage	
Mulinia lateralis	Suspension feeder
Macoma mitchelli	Deposit feeder
Rangia flexuosa	Suspension feeder
Texadina sphinctostoma	Deposit feeder
Texadina barretti	Deposit feeder
Streblosio benedicti	Surface deposit feeder
Capitella capitata	Surface deposit feeder
Mediomastus californiensis	Subsurface deposit feeder
Polydora ligni	Surface deposit feeder
Corophium louisianum	Suspension feeder
Oyster-reef assemblage	
Boonea impressa	Parasitic
Texadina sphinctostoma	Deposit feeder
Crassostrea virginica	Suspension feeder
Ischadium recurvum	Suspension feeder
Brachidontes exustus	Suspension feeder
Mulinia lateralis	Suspension feeder
Nereis succinea	Surface deposit feeder
Polydora ligni	Surface deposit feeder
Mediomastus californiensis	Subsurface deposit feeder
Streblospio benedicti	Surface deposit feeder
Parandalia fauveli	Carnivore or omnivore?
Melita nitida	Unknown
Rithropanopeus harrisii	Scavenger or omnivore?
Cassidinidea lunifrons	Scavenger or omnivore?
West Bay	
Grassflat assemblage	
Amygdalum papyrium	Suspension feeder
Laevicardium mortoni	Suspension feeder
Chone duneri	Suspension feeder
Nereis succinea	Surface deposit feeder
Streblospio benedicti	Surface deposit feeder
Ampelisca abdita	Deposit feeder
Edotea montosa	Scavenger or omnivore?

sediments are dominantly mud or sandy mud. River input brings large quantities of organic laden muds that provide food for deposit feeders and suspension-feeding species that are able to tolerate high turbidities. However, this assemblage is also subjected to greater natural saline fluctuations than are other bay assemblages and the benthic community living in it is probably highly stressed. Although total benthic standing crop or density may be high due to large numbers of estuarine endemics, diversities are low because most benthic species are not able to tolerate the wide range in salinity.

Suspension feeders, especially bivalves, are most abundant on oyster reefs and in marine grassflats (table 16). These habitats are structurally complex and substrates, food, and shelter are available for most feeding types.

Biomass and Density

The most important part of biomass production in all biotopes (niches) of an estuary is that produced by the primary consumers, which include the benthos (Armstrong, 1987). Benthic organisms are in an intermediate position in the estuarine food chain, serving as primary and secondary consumers and transferring energy to higher trophic levels such as demersal fish and other predatory organisms.

Biomass is defined as the amount of living substance or living weight of the organisms being studied. Biomass is usually expressed as weight per unit area or grams per square meter. It can also be expressed in dry ash-free weight (Crisp, 1984). Other alternatives for measuring biomass include the chemical analysis of tissues for nitrogen and for caloric content (biomass expressed as energy) (Crisp, 1984). The relative merits of numbers or density (numbers of organisms per square meter), biomass, and energy flow are discussed by Odum (1971). Odum (1971) stresses that numbers overemphasize the importance of small organisms, and biomass overemphasizes the importance of large organisms, but energy flow provides a more suitable index for comparing all populations in an ecosystem.

Infaunal biomass measurements have been used to calculate annual benthic production (Nichols, 1977; Flint and others, 1981). Production can be defined as the total amount of tissue in the population under study during a given time period and can rarely be measured directly (Allen, 1971). Its measurement calls for knowledge of the biomass of the population at the beginning and end of the period and of the mass of living components that have been lost by death or emigration during the period (Allen, 1971). Nichols (1977) discusses the use of the ratio of production to biomass (turnover ratio) for making approximate estimates of annual production from biomass measurements. Production is computed by multiplying the mean of four seasonal estimates of biomass for each species by 4.5, a turnover ratio that Nichols (1977) felt is reliable for a first estimate of the productivity of the common species populations. Total production is obtained by summing the products for each of the common species. Others have questioned the accuracy of the turnover ratio in estimating production. Allen (1951) has stated that the simple multiplication of mean biomass by the number of turnovers in a year would lead to an underestimate of production. Allen (1971) emphasized that it is only when growth and mortality figures are known that a mathematical relationship between biomass and production can be predicted.

Flint and others (1981) used regression equations to establish a relatively high correlation between infaunal total density and total biomass in the Corpus Christi–Nueces Bay system. After calculating expected biomass from total density measurements, expected and observed biomass measurements were then compared with fishery harvest figures for Corpus Christi Bay from 1973–1982 (Flint and others, 1981; Flint, 1985). Flint (1985) found a strong correlation (r^2=0.67) between shrimp fishery yields and mean annual benthic biomass (wet weight). Fisheries production was high following a large benthic bloom in 1980. Flint concluded that either the shrimp responded to the same physical/chemical changes as the benthic infauna or the shrimp derived at least part of their nutrition from the infauna and responded to an increased food source from the large benthic bloom in 1980.

Biomass and density measurements of macrobenthic communities vary considerably (tables 17 through 21), and high values often depend on many ecologic factors that affect either large, heavy

Table 17. Biomass and density of benthic macroinvertebrates from bays–estuaries–lagoons on the Texas coast and other coastal areas.

Location	Total* or mean biomass (g/m^2)	Total** or mean density ($organisms/m^2$)	Average percent sand or textural type	Sieve size (mm)	References
Texas bays-estuaries-lagoons					
Nueces Bay	49		49	0.5	Flint and Kalke, 1985
Corpus Christi Bay	86	2,238	2	0.5	Flint and Kalke, 1985; Armstrong, 1987
Corpus Christi Bay	26	1,925	94		Flint and Kalke, 1985; Armstrong, 1987
Upper San Antonio Bay		2,250	36	0.5	Matthews and others, 1975
Mid-San Antonio Bay		1,890	37		Matthews and others, 1975
Lower San Antonio Bay		590	27		Matthews and others, 1975
Other bays-estuaries-lagoons					
Mouth of Rhode River, Maryland (Chesapeake Bay)	119*	34,000**	5	0.5	Hines and Comtois, 1985
Mouth of Rhode River, Maryland (Chesapeake Bay)	177*	22,000**	73		Hines and Comtois, 1985
Gulf of Maine	3.18		silt and clayey silt	1.0	Larsen, 1979
Cape Cod Bay, Mass.	1.6-177.9*	11,190-30,150	81-96	1.0	Young and Rhoads, 1971
Cape Cod Bay, Mass.	10.3-16.7*	7,870-14,230	6-33		Young and Rhoads, 1971
Buzzards Bay, Mass.		1,629-12,576**	87-99	0.5	Sanders, 1958
Buzzards Bay, Mass.		1,187-7,982	7-56		Sanders, 1958
Hanno Bay, Baltic	38-90	5,410-6,169	fine sand	1.0	Persson, 1983
Puget Sound, Wash.	16.5-18.8	389-1,122	mud	?	Lie and Evans, 1973
Puget Sound, Wash.	2.9-13.7	544-3,427	silty sand-sand		Lie and Evans, 1973

*Total biomass

**Total density

Table 18. Mean density of benthic macroinvertebrates in marine grasses.

Location	Mean density (organisms/m^2)	Sieve size (mm)	Sources
Texas bays–estuaries–lagoons			
South Bay	532	1.0	White and others (1986)
Lower Laguna Madre	1,911	1.0	White and others (1986)
Baffin Bay	3,354	1.0	White and others (1989)
Upper Laguna Madre	1,601	1.0	White and others (1983)
Redfish Bay	793	1.0	White and others (1983)
Espiritu Santo Bay	5,726	1.0	White and others (1989)
Espiritu Santo Bay (seasonal)	9,153	1.0	
West Bay	4,167	1.0	White and others (1985)
Other-bays–estuaries–lagoons			
Indian River, FL	17,479	0.5	Virnstein and others (1983)
Chesapeake Bay, VA	48,900	1.0	Orth (1977)
Apalachicola Bay, FL	38,780	0.5	Sheridan and Livingston (1983)
Pensacola Bay, FL	6,077	0.5	Stoner and others (1983)
Beaufort, NC	672	6.0	Williams and Thomas (1967)

Table 19. Mean density of benthic macroinvertebrates versus mean percent sand in Espiritu Santo, Matagorda, Galveston, and Trinity Bays.

Location	Mean density (organisms/m2)	Mean percent sand per station	Number of stations
Espiritu Santo Bay	86	2	1
	589	35	8
	402	70	15
	532	90	7
Matagorda Bay	744	8	17
	833	33	10
	877	61	3
	1,598	90	15
Galveston Bay	216	9	6
	211	39	3
	333	65	10
	287	91	7
Trinity Bay	111	6	7
	65	34	6
	86	68	1
	53	89	3

Table 20. Mean density versus mean percent sand for benthic macroinvertebrate assemblages in Texas bays.

Assemblage	Mean density (organisms/m^2)	Mean percent sand per station
Galveston/Trinity/East Bays		
Oyster reef	604	44.1
River influenced	197	35.7
Open bay center	168	35.1
Bay margin	260	77.6
Inlet influenced	339	59.2
Matagorda Bay		
Open bay center	861	21.7
Inlet influenced	1,969	87.9
Oyster reef	776	60.0
Bay margin	993	92.7
Lavaca Bay		
River influenced	670	54.0
Open bay center	1,538	23.4
Oyster reef	5,417	94.8
Corpus Christi Bay		
Inlet influenced	585	61.1
Bay margin	703	53.6
Open bay center depauperate	111	10.0
Open bay center	285	6.5
Oyster reef	1,298	----
River influenced	388	70.1
San Antonio Bay		
River influenced	2,702	34.3
Oyster reef	2,124	95.1
Enclosed bay center	45	39.2
Bay nargin	316	96.8

Table 21. Biomass and density of benthic macroinvertebrates in marshes, mudflats, and sandflats.

Location	Mean biomass (g/m^2)	Mean density (numbers/m^2)	Sediment type (percent sand)	Marsh vegetation or habitat	Remarks (sieve size in mm)	References
West Bay. Texas		3-4.000	sand and muddy sand	Spartina alterniflora	total fauna (0.4)	Gilmore and Trent. 1974
Newport River. North Carolina	3	7,600	sand	Spartina alterniflora	macroinfauna (0.6)	Cammen. 1979
North Florida	123	475	sand	Juncus roemerianus	total fauna (1.0)	Subrahmanyam and others. 1976
St Louis Bay. Miss.		316	sandy mud (37)	Juncus roemerianus	mollusks only (1.0)	Bishop. 1981
St. Louis Bay Miss.		396	sandy mud (44)	Spartina cynosuroides	mollusks only (1.0)	Bishop. 1981
San Francisco Bay. Calif	13-24	53.000-155.000	mud (14-24)	mudflat	macroinfauna (1.0)	Nichols. 1977a
Lynher estuary England	13.2*		mud (1-3)	mudflat	total fauna (0.5)	Warwick and Price. 1975
Great Bay New Hampshire	17 8	476	fine sand	sandflat	macroinfauna	Winston and Anderson. 1970
Great Bay New Hampshire	5 2-27 9	135-987	silt and silty clay	mudflat	macroinfauna	Winston and Anderson. 1970
Grevelingen Netherlands	0 3-33.9		very find sand with some mud	sandflat	total fauna (1.0)	Wolff and deWolf. 1977

*Total benthic biomass

organisms (e.g., bivalves and large polychaetes) in the case of biomass values or small, numerous organisms (e.g., small polychaetes) in the case of density values. Sediment type and the presence of seagrasses (table 18) and marsh vegetation (table 21) are just a few of the many important ecologic variables affecting benthic biomass and density measurements.

High biomass and density measurements have been reported for sand, mud, shell/gravel, or mixed (sandy mud or muddy sand) sediments. Studies finding highest biomass and/or density values in sandy sediments include Sanders (1958) in Buzzards Bay, Massachussetts (table 17), Young and Rhoads (1971) in Cape Cod Bay (table 16), Maurer and others (1978) in Delaware Bay, and Hines and Comtois (1985) in central Chesapeake Bay (table 17). Highest values in muds are cited in Lie and Evans (1973) in Puget Sound (table 17) and Flint and others (1981, 1982) and Flint and Kalke (1985) in Corpus Christi and Nueces Bays. Studies with highest values in mixed sediments or shell/gravel include Gillard (1974) in upper Galveston Bay; Holland and others (1975) in Corpus Christi, Nueces, Aransas, and Copano Bays; Matthews and others (1974,1975) in San Antonio Bay (table 17); Parker (1975) in Cape Cod Bay and Texas bays; and Reinharz and O'Connell (1983) in upper and central Chesapeake Bay. This is merely a partial list of studies that include discussions of biomass/density and sediment relationships, but it illustrates differences between various bay-estuarine systems. Comparison of the results from these diverse areas is confounded not only by the complexity of the environment in the different bay-estuarine systems but also by the use of different mesh-size sieves (tables 17 through 19) and different sampling periods. Although both larger and smaller mesh sizes are used, 0.5 and 1.0 mm sizes are probably the most common sizes used in benthic studies. Sheridan and Livingston (1983) discuss this problem and note that other studies have found 2 to 6 times the number of individuals in the smaller mesh sieve of 0.5 mm versus the larger 1.0 mm size.

The only studies in Texas estuaries correlating benthic infaunal biomass with sediment were those of Flint and others (1981, 1982) and Flint and Kalke (1985). Armstrong (1987) reported on the results of their studies (Flint and others, 1981, 1982; Flint and Kalke, 1985) of benthic infaunal production in Corpus Christi and Nueces Bays and on the inner shelf near Mustang Island from 1981 to 1983. Standing stock biomass was greatest at the muddy, midbay site (station 7) in Corpus Christi Bay and least at the sandy, inner-shelf site (Fig. 95 and table 17). Sediment at the midbay site was more than 70 percent clay. There was a great deal of variation in standing stock biomass at the midbay site, with biomass ranging from less than 4 g/m^2 after the first year of study to more than 100 g/m^2 after the second year. Biomass at the mixed-sediment site in Nueces Bay (station 2) consistently showed minimum standing stocks in the summer and fall (0.4–20.1 g/m^2) and maximum biomass in the winter and spring (49.4–151.8 g/m^2). Biomass at the sandy, inner-shelf site (station 10) was much less variable than biomass at the bay sites, probably because of the stabilizing influence of oceanic waters (Armstrong, 1987).

Densities of organisms from four different bays on the Texas coast illustrate the variation in density with mean percent sand (table 19). In Matagorda Bay, densities were highest in sandy (90 percent sand) sediments, whereas in Trinity Bay, densities were highest in muds (6 percent sand). Densities in Galveston Bay and Espiritu Santo Bay were highest in mixed sediments of sandy mud or muddy sand.

Benthic Assemblages

High biomass and density values are dependent on many variables and not restricted to a particular sediment type; however, samples taken from oyster reefs and seagrasses generally always have high density values (tables 18 and 20). Seagrasses provide food, diverse habitat, and protection from predators, and it is not surprising that many studies (O'Gower and Wacasey, 1967; Santos and Simon, 1974; Orth, 1977; Virnstein and others, 1983) have reported higher densities of macrobenthos in seagrass than in adjacent bare sand. Mean densities of macrobenthos from seagrass beds along the Texas coast range from 532 organisms/m^2 in South Bay to 9,153 in Espiritu Santo Bay (table 18). Oyster reefs provide a heterogeneous substrate and suitable surfaces for sessile epifauna to attach. Suspension feeders are abundant on oyster reefs or other shell/gravel substrates. Mean densities of organisms on reefs in various Texas bays range from 604 to 5,417/m^2 (table 20).

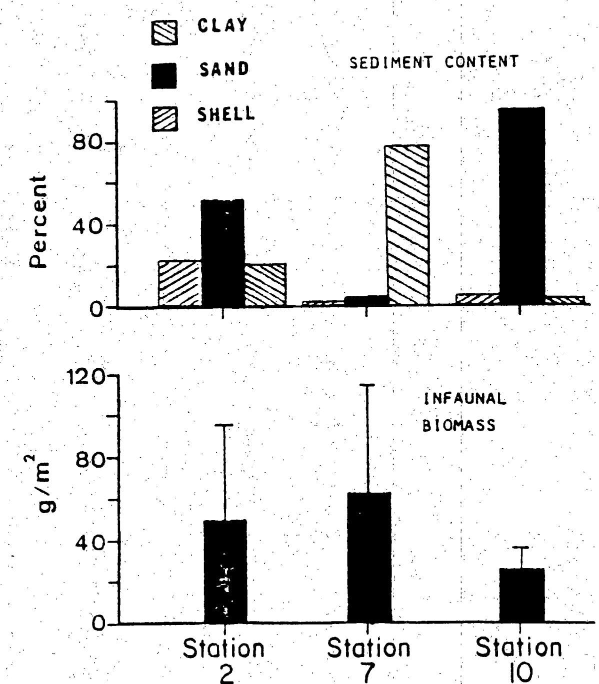


Figure 95. Average measures for sediment characteristics and benthic macroinfaunal biomass at three stations in Corpus Christi Bay from 1981 to 1983. Bars represent percent confidence intervals around the means. (Modified from Armstrong, 1987.)

Densities are also relatively high in areas with an inlet-influenced assemblage (table 20; also see the macroinvertebrate assemblage section in White and others, 1983, 1986). Sediments in these areas range from 59 to 88 percent sand (muddy sand to sand), and macroinvertebrate densities range from 339 to 1,969 organisms/m^2. An inlet-influenced assemblage occurs in bay–estuary–lagoon systems near tidal inlets, and species composition is typical of both bay and nearshore-shelf areas. Densities and diversities in areas with an inlet-influenced assemblage are probably relatively high because of the stability of the environment—salinities, except in extreme cases of high, localized rainfalls, are probably maintained near oceanic levels.

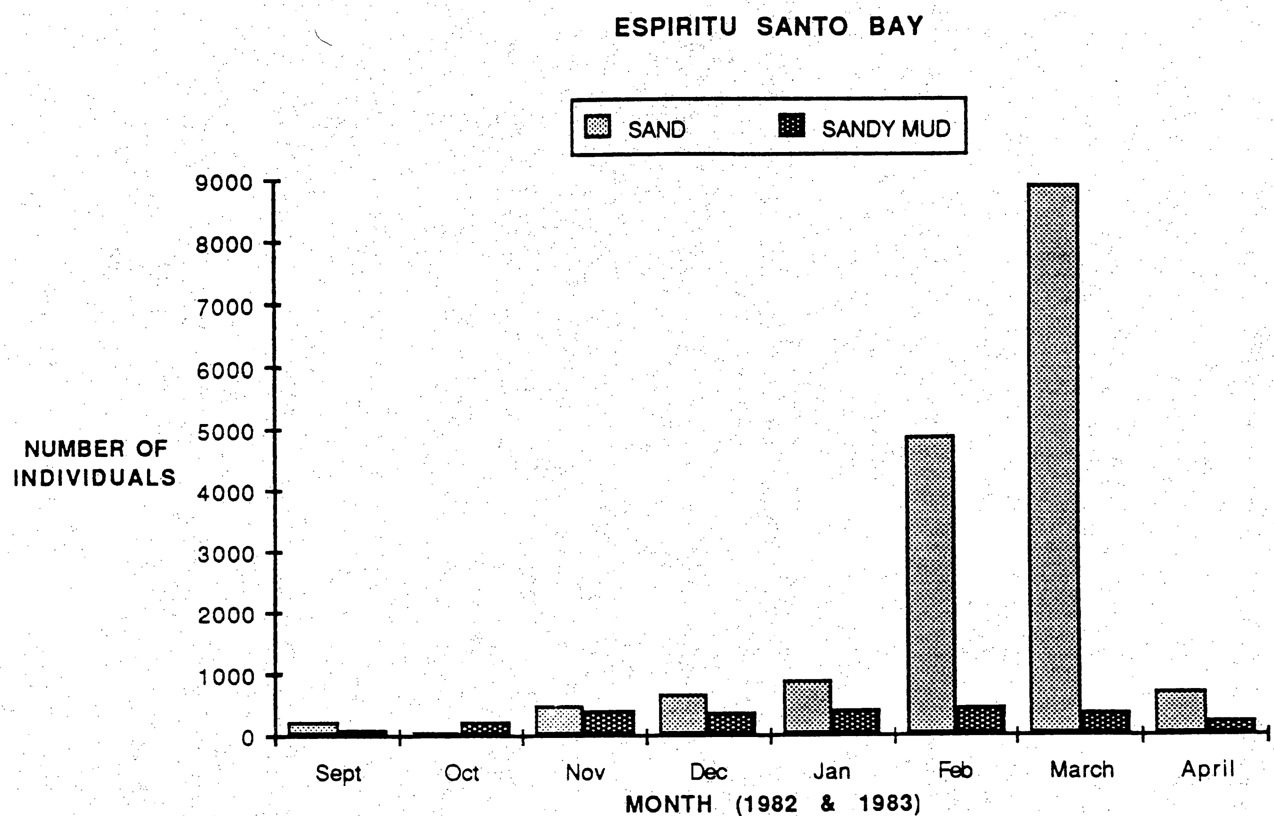
Densities in areas with a river-influenced assemblage are highly variable, ranging from 197 organisms/m^2 in the Galveston–Trinity–East Bay area to 2,702 in San Antonio Bay (table 20). River-influenced assemblages occur in the upper bays and tidally influenced parts of rivers. Organisms in these areas are subjected to natural stress primarily from salinity fluctuations (Bechtal and Copeland, 1970; Holland and others, 1973). Sediment type ranges from 34 to 70 percent sand (sandy mud to muddy sand). Densities in upper San Antonio, Hynes, and Guadalupe Bays are especially high because of large numbers of mollusks, primarily the brackish-water gastropod *Texadina sphinctostoma* and the bivalves *Rangia cuneata*, *R. flexuosa*, and *Mulinia lateralis*. Matthews and others (1975) and Harper and Hopkins (1976) also found high populations of brackish-water species, including *T. sphinctostoma*, in upper San Antonio Bay. In their seasonal study, Harper and Hopkins (1976) noted that benthic populations were highly variable and that some species in the upper bay increased dramatically after a spring flood. They concluded that the bloom was probably in response to increased nutrients brought in by the Guadalupe and San Antonio Rivers and only indirectly related to decreased salinities.

Macrobenthic communities in estuaries are dominated by populations that show large seasonal and year-to-year fluctuations in abundance (Holland, 1985). Many species respond to salinity changes, especially those of an extreme nature, with large population increases (Harper and Hopkins, 1976; Flint and others, 1981) or decreases (Stone and Reish, 1965; Johnson, 1980). Another source of variation for most species is that associated with seasonal recruitment cycles as illustrated by a seasonal study of macrobenthic populations in Espiritu Santo and Lavaca Bays (Fig. 96). The large increase in density during March at the sandy (89 percent sand) station in Espiritu Santo Bay was primarily due to an increase in numbers of the suspension-feeding bivalve, *Lyonsia hyalina floridana*. This may be the time of year for the annual recruitment pulse for *Lyonsia*, as Flint and others (1981) also reported large numbers of *Lyonsia* in Corpus Christi Bay during February and March. Benthic populations increased in the winter and early spring at both stations in Lavaca Bay (Fig. 96), although the fluctuation was not as large as at the sandy station in Espiritu Santo Bay. Peaks in benthic populations in the estuaries on the Texas coast generally occur in the winter and early spring (Harper and Hopkins, 1976). Benthic populations at the sandy-mud (29 percent sand) station in Espiritu Santo Bay were small and fluctuated very little during the 8 months of study (Fig. 96).

Changes in sediment parameters, even on a small scale, can result in population fluctuations on a seasonal basis (Holland, 1985). Holland (1985) found that consistent but small-scale changes in silt-clay content of stations in a muddy-sand habitat accounted for seasonal variations in populations within that habitat. Other environmental factors, such as salinity, dissolved oxygen, and temperature, did not vary within the muddy-sand habitat over ranges sufficient to influence macrobenthic abundance.

Marshes

There is an extensive literature on salt-marsh animal distributions, life histories, and ecology (Daiber, 1982), and even many studies of faunal groups and their relationship to sediment in the marsh. For example, Whiting and Mashiri (1974), Daiber (1982), Barnwell and Thurman (1984), and Thurman (1984) have studied the close correlation between the distribution of the fiddler crabs, *Uca* and *Sesarma*, and sediment type and organic content of the sediment. However, relatively few studies have examined the relationship between sediment and biomass or density of the total benthic community in the marsh.



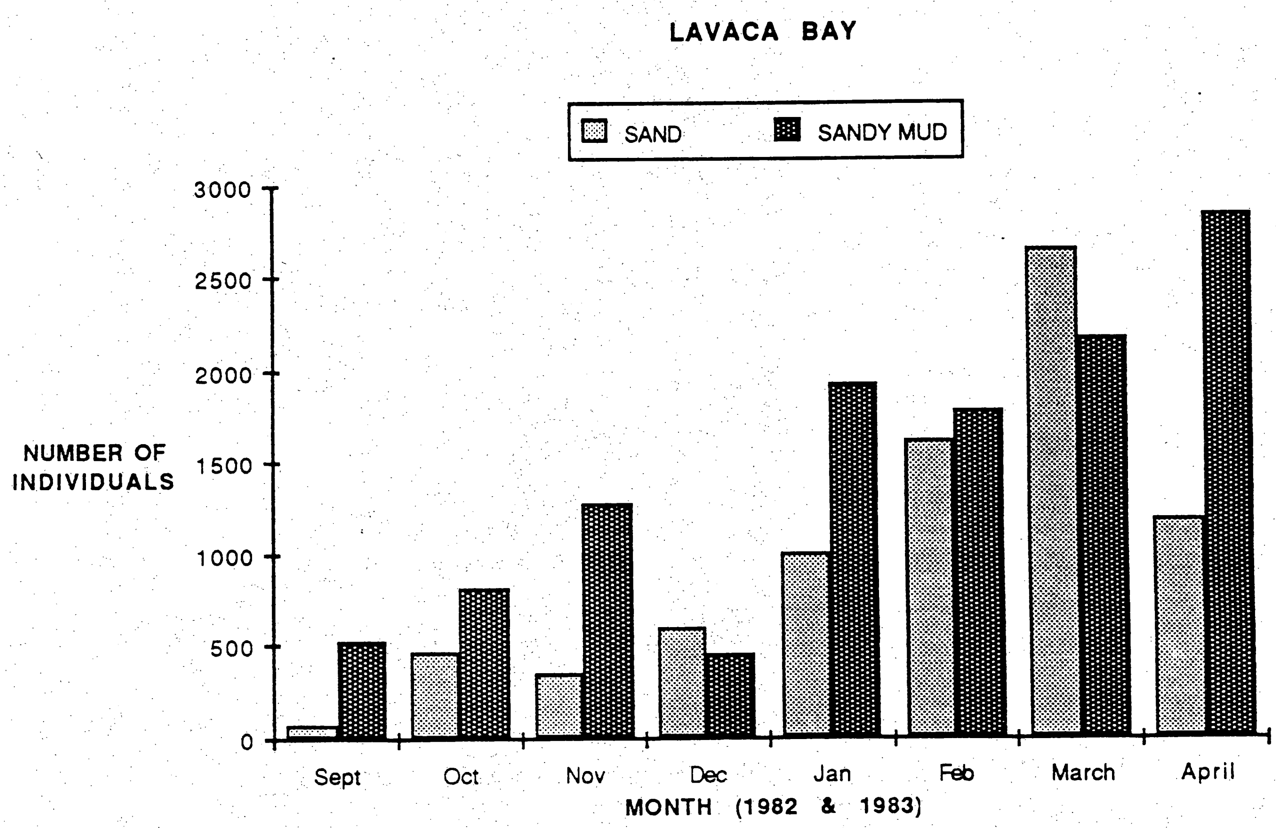


Figure 96. Density (number of individuals/m^2) by month of benthic macroinvertebrates at sandy and sandy mud stations in Espiritu Santo and Lavaca Bays.

Gilmore and Trent (1974) compared the abundance of benthic macroinvertebrates in West Bay, Texas, between a natural marsh, an adjacent marsh altered by channelization, bulkheading and filling, and an open-bay area. Relating abundance to substrate in the three areas, Gilmore and Trent (1974) found that densities for all organisms combined were highest in the marsh (table 21) and least abundant in the open bay. Crustaceans showed a preference for sandy substrates in the marsh, whereas polychaetes were most abundant in canals and at stations with low to intermediate amounts of silt and clay.

Cammen (1979) took monthly benthic samples from a *Spartina* marsh in North Carolina. Infaunal abundance was greatest in late winter and early spring and least in summer and early autumn. Numbers of individuals ranged from 2,200 to 15,500/m^2 (mean density of 7,600) and biomass ranged from 1.3 to 6.1 grams ash-free dry weight (AFDW)/m^2 (mean biomass of 3 g AFDW). Cammen (1979) estimated that annual production of the infauna was 5.9 g AFDW/m^2. Production for the polychaete *Nereis succinea*, which accounted for most of the biomass, was estimated to be 4.1 g AFDW/m^2. Sediment at the marsh sites was mainly medium to fine sand.

Subrahmanyam and others (1976) studied the infauna and epifauna in a *Juncus* marsh in north Florida. The gastropod *Littorina irrorata* accounted for 81 percent of a total biomass of 123 grams/m^2 (table 21). Peaks of invertebrate abundance occurred in the winter and fall. Subrahmanyam and others (1976) reported that organism densities and diversities are generally higher in the more flooded or low marsh zones than in the mid-marsh or high marsh zones, because estuarine species tend to invade lower marshes more easily and there is a greater availability of organic detritus in the low marsh. Subrahmanyam and others (1976) compared marsh densities of their study and other studies with densities of macroinvertebrates from estuaries, and, contrary to the results of Gilmore and Trent, they found that marsh densities appeared to be lower. They speculated that this was because marsh sediments are covered with plant rhizomes and living space is reduced (Subrahmanyam and others, 1976).

Mudflats and Sandflats

The mudflat invertebrate community is an important link in the cycling of organic matter to the estuary. Several investigators of mudflat environments (Boyden and Little, 1973; Nichols, 1977a) have proposed high productivity for this habitat despite small numbers of species.

Nichols (1977a) found that infaunal densities and biomass were very high in the mudflats of San Francisco Bay (table 21). Densities ranged from 53,000 to 155,000 organisms/m^2, depending on the station; however, actual densities should be much higher as Nichols (1977a) speculated he was losing more than one-half of all the individuals of the bivalve *Gemma gemma* through the 0.5-mm-mesh sieve. Total biomass varied from 13 to 24 g AFDW/m^2 without large seasonal variations. Three species of bivalves made up most of the biomass. Nichols (1977a) observed that the large densities of infaunal species indicate the mudflats act as sinks for organic matter produced on the marsh, in the water column, and on the mud itself. The high secondary productivity supported a large shorebird community.

Winston and Anderson (1970) studied the amount of bioturbation in sediments at a sandflat station and at five mudflat stations in Great Bay, New Hampshire. Densities and biomass measurements at the sandflat station were about halfway between the high and low measurements for the mudflat stations (table 21).

Wolff and de Wolf (1977) took monthly biomass measurements at five stations in the sandflats of the Grevelingen estuary, the Netherlands. The highest biomass values were found low in the intertidal zone, with a gradual decrease up to the high water mark. Below the low water mark, biomass was generally low. Mollusks and polychaetes, particularly *Arenicola marina*, comprised most of the biomass at each station.

Species Diversity and Species Richness

The oldest and most fundamental concept of diversity is species richness, or the number of species in a community. Direct species counts provide one of the simplest, most practical, and most objective measures of species richness (Peet, 1974). However, direct counts do not provide enough information about the underlying community, as they do not show how individuals are distributed among the constituent species. As a result, various diversity indices have been devised that are influenced by both species richness and species dominance or how evenly the species are distributed. Probably the most widely used index is the Shannon index or Shannon-Weaver diversity index (H') (see White and others, 1983, for formula). The Shannon index will increase with the number of species and as the proportions of individuals among the species become more equal. Diversity index interpretations should be made cautiously (McIntosh, 1967), because it is very easy to read into the numbers meanings that are not there. This problem is inherent in the Shannon-Weaver formula, because it is affected by both species number and species dominance. Therefore, a single diversity number may be misleading.

Factors potentially increasing species diversity may be summarized as follows: (1) With time, all communities tend to increase in diversity. (2) With increased structural complexity, diversity may be expected to increase. (3) With a predictable environment having a constancy of climate, diversity will increase. (4) With increased competition, diversity will increase. (4) With increased number of predators, competition among prey species will be reduced and diversity will be increased (Gray, 1974). Of these factors, spatial heterogeneity or the structural complexity of the environment is the only one that relates directly to the influence of sediment on community structure (Gray, 1974). Coarse and heterogeneous sediments are more structurally complex and have higher diversities than fine and homogeneous sediments (Gray, 1974); therefore, gravel/shell and muddy sands or sandy muds are generally more diverse than muds or sands.

Sanders (1968) combined the predictable environment, time, and competition hypotheses to form the stability-time hypothesis. This hypothesis states that physical instability in an environment prevents the establishment of diverse communities, and conversely, in a stable environment, with time, speciation and immigration will lead to high diversity. Physical changes such as rapid and severe changes in salinity and temperature, erosion, and rapid deposition may produce a physically controlled community that is characterized by low diversity (Johnson, 1970). Ecological systems in estuaries that are subject to high levels of natural stress include high-energy beaches that are stressed from breaking waves and deltas that receive high rates of sedimentation (Copeland, 1970). Wohlschlag and Copeland (1970) emphasize that estuaries are fragile and that even slight stresses on estuarine systems over long periods of time tend gradually to reduce species diversity, even though yields of common species to exploitation tend to be maintained at reasonable levels while the stresses themselves may tend to remain unrecognized. Shallow bays and estuaries on the Texas coast experience large and unpredictable changes in the environment, and environmental instability may override structural complexity as the major factor controlling diversity.

River-influenced areas are especially susceptible to drastic and sometimes long-term changes in salinity, and benthic assemblages of these areas are characterized by having low diversities and numbers of species (table 22). Sediments are fairly homogeneous and mostly muds except for sands or muddy sands near bay margins. River-influenced assemblages may be found in all or most parts of a bay, such as upper Galveston and Trinity Bays (White and others, 1985), or they may be restricted to creeks or rivers, such as the Colorado River near Matagorda (White and others, 1988). Benthic species in these oligohaline to mesohaline zones (salinities of 0 to 15 ppt) are generally estuarine endemics, such as *Rangia cuneata*, or euryhaline opportunists, such as *Streblospio benedicti* (Boesch, 1977; Schaffner and others, 1987).

Bay habitats where structural complexity is important and diversities and species richness are generally very high are oyster reefs and other shelly sediments and marine grassflats (tables 22 and 23). These environments provide surfaces for the attachment of epifauna and many potential niches for infauna. Oyster reefs and marine grassflats may provide many more refuges for prey species than might structurally simple localities, and more refuges should result in higher diversity of prey

Table 22. Mean percent sand, species diversity, and mean number of species of benthic macroinvertebrate assemblages.

Assemblage	Mean percent sand per station	Range in diversity (H')	Mean number of species
Galveston/Trinity/ East Bays			
Oyster reef	44.1	1.5-2.23	10.3
River influenced	35.7	0.00-2.15	3.5
Open bay center	35.1	0.00-2.00	5.4
Bay margin	77.6	0.56-2.49	6.6
Inlet influenced	59.2	0.58-2.56	9.9
Matagorda Bay			
Open bay center	21.7	0.00-2.88	14.4
Inlet influenced	87.9	1.09-2.85	18.9
Oyster reef	60.0	2.87	34.0
Bay margin	92.7	0.00-2.67	13.6
Lavaca Bay			
River influenced	54.0	0.00-2.37	7.8
Open bay center	23.4	0.56-2.67	14.0
Oyster reef	94.8	2.79	42.0
Corpus Christi Bay			
Inlet influenced	61.1	0.64-2.87	15.3
Bay margin	53.6	0.00-2.17	7.4
Open bay center depauperate	10.0	0.00-1.89	3.8
Open bay center	6.5	0.89-2.46	8.0
Oyster reef	------	0.98-2.33	10.0
River influenced	70.1	1.08	4.0
San Antonio Bay			
River influenced	34.3	0.00-1.72	5.1
Oyster reef	95.1	1.05-1.86	8.4
Enclosed bay center	39.2	0.00-1.55	1.8
Bay margin	96.8	0.00-1.77	4.6

Table 23. Mean percent sand, species diversity, and mean number of benthic macroinvertebrates in marine grassflat assemblages.

Location	Mean percent sand	Range in diversity (H')	Mean number of species
Lower Laguna Madre	69	0.56-3.06	22.0
Upper Laguna Madre	79	0.00-2.70	14.8
Redfish Bay	50	0.00-2.38	12.2
Corpus Christi Bay	95	1.97-2.65	13.5
Espiritu Santo Bay	70	1.07-2.44	28.0
East Matagorda Bay	89	2.03-2.31	13.5
Christmas Bay	30	2.39	24.0

(Menge and Sutherland, 1976). Also, these habitats might decrease the foraging efficiency of the predator (Menge and Sutherland, 1976). Craig and Bright (1986) found high populations of the bivalve *Mercenaria mercenaria texana* in shelly sediments in Christmas Bay, and they speculated that shell fragments may provide young bivalves protection from predators, especially blue crabs, and a favorable substrate to which the bivalve larvae can attach in the byssal stage.

White and others (1985; 1989) found that stations in the Galveston and Matagorda Bay systems containing more than 10 percent shell (gravel) generally had more species (Figs. 97 and 98). Of the eight stations in the Galveston Bay system having more than 10 percent shell, six occur in lower Galveston Bay, one in West Bay, and one in East Bay. Of the five stations having more than 10 percent shell in the Matagorda Bay system, three occur in Matagorda Bay and two in Carancahua Bay. Total numbers of species in Carancahua Bay are highest at the shelly stations. Of 10 stations sampled monthly for 8 months in Espiritu Santo Bay, species diversity (H') was highest at the only station with sediment containing a high percentage of shell (38 percent shell). Holland and others (1975) also reported high numbers of species at stations with shelly sediments in Corpus Christi Bay.

Diversities (H') of benthic macroinvertebrates in marine grassflat assemblages range from 0.00 in upper Laguna Madre and Redfish Bay to 3.06 in lower Laguna Madre (table 23). Most diversities (H') at grassflat stations, especially in lower Laguna Madre, are high (H' above 2.0). Species richness (mean number of species per station) is also high, ranging from 12.2 in Redfish Bay to 28.0 in Espiritu Santo Bay (table 23). Except for the oyster-reef and grassflat assemblages, the mean number of species per station for most other bay assemblages is less than 10.0 (tables 22 and 23).

Other investigators (O'Gower and Wacasey, 1967; Heck and Wetstone, 1977; Virnstein and others, 1983; Lewis and Stoner, 1983; Orth and others, 1984) have also found highly diverse faunal communities in seagrass beds, particularly in comparison with nonvegetated areas. Orth and others (1984) suggested that the abundance of many species, both epifauna and infauna, was positively correlated with two distinct aspects of plant morphology, (1) the root-rhizome mat, and (2) the plant canopy. Heck and Wetstone (1977) found that invertebrate species number was not

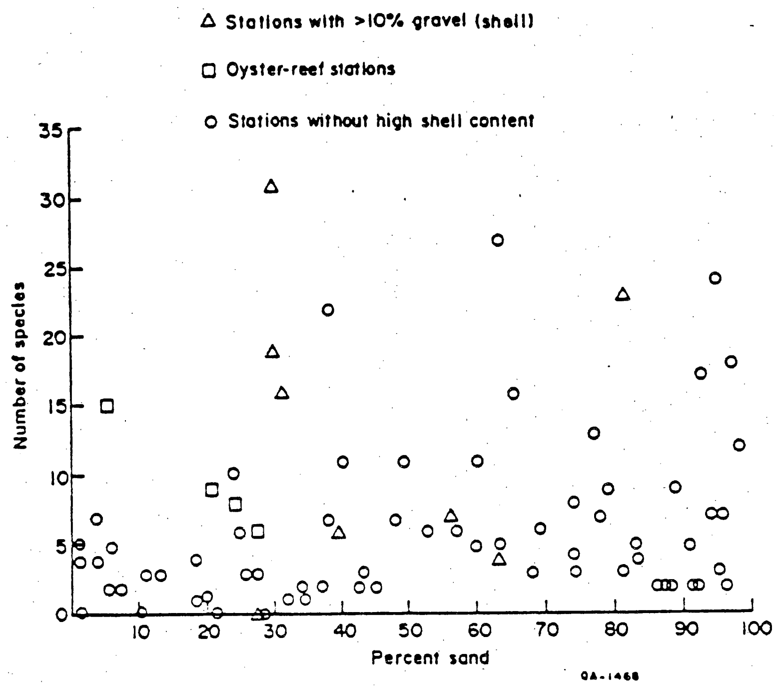


Figure 97. Scattergram of total species and percent sand in the bays of the Galveston–Houston area. (From White and others, 1985.)

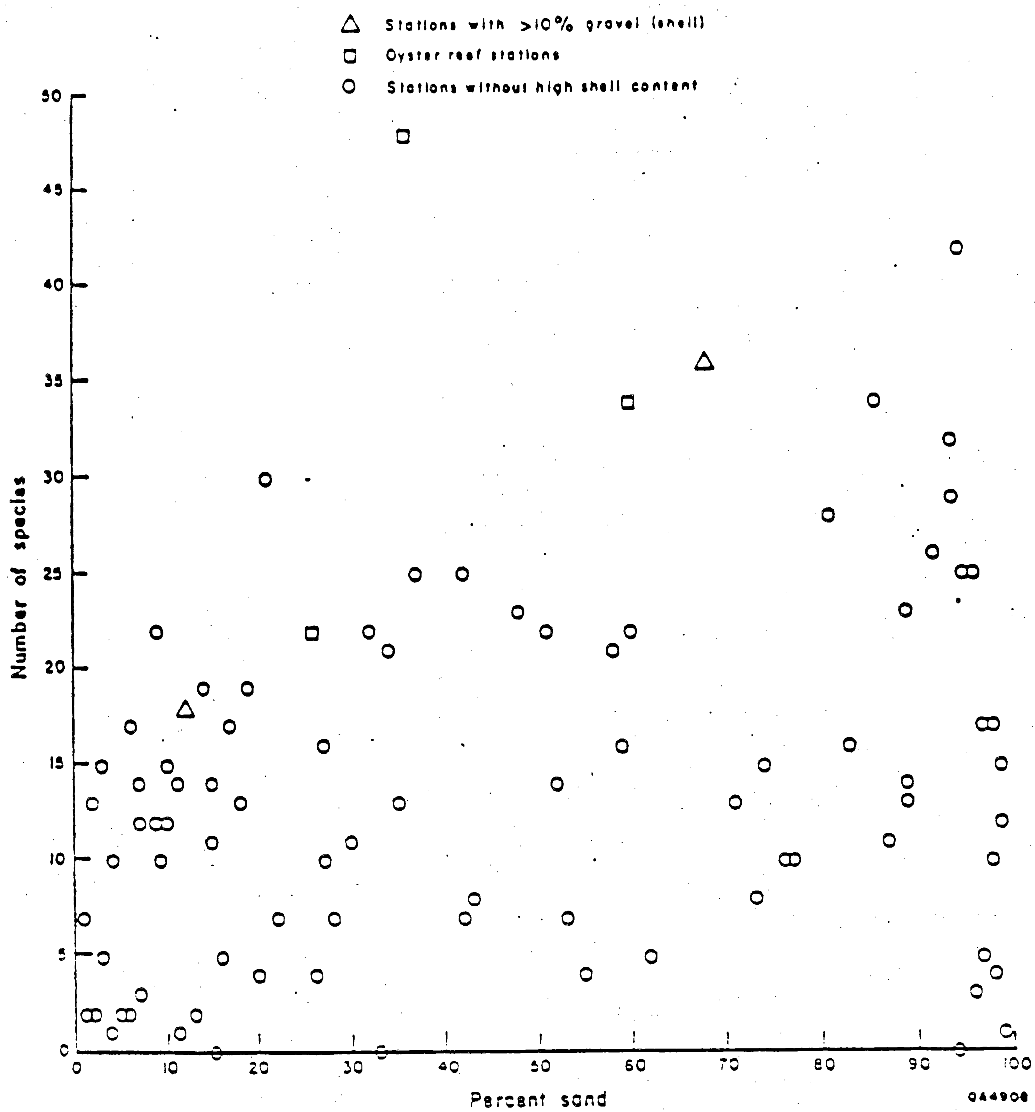


Figure 98. Scattergrams of total species and percent sand in the bays of the Port Lavaca area. (From White and others, 1989.)

significantly correlated with plant species number, but it was significantly correlated with plant biomass.

Other studies that show physical instability affecting species diversity in the bays include Woodward-Clyde Consultants (1977) and Holland and others (1973). Woodward-Clyde Consultants (1977) found that the benthic community was depauperate in species where the physical instability of the substrate was the greatest. For example, there were relatively few species on the shallow, western shoreline of Matagorda Bay, where wave action disturbs the sandy sediment. Also, intensive shrimp trawling activity in the Matagorda Ship Channel created an unstable substrate and was responsible for low numbers of species and individuals at the channel station. On the other hand, physical disturbance was relatively low in the deep, level bottom habitat in Matagorda Bay. Substrates were predominantly silt and shell, and numbers of species were relatively high at all stations.

Holland and others (1973) studied the structure of the benthic community in the Galveston Bay system to ascertain water quality. They applied various diversity indices, including the Shannon index (H′), to data collected during four sampling periods in 1971 to 1972 at five stations in the Galveston Bay system. Two stations are located in the upper bay, one near the Houston Ship Channel–Clear Lake region and one in Trinity Bay. The three lower bay stations are in the middle of West Bay, near the Texas City Ship Channel, and in East Bay. Holland and others (1973) found that three of the stations were areas of "normal estuarine stress," or stations having macrobenthic H′ values above 2.0. The stations with normal stress were in upper Galveston Bay near Clear Lake, in East Bay, and in central West Bay. The Trinity Bay and Texas City Ship Channel stations showed evidence of great stress. Holland and others (1973) concluded that the Trinity Bay station was probably stressed naturally, primarily by salinity fluctuations. The Texas City Ship Channel site showed intermittent stress, possibly owing to manmade pollution.

Gilmore and others (1976) found that taxa diversity (numbers of taxa/m^2) declined from the high salinity lower Lavaca Bay to the low salinity upper bay and river area. Taxa diversity was higher during late winter and early spring when sustained fresh-water inflow was generally low.

Gilmore and others (1976) associated high diversity with high salinity, low turbidity, and low nutrient concentrations.

Summary

Benthic organisms are important in estuarine ecosystems because they provide an essential link in the estuarine food web, and they influence the fate of sediment and contaminants in the sediment or water column. A knowledge of animal-sediment relations is important because the benthos are closely associated with sediments, and any change in quantity, suspension, distribution, or deposition of sediments may affect the trophic structure, density, diversity, biomass, and ultimately the productivity of the benthos. Also, the loss of marsh habitat from the combination of a loss in fluvial sediment, subsidence, and sea-level rise would affect the total productivity of the benthos. Any natural or man-induced changes that affect the benthos may upset the fragile ecological balance that is present and may ultimately affect fishery production.

The following findings resulted from this literature synthesis on macrobenthos–sediment relationships:

(1) Trophic structure

(a) Deposit feeders are most abundant in fine-grained sediment with high organic content. In Texas estuaries, deposit feeders are most abundant in the river-influenced assemblage of upper bays.

(b) Suspension feeders are most abundant on sediments of fine to medium sand and on shelly substrates. In Texas estuaries, suspension feeders and other trophic groups are most abundant on oyster reefs and other shelly substrates and in marine grassflats.

(2) Biomass and density

(a) High biomass and density measurements have been reported for sand, mud, shell/gravel, or mixed sediments.

(b) The only studies of Texas estuaries correlating benthic infaunal biomass with sediment were for the Corpus Christi–Nueces Bay system. In those studies, benthic biomass was highest at a muddy, midbay station.

(c) Benthic densities on oyster reefs and other shelly substrates, in marine grassflats, and near tidal inlets are generally high.

(d) Densities in areas with a river-influenced assemblage are highly variable because these areas are stressed from natural disturbance, primarily salinity fluctuations.

(e) Benthic densities are variable seasonally and are probably most dependent on seasonal recruitment cycles, although changes in sediment parameters, even on a small scale, can result in density fluctuations.

(f) There have been no studies of benthic biomass in Texas marshes, mudflats, or sandflats.

(g) Benthic biomass and densities are generally high in marshes, mudflats, and sandflats.

(h) The only study of Texas marshes that correlated benthic densities and sediment was in West Galveston Bay.

(3) Species diversity and species richness

(a) Coarse and heterogeneous sediments, such as gravel/shell and sandy muds or muddy sands, are more structurally complex and thus more diverse than sands or muds.

(b) In Texas estuaries, environmental instability may override structural complexity as the major factor controlling species diversity, especially in areas with a river-influenced assemblage.

(c) Estuarine habitats where structural complexity is important and diversities and species richness are high are oyster reefs and other shelly sediments and marine grassflats.

(d) In areas where physical instability of the substrate is high, such as areas with intensive shrimp trawling activity, benthic diversities and species richness are low.

SUMMARY AND CONCLUSIONS

Sedimentation along the Texas Gulf Coast is affected by various interactive processes including riverine discharge, astronomical and wind-generated tides, waves and currents, episodic events such as fresh-water flooding and tropical cyclones, biodeposition, subsidence and sea-level rise, and human activities. The primary sources of sediments delivered to the estuaries are the rivers that cross the coastal plain. Deposition of the fluvial sediments along the alluvial river valleys and at the river mouths has produced extensive fluvial-deltaic deposits on which marshes and other wetlands—essential components of a healthy estuarine ecosystem—have developed. Sediments delivered to the marshlands not only provide a source of nutrients for sustained plant growth, but they also provide an inorganic foundation necessary to maintain the substrate above a rising sea level. The submergence of more than 4,000 ha (10,000 acres) of fluvial-deltaic wetlands between the mid-1950's and the late-1970's in two areas along the Texas coast signifies that sediments in these areas are not accumulating at rates sufficient to counter the effects of relative sea-level rise.

Wetlands are being lost at a dramatic rate on the Mississippi River delta. In fact, land-loss rates have accelerated geometrically during the 20th century, largely as a result of natural processes, of harnessing the Mississippi River deltaic-sedimentation processes, and of accelerated subsidence (natural and possibly human induced) (Gagliano and others, 1981; Boesch and others, 1983; Wells and Coleman, 1987). Results of several investigations on Louisiana marsh sedimentation indicate that marsh aggradation (vertical accretion) rates are not keeping pace with relative sea-level rise.

There have been few studies of sedimentation in Texas marshes. Investigations have focused principally on shoreline changes to document retreat (erosion) and advancement (accretion) of the shoreline. The loss of interior marshes in fluvial-deltaic areas has only recently been investigated systematically (White and Calnan, 1989) to determine the historical trends in marsh transformation to open water, a process that previously has been documented only in selective areas (White and others, 1985; 1987).

The delivery of fluvial sediments to the bay–estuary–lagoon systems has been a process operating through much of the Holocene Epoch, and of course continues today. The most extensive look at the accumulation of sediments in the bays and estuaries of Texas was done by Shepard (1953); the most thorough investigation of a single bay system (Lavaca Bay) was accomplished by Wilkinson and Byrne (1977). In general, Shepard (1953) concluded, on the basis of bathymetric surveys made in the latter half of 1800's and mid-1930's, that Texas bays and estuaries were shoaling (becoming shallower) at an average rate of 3.8 mm/yr (0.15 in/yr). The highest rates of shoaling occurred at the heads of bays where deposition of fluvial sediments was at a maximum. In bays located away from fluvial input, shoaling rates were much lower, and in some areas deepening of the bay floors had occurred. Wilkinson and Byrne (1977) concluded that historical rates of sedimentation in Lavaca Bay are higher than rates over a geologic time frame (past 8,000 to 10,000 yr) and suggested the higher historic rate may be related to land-use practices (cropland) in the drainage basin.

Only two rivers discharging into Texas estuaries have significantly extended their deltas since the mid-1800's: the Colorado and Trinity Rivers. The Colorado River delta has a unique history of very rapid progradation across the eastern arm of Matagorda Bay following the removal of a log raft upstream that had blocked sediment along the lower reaches of the river channel. Shepard (1953) estimated that the Trinity River had extended its delta about 0.5 km (0.3 mi) since the mid-1800's. On the Guadalupe River delta a small subdelta has prograded into Mission Lake as a result of the artificial diversion of river discharge and sediment load into the shallow lake.

A major process countering the trend toward net sedimentation or shoaling of bay floors and net aggradation of marsh areas is relative sea-level rise. Generally composed of two components, a lesser component of eustatic (global) sea-level rise and a more significant component of land-surface subsidence, relative sea-level rise ranges up to more than 12 mm/yr (0.47 in/yr) along portions of the Texas Gulf coast (Swanson and Thurlow, 1973). Rates are considerably higher than this in areas undergoing human-induced subsidence due to underground fluid withdrawal, such as in the Houston area. More recent comparisons of bathymetric data in the Galveston–Trinity Bay system indicate that water depths have increased as a result of subsidence (Morton and McGowen, 1979).

In the past 40 years, there has been a marked decline in fluvial sediments delivered by many coastal rivers. Among the rivers are the Trinity, San Jacinto, Brazos, Colorado, and Nueces. The sediment loads of other rivers have also possibly diminished, but in many instances sediment-load measuring stations are not located close enough to the coast to adequately reflect the decline. Sediment load in several rivers, for which there is data, is less than half the previous load measured before the 1950's, and in a some cases the load is less than 15 percent of previous amounts. Decreases in stream sediment load are related to different factors including implementation of soil conservation measures. But comparisons of reservoir development in the drainage basins with reductions in stream sediment load indicate reservoirs are probably the major factor. Large reservoirs can trap from 95 to 100 percent of the sediment delivered to them, and reduction in peak flows below the dams decreases the ability of the stream to transport sediments accumulating downstream at the mouths of tributaries. The largest quantities of sediment are delivered to the estuaries during major flood events, which are controlled along streams with large reservoirs.

Channel degradation downstream from reservoirs can contribute sediment to estuarine areas, but the amounts are hard to quantify because of numerous variables involved. Isphording (1986) reported an increase in sand and clay deposition in Apalachicola Bay after reservoir development, but he also reported a striking reduction in silt. Silt, which had been previously supplied under natural conditions, was trapped along with sand by the reservoir. He suggested that clay was washed over the spillways and continued downstream to the bay, and he attributed increases in sand to channel erosion downstream from the reservoirs. Studies by Williams and Wolman (1983) of effects downstream from reservoirs generally indicate substantial decreases in average annual peak discharges with marked reductions in suspended sediment load for hundreds of kilometers downstream. In some major rivers, annual sediment loads did not equal pre-dam values for hundreds or thousands of kilometers. Degradation of channel beds generally occurred during the first decade or two after dam completion.

The environments at the mouths of many rivers along the Texas coast have been significantly modified through canal dredging and sediment disposal activities. These modifications have altered the hydrologic regime and sediment dispersal pathways in some areas, which can hinder natural sedimentary processes and promote erosion.

With relative sea-level rise along the Texas coast matching rates reported along much of the Gulf coast of Louisiana, and with marked declines in fluvial sediments delivered to many Texas coastal fluvial-deltaic and estuarine systems, it is probable that marsh aggradation rates and bay-floor sedimentation rates are no longer keeping pace with rates of relative sea-level rise in many areas. This conclusion is supported by observations in selected areas, including a systematic investigation of historical changes in interior marshes in fluvial-deltaic areas, and measurements of marsh aggradation rates in two areas along the Texas coast (White and Calnan, 1989). More recent comparisons of bathymetric data in Texas bay–estuary–lagoon systems would provide up-to-date information on shoaling or submergence/erosion rates to compare with Shepard's earlier work. Also, establishment and operation of stream-discharge and sediment-load (including bed load) measuring stations on many streams at locations closer to the coast would allow a better estimation of the quantities of fluvial sediments delivered to the bay–estuary–lagoon systems of Texas.

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