Geologic History, Depositional Environment, Processes, and Hydrology of Galveston Island, Texas

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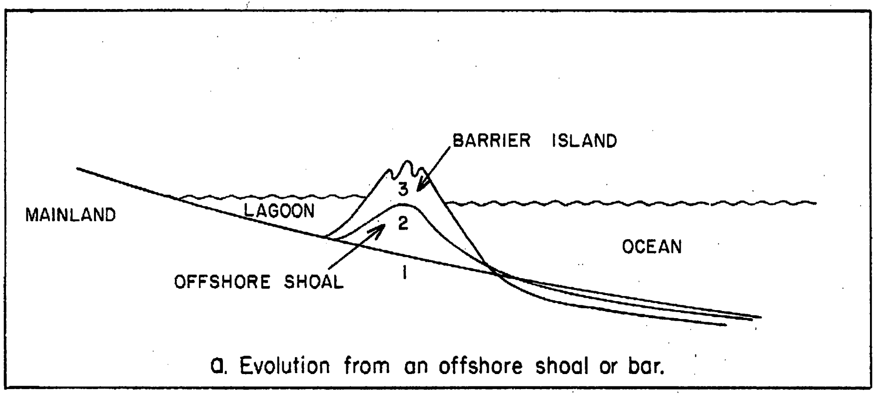
CONTENTS

Figures

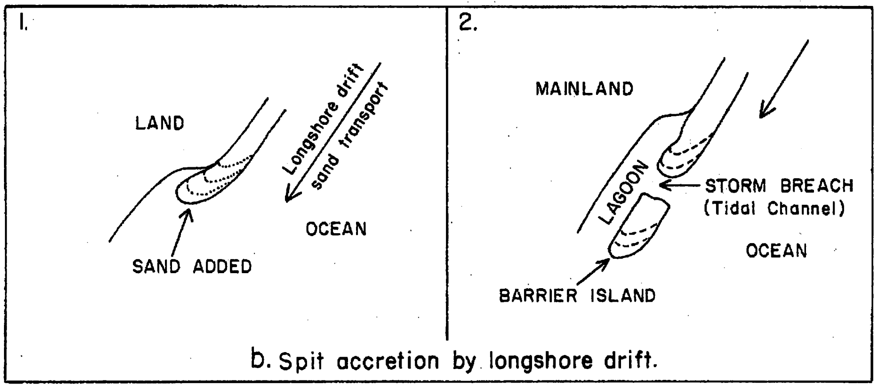
INTRODUCTION

Galveston Island is a very young geologic feature when compared with the Earth: recent estimates place Earth's age at approximately 4.5 billion years. Galveston Island and other Texas barrier islands, on the other hand, began forming as submerged bars no more than 4,500 to 5,000 yr ago, according to radiocarbon dating of shells (Fisk, 1959). The following descriptions of development history, present shoreline conditions, and processes and environments have been modified from LeBlanc and Hodgson (1959), Bernard and others (1970), Fisher and others (1972), Morton, (1974), McGowen and others (1977), Weise and White (1980), and Paine and Morton (1989).

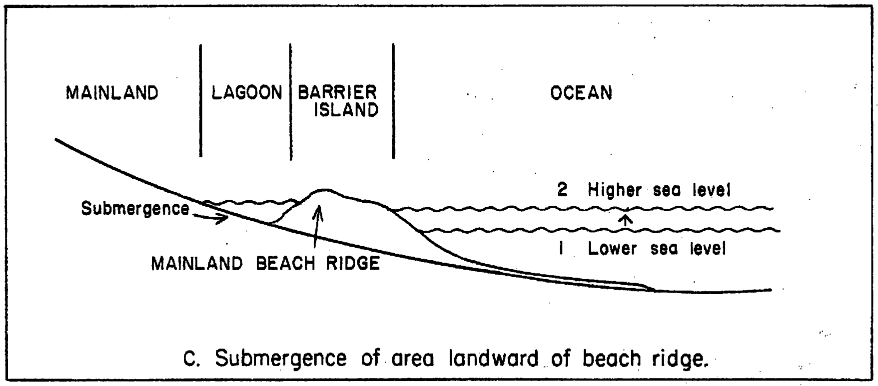
The origin of barrier islands has been debated for years. It is obvious, however, that barriers are formed and modified by different processes or combinations of processes (Schwartz, 1971), depending on such variables as sediment source, sediment type and supply, rate and direction of relative sea-level changes, basin shape, slope of the continental shelf, direction and strength of currents and waves, and magnitude of tides. Three of the most discussed theories of barrier-island origin are (1) development of a barrier island from an offshore shoal or submerged sandbar, (2) development by spit accretion (building) resulting from longshore drift, and (3) development by drowning of the area landward of mainland beach sand ridges (Wanless, 1974) (fig. 1). One possible explanation for the origin of Galveston Island is that it developed from offshore shoals (fig. 1a), later growth being aided by spit accretion (fig. 1b). The offshore shoals might have been old mainland beach ridges submerged during a rise in sea level (fig. 1c). All three processes, consequently, may have played a role in the origin of Galveston Island. It is very likely, moreover, that various segments of the island underwent different processes at different rates during their development.



a. Evolution from an offshore shoal or bar.



b. Spit accretion by longshore drift.



c. Submergence of area landward of beach ridge.

Figure 1. Three theories of barrier-island origin: (a) evolution from an offshore shoal or bar, (b) evolution by spit accretion resulting from longshore drift of sand, and (c) evolution by flooding of area landward of mainland beach sand ridges during a rise in sea level. Modified from Wanless (1974).

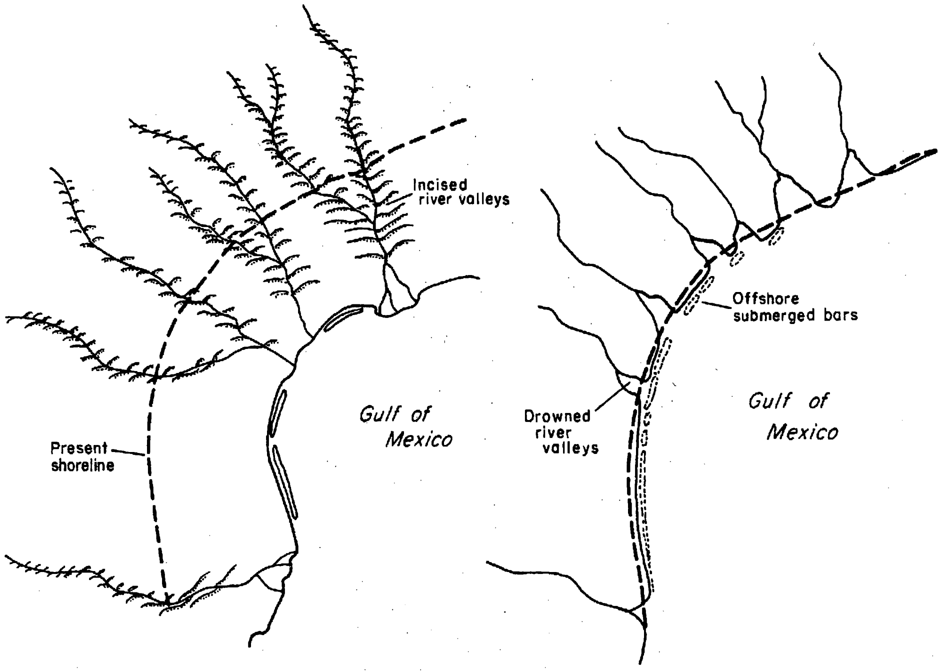
HISTORY OF ISLAND DEVELOPMENT

Geologists generally agree upon the basic stages in Galveston Island's development, although the precise time that each stage occurred is still debated. A discussion of the development of Galveston Island should begin with geologic events immediately preceding its origin. Figure 2 is a schematic representation of the stages leading to the formation of Texas barrier islands, based on LeBlanc and Hodgson's (1959) interpretation of the history of the Texas Gulf Coast. The figure is not meant to reveal the exact geography of the coastline at the various stages but, rather, to show a series of models illustrating probable relationships among sea level, rivers, divides, subaqueous shoals, and islands.

Beginning about 30,000 yr B.P., near the end of the Pleistocene, sea level was low in response to the last episode of glaciation, and rivers along the Texas coast, as well as throughout the world, could no longer shift from their courses. Dropping sea level caused extensive erosion of streams into older, underlying fluvial and deltaic deposits. By the time sea level had dropped more than 400 ft—and rivers were building deltas along a new shoreline scores of miles out on the continental shelf—deep, broad, scalloped-shaped valleys were being cut across the earlier Pleistocene fluvial delta plains. The incised valleys of the Trinity and San Jacinto Rivers record this event.

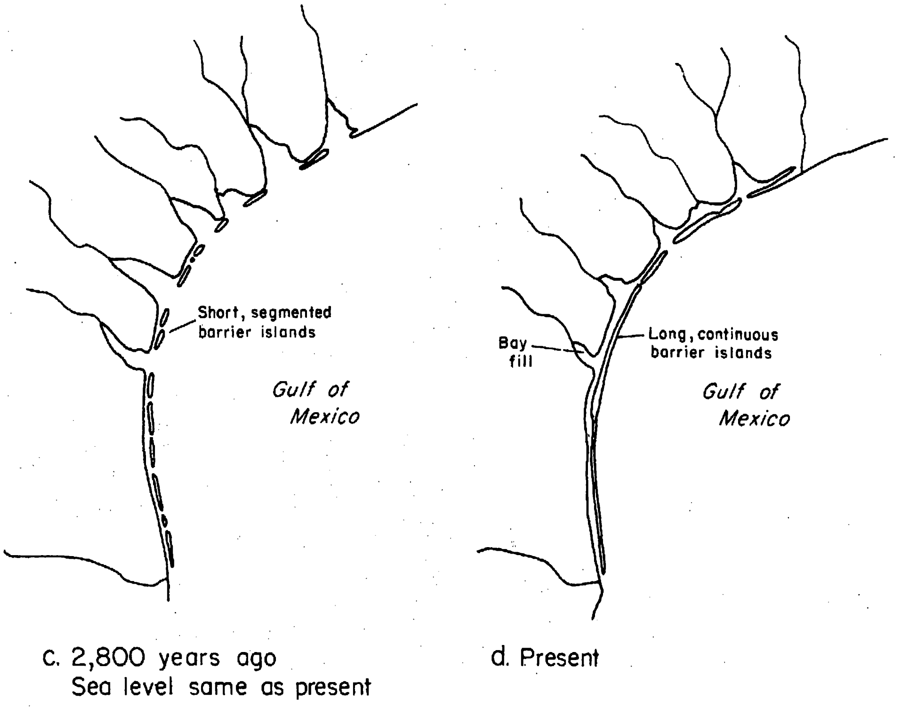
About 18,000 yr ago, near the end of the final (Wisconsin) glacial stage at the end of the Pleistocene, worldwide sea level was about 300 to 450 ft lower than it is now (Curray, 1960). At that time, the shoreline lay much farther gulfward on what is now the submerged continental shelf bordering the Gulf of Mexico. Rivers draining Texas carried sediments across the shelf and deposited them in the Gulf in areas that are now about 50 mi offshore. Upstream, however, rivers scoured deep valleys across the Coastal Plain and emergent inner shelf (fig. 2a).

By about 4,500 yr ago, and after a long period of glacial melting, sea level reached within approximately 15 ft of present sea level (fig. 3). The final small changes in sea level have resulted from compaction of sediment, subsidence of the Gulf Coast area, and minor glacial fluctuations



a. 18,000 years ago
End of Wisconsin Glacial Stage
Sea level −300 to −450 feet

b. 4,500 years ago
End of Holocene
Sea level −15 feet



c. 2,800 years ago
Sea level same as present

d. Present

Figure 2. Sketches representing several stages in the history of the southern Texas coast: (a) 18,000 yr ago, (b) 4,500 yr ago, (c) 2,800 yr ago, and (d) at present. The sketches do not indicate exact configurations of the shoreline but, rather, show relationships among sea level, rivers, divides, subaqueous shoals, and islands. After Weise and White (1980).

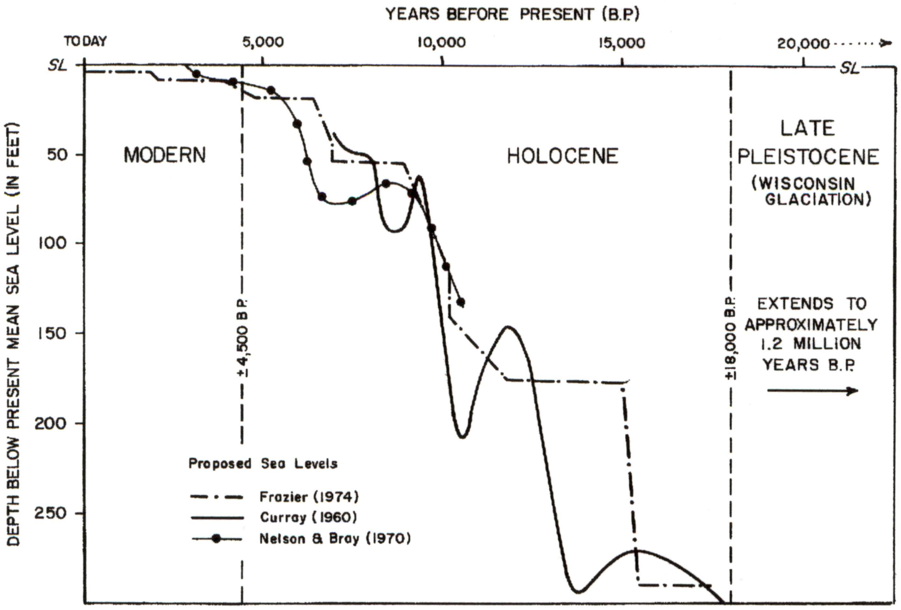


Figure 3. Proposed sea-level changes during the last 20,000 yr; sketch defines use of Modern and Holocene. After Fisher and others (1972).

(Brown and others 1976). The old river valleys carved during the lower stand of sea level flooded (fig. 2b) and became the bays and estuaries along the present Texas coast.

Parts of the Trinity and San Jacinto valleys were drowned by marine water, producing estuaries known as Trinity and Galveston Bays. Modern shoreline erosion has since enlarged the bays, and the deeper parts of the submerged valleys have been filled slowly by bay sediment, although relict meander-cut valley walls are still well defined along the west side of Galveston Bay.

Several thousand years ago, when sea level stabilized near today's level, sand shoals (or bars) that had formed just offshore began to merge. The old submerged river delta and barrier-island deposits laid down farther seaward during times of lower sea level (Pleistocene glacial episodes) were eroded to supply sand for the joining sandbars. As waves and currents carried the eroded sand in toward the shore from the submerged deposits and along the shore from rivers, the bars built up and emerged as a chain of short barrier islands (fig. 2c). These initial islands were positioned primarily on the divides between the old Pleistocene river valleys. The stream valleys thus served as broad tidal passes leading to bays and lagoons behind the emerging islands.

Much of the sand transported by longshore drift (currents moving parallel to the shore) was deposited on the downcurrent ends of the barrier islands, resulting in spit accretion (fig. 4). After a history of shifting, abandonment, and reestablishment by storm breaches, many tidal inlets were eventually closed. A number of short islands were consequently joined to form the longer island present today (fig. 2d).

The barrier islands were built vertically, principally by eolian (wind) processes, and slightly gulfward, by marine processes, as sand carried in from the shelf was added to the shorefaces of the islands. They also built bayward by storm washover deposition. Radiocarbon dating has been used to determine the age and sequence of development of Galveston Island (fig. 5).

When sea level approached its approximately present level, five principal natural changes began along the coast: (1) deeper parts of the Trinity and San Jacinto estuaries began to fill with sediment eroded from the walls of drowned valleys; (2) the Trinity and San Jacinto bayhead deltas began their slow filling of the uppermost parts of the estuaries; (3) headward erosion by short

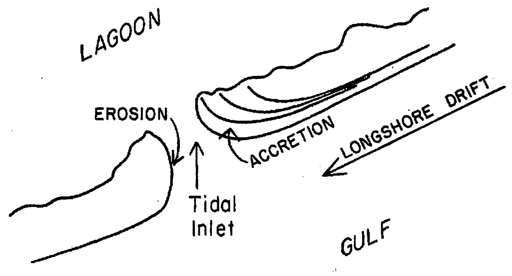


Figure 4. Spit accretion. Sand carried by longshore currents is deposited on the downcurrent end of a barrier island. The upcurrent end of the adjacent island may erode, causing the tidal inlet to shift in the direction in which the currents move. The tidal inlet will be closed, however, if the rate of accretion exceeds the rate of erosion. After Weise and White (1980).

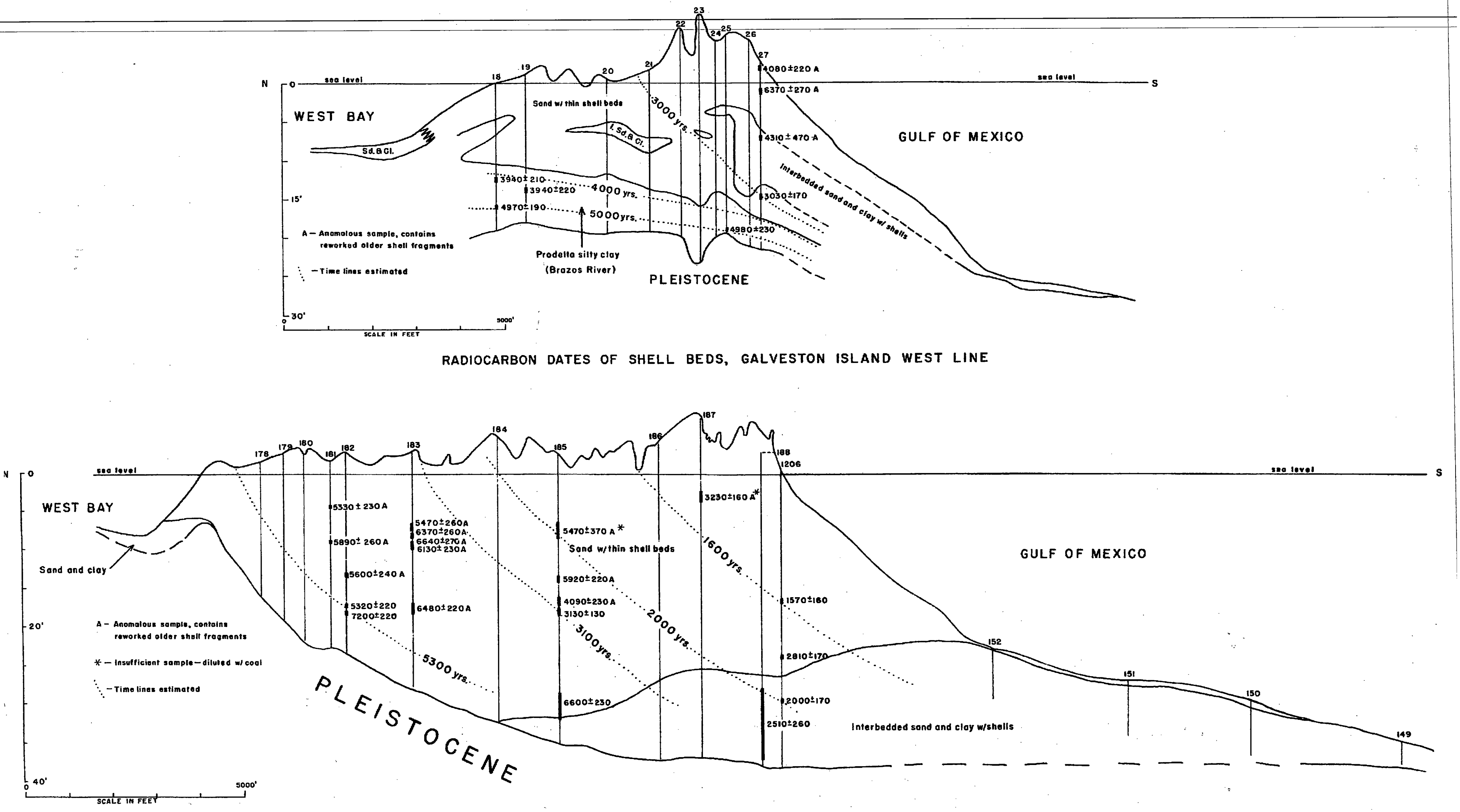


Figure 5. History of the development of Galveston Island based on radiocarbon dating. After Bernard and others (1970).

streams continued within Pleistocene interdistributary areas, where significant compaction of mud is now occurring; (4) East Bay and West Bay developed as elongate lagoons behind Bolivar Peninsula, which grew southwestward by spit deposition and shoreface deposition from eroded deltaic headlands near High Island, and behind Galveston and Follets Islands, which developed as coalescing, exposed offshore bars that also grew seaward by shoreface deposition; and (5) marshes encroached upon subsiding Pleistocene delta deposits and bay areas that were filled by storm washover fans and bay-margin deposits.

Barrier-Island and Peninsula Accretion

When sea level reached its 4,500-yr-B.P. level, sands eroded from Pleistocene deltaic headlands, and submerged Pleistocene sands on the inner shelf moved southwestward by longshore currents and shoreward by wind-generated waves. These sands were deposited as spits and offshore bars that eventually coalesced into the present 55 mi^2 of sand that compose Bolivar Peninsula, Galveston Island, and Follets Island sandstone bodies. Relict beach ridges testify to a slow seaward growth or accretion by sand from longshore and onshore currents. Shoreface sands grade gently seaward into shelf mud and silt. High-energy, shifting tidal passes have maintained communication between bays and the Gulf, and persistent winds and hurricanes blow beach sand into dune ridges. Storms have breached the low, narrow sandstone bodies many times, building washover fans landward into shallow bays. The back sides of the barriers are fringed by tidal sand marsh.

During the past 4,500 yr, compaction of sediment and slow subsidence of the Gulf Coast basin have resulted in relative changes in sea level of about 10 to 15 ft.

PRESENT SHORELINE CONDITIONS

There appears to be a natural sequence of stages in the life of a barrier shoreline: (1) an accretionary, or building, phase; (2) a phase of stability, or equilibrium; and finally, (3) a stage of erosion, or destruction (McGowen and others 1977). Various segments of Galveston Island probably have experienced these phases at different rates and at different times.

On the basis of long-term beach stability, the Gulf shoreline of Galveston Island can be divided into three zones (Paine and Morton, 1989): (1) a zone of sand accumulation between the seawall and Bolivar Roads (East Beach), (2) a zone characterized by very little beach sand in front of the Galveston seawall, and (3) a generally recessional zone between the west end of the seawall and San Luis Pass (West Beach).

Most of East Beach advanced between 1974 and 1982 (fig. 6). Rates of shoreline advance increased toward the northeast from 3.5 ft/yr near the seawall (station 4) to 23.4 ft/yr adjacent to the jetty (station 1). Rates of East Beach advance were higher than those observed between 1956 and 1970, when only the two stations nearest the jetty advanced, station 3 remained stable, and stations 4 and 5 eroded (Morton, 1974). Rates of advance between 1974 and 1982, however, were below the long-term (1930 through 1970) rates along East Beach.

The shoreline along the seawall (stations 5 to 12) was relatively stable between 1974 and 1982. Significant accumulations of sand are found only in pockets adjacent to the short groins protruding into the Gulf; in many areas, riprap protecting the base of the seawall is the shoreline. Rates of erosion are lower than longer term (1933 through 1973) rates simply because no beach remains.

The 9 mi of shoreline west of the seawall (stations 13 to 21) retreated between 1974 and 1982. Retreat rates increased from 2.3 ft/yr 9 mi from the seawall (station 21) to 11.6 ft/yr at the west end of the seawall. Rates were even more recessional between 1956 and 1973. Long-term (1933 through 1973) rates in this area were also recessional but lower than rates during more recent monitoring periods (1956 through 1973 and 1974 through 1982).

The highest rates of shoreline movement between 1974 and 1982 on Galveston Island were recorded along 3 mi of beach east of San Luis Pass (stations 28 to 31). Shoreline adjacent to the pass retreated as much as 33.8 ft/yr; slightly farther east, it eroded more slowly (3.3 to 9.8 ft/yr at

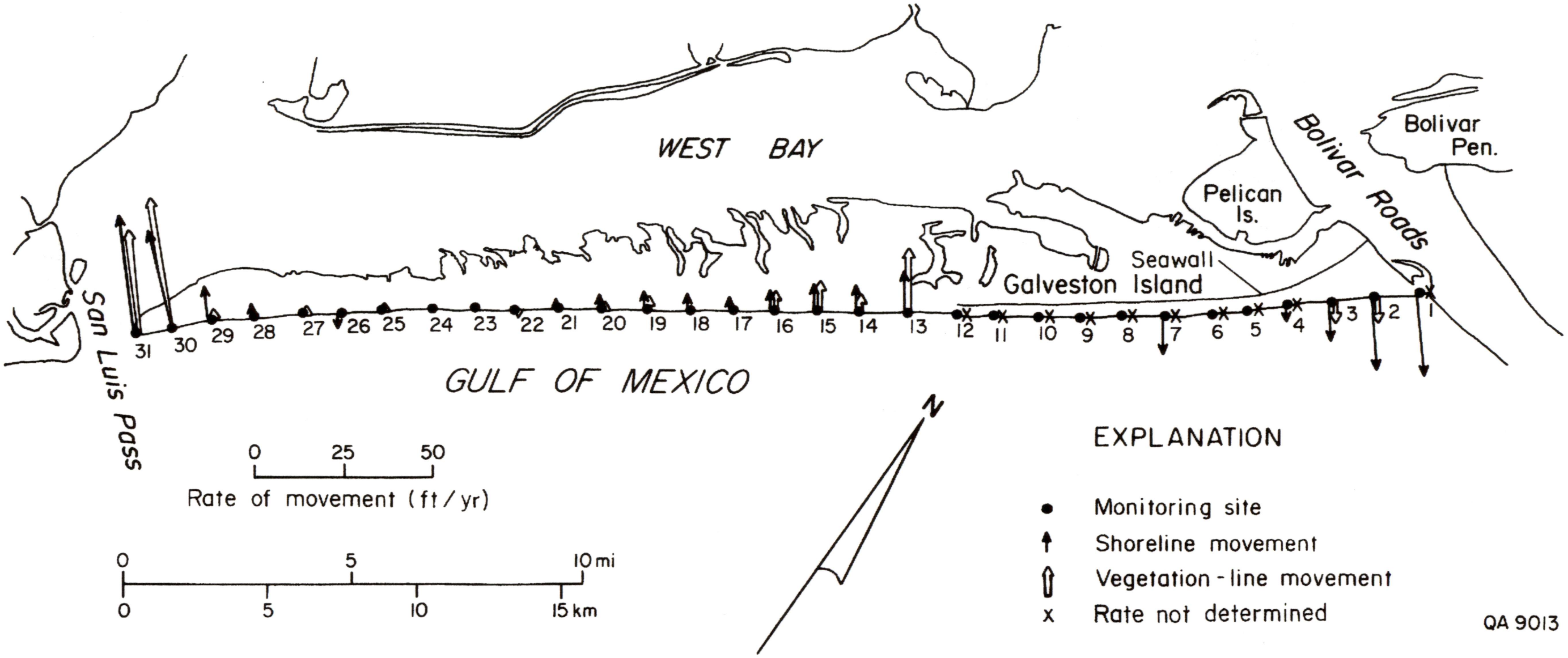


Figure 6. Shoreline and vegetation-line changes between Bolivar Roads and San Luis Pass, 1974 through 1982. After Morton (1974).

stations 28 and 29). In most of this area, retreat between 1974 and 1982 occurred at a rate much higher than the long-term (1930 through 1973) average.

The shoreline between the erosional beaches near San Luis Pass and the area west of the seawall (stations 22 to 27) was stable or slowly advancing (as much as 1.9 ft/yr) between 1974 and 1982. Beach profiles conducted by the U.S. Army Corps of Engineers (1968 through 1980) substantiate the stability of this segment between 1973 and 1980.

A beach replenishment project on Galveston Island in the area seaward of the seawall was completed in 1995. Sand was transported from channel-dredging operations to the beach area between the jetties along the seawall. The sand provided by this operation replaced beach sand that had been depleted and removed by erosion.

DEPOSITIONAL ENVIRONMENTS AND PROCESSES

Beach

Beaches (fig. 7) that are accreting or building seaward have two distinct zones: (1) forebeach, the seaward-sloping, smooth part of the beach that is affected daily by swash, and (2) backbeach, which is normally separated from the forebeach by a berm. Because certain segments of Galveston Island are building outward or are in equilibrium, they display both forebeach and backbeach zones.

Beach segments that have ongoing erosion or that have physical energy exceeding sediment availability have relatively narrow forebeaches and backbeaches, and backbeaches may commonly be missing. They consist of high proportions of shell to sand in the lower and upper swash zones. Berms as much as 5 ft high separate the forebeach and backbeach.

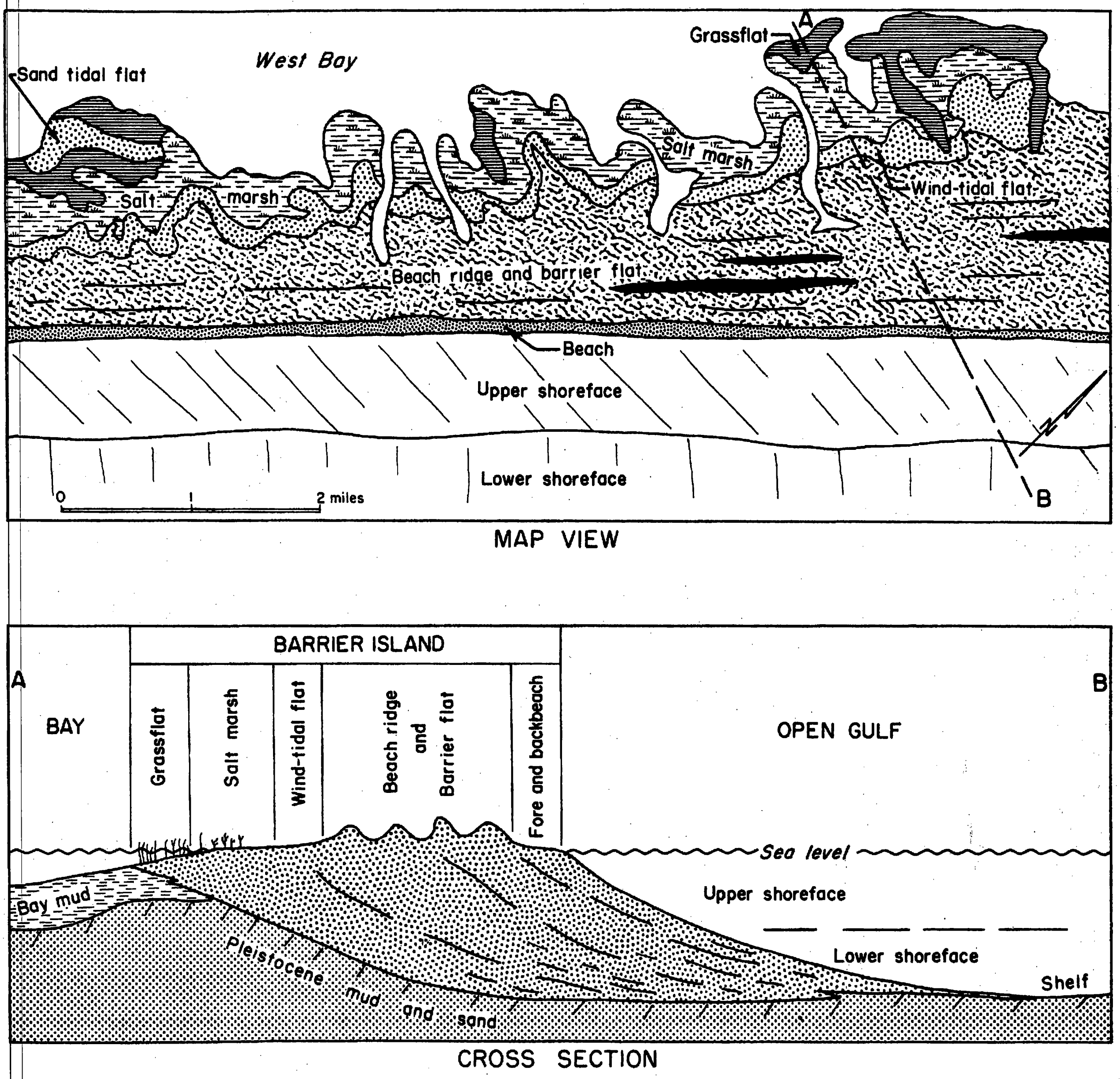


Figure 7. Modern barrier-bar environments and facies, Galveston Island. Cross section after Fisher and others (1972).

Beach Ridge and Barrier Flat

The beach ridge and the barrier flat make up the major environment of the barrier system on Galveston Island. Terrain is characterized by a series of subparallel ridges and swales generally oriented along the main trend of the barrier island (fig. 7). Each ridge represents a former shoreline position during earlier stages of barrier development. The greatest concentration of ridges on Galveston Island lies near Galveston and extends westward for about 25 mi. Ridge height is generally about 5 ft. Locally, ridge crests have a maximum elevation of 10 ft above sea level. Beach ridges, generally paralleling the shoreline, may extend for several miles; the area of beach ridges and barrier flats is widest on the northeast end of Galveston Island but becomes much narrower on the southwest end near San Luis Pass. Certain beach ridges on Galveston Island curve sharply toward the bay, representing spit accretion into old, relict tidal passes that have subsequently been filled by the migrating spits. Spit migration occurred southwestward in the direction of present longshore drift, as indicated by the trend of curved beach ridges.

The growth of beach ridges 5 to 10 ft above sea level is a function of several interacting coastal processes. Sand and shell material from which the ridges have been constructed was derived from offshore and moved onshore by wind-generated currents. Under normal sea conditions the strandline builds seaward by accumulation of sand on the beach. Spring tides and storms raise sea level, temporarily allowing sand to accumulate as berms a few feet above mean sea level. With return to normal sea level, the berm is modified by wind and biologic processes. Subsequent spring tides and storms create another berm, which is accreted to the previous berm.

Also situated between the beach and the wind-tidal flat are areas in which no obvious beach ridges or swales exist and which constitute the vegetated barrier flat (fig. 7). The barrier flat lies about 5 ft above mean sea level on the Gulf side of the barrier and between sea level and 5 ft along the bayside. The surface of the barrier flat slopes gently bayward. Vegetation on the flat, as well as on the beach ridges, is predominantly grasses that are tolerant of salt spray. Locally some small mottes of oak are present. The barrier flat is formed chiefly by sediments blown from the area of the beach ridge and also by sediments deposited by storm washovers.

Wind-Tidal Flat

A flat, barren, relatively featureless surface occurs along the back side of the barrier islands between the vegetated barrier flat and beach ridge and the salt marshes along the bay shore. This area constitutes the wind-tidal flat. Inundation by salt water occurs a few times each winter during passage of a polar front, and duration of flooding is directly related to the duration of the north wind. Because the area is flooded only a few days each year, most of the surface and near-surface salt water evaporates, leaving a thin salt crust on the flat surface. Although blue-green algae flourish on these flats during and shortly after flooding, the environment is largely barren of vascular plants. Some local salt-marsh vegetation exists, and *Uca*, the fiddler crab, commonly burrows the lower parts of the flats.

Salt Marsh

The back side of the barrier islands, extending bayward of the wind-tidal flat, supports salt marshes. The marshes display an orderly plant succession from the bay line to higher parts of the barrier. The succession is controlled by factors such as degree of inundation, salinity of the substrate, and height of the marsh surface above bay water level. From the bay line toward the higher marsh areas, the typical plant succession is (1) *Spartina alterniflora;* (2) *Batis, Salicornia, and Distichlis;* (3) *Spartina patens, Monanthochloe, Suaeda,* and *Borrichia;* and (4) sparse marsh vegetation in hypersaline areas.

Marshes are indented on the bayside by tidal cuts that are curved to the west, reflecting general westerly longshore drift in the bays. During northers some oyster shell and shell from other bay species are washed into the marsh, causing thin, narrow, discontinuous beaches to develop. With the exception of shell beaches, sediments underlying the marshes become coarser or sandier from the bay margin to the higher parts of the marsh. Sediments underlying low marshes are generally dark gray mud or muddy sand that are intensely burrowed by worms, crustaceans, and mollusks and mottled by penetration of plant roots. Sediments underlying higher marshes are dominantly sand and muddy sand. Sediments of the high marsh are reworked primarily by plant roots and fiddler crabs.

Subaqueous Sand Flats

The barrier island extends some distance beyond the marsh line into the bay; this extension is marked by a shallow sand flat that terminates rather abruptly in water between 2 and 6 ft deep. Sand flats are commonly stabilized by marine grasses—for example, bayward of Follets Island—that are commonly designated as grassflats. Other parts of the sand flat are barren or only slightly vegetated. Sediments that accumulate on the grassflats and sand flats are chiefly fine-grained sands derived from the adjacent tidal passes and tidal deltas. Sand is transported to the west along the bayside of Galveston and Follets Islands by longshore current set up by wind-generated waves (fig. 8).

Tidal Passes and Tidal Deltas

Natural breaks between barriers through which there is tidal exchange between bay and Gulf waters are termed *tidal passes*. Sediments move into the bay with flood tides, and part of the sediment load accumulates as fan-shaped bodies near the terminus of the tidal channel; these compose *flood-tidal deltas* (fig. 9). During ebb tide, sediments are transported from the bay seaward through the tidal pass. Because physical processes are much stronger on the Gulf side of the barriers than on the bayside, much of the sediment is moved immediately southwestward by longshore currents. Accordingly, ebb-tidal deltas are poorly developed and form a simple seaward bulge, along with some sand shoals near the mouth of the pass. Unlike flood-tidal deltas, the ebb-tidal deltas of the Texas Coast never become emergent environments.

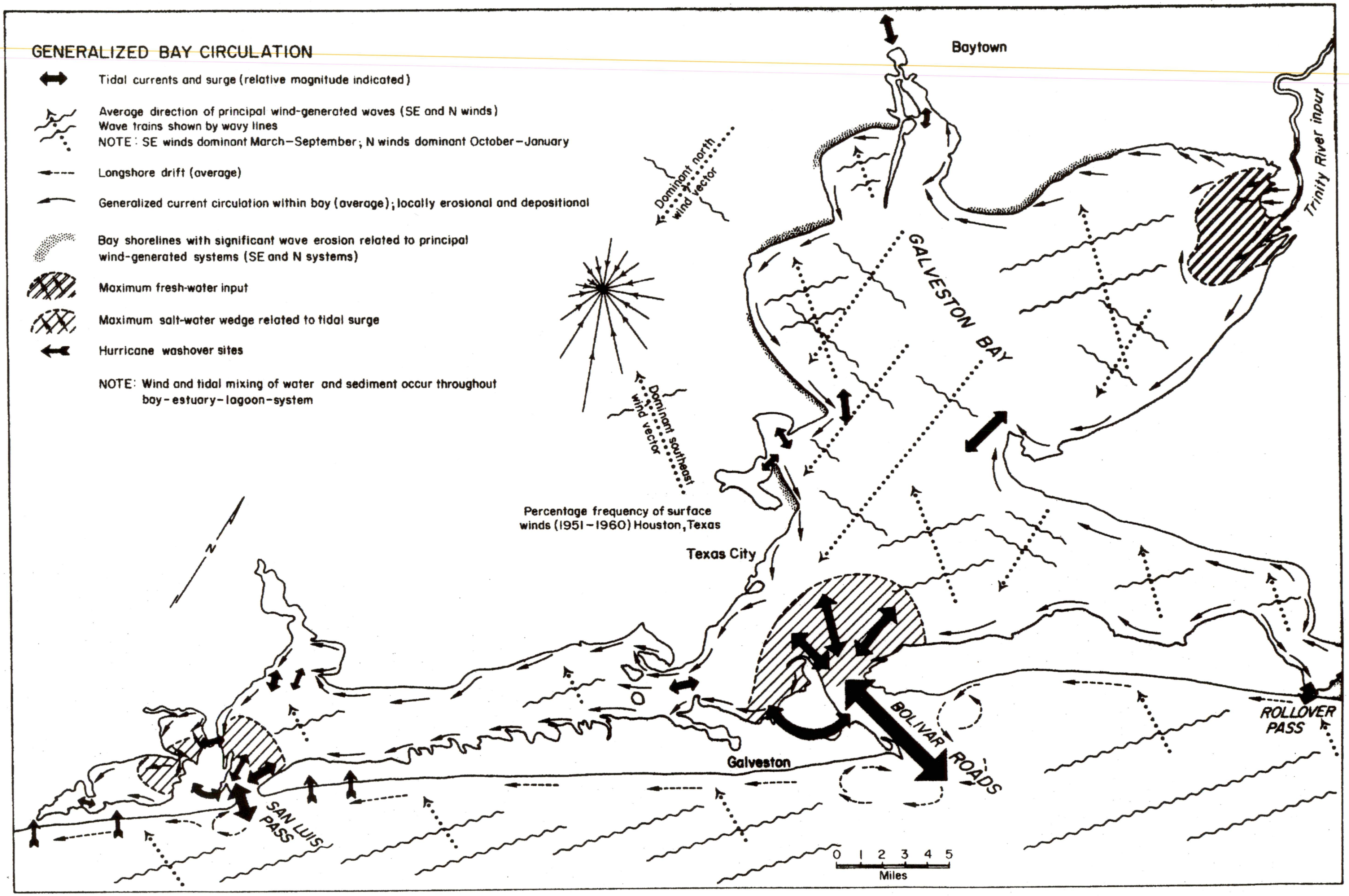


Figure 8. Circulation, waves, sediment transport, and other physical processes, bay-estuary-lagoon system, Galveston–Houston area.

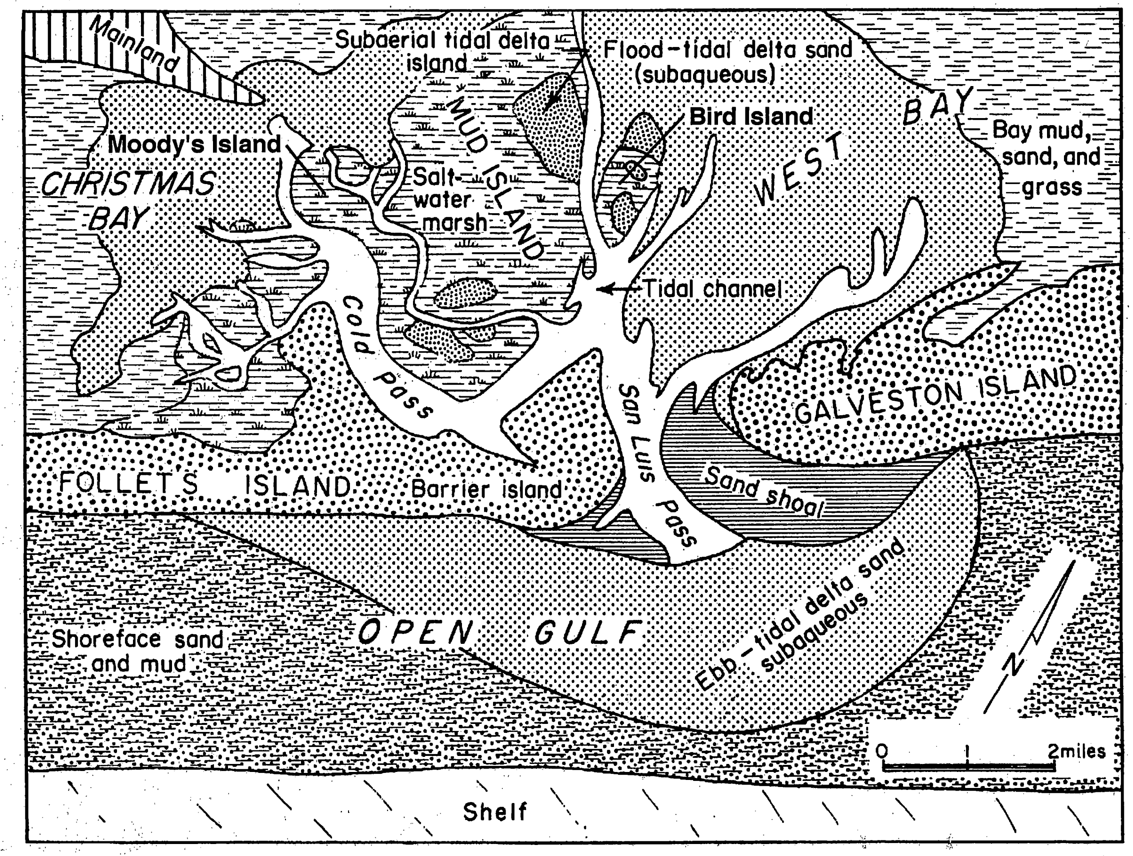


Figure 9. Modern tidal-delta facies, San Luis Pass, West Bay, Brazoria and Galveston Counties, Texas. These facies are developed by ebb- and flood-tidal currents. Marsh environments occupy much of the shallow delta fan. After Fisher and others (1972).

Major tidal passes on the Texas Coast, such as Bolivar Roads at Galveston (fig. 8), are situated over buried Pleistocene valleys. Depth of the buried relict valley at Bolivar Roads is approximately 260 ft. These valleys are filled in their deepest parts by river gravels and sands, succeeded upward by deltaic sediments, estuarine deposits, and, near the surface, by tidal-channel deposits. Deposits in the deeper parts of the tidal channels consist of a broken shell lag. Channels are unstable; they tend to migrate in the direction of longshore drift. As the channel migrates it is successively filled by spit accretion.

Flood-tidal deltas consist of shell and sands near the mouth of the main tidal channel, sediments becoming finer on the distal parts of the deltas toward the bay. Tidal deltas become emergent when storms raise the water level in the bay, allowing sediment to build vertically. With subsidence of the storm and associated high tides, parts of the flood delta become emergent and may be subsequently stabilized by marsh vegetation. Mud, Moody's, and Bird Islands on the bayside of San Luis Pass are examples of emergent flood-tidal deltas (fig. 9).

Washover Channels and Fans

During hurricane surges and storms, the barrier island locally may be breached (fig. 10). Storm-generated currents cut channels through the barrier and carry sand to the bayside, where it is deposited as a washover fan. During normal periods, sand transported by longshore currents fills or heals these washover channels along the Gulf side. They may be reopened during subsequent storms. The washover fans along the bayside, consisting mainly of unvegetated sand, ultimately may become stabilized by marsh vegetation.

HYDROLOGIC CONDITIONS

Sands that have accumulated to form Galveston Island provide a natural reservoir for ground water. A cross section of the island (fig. 11) illustrates the relative proportion of sand available and a generalized representation of the base of the fresh-water lens. Fresh ground water is commonly

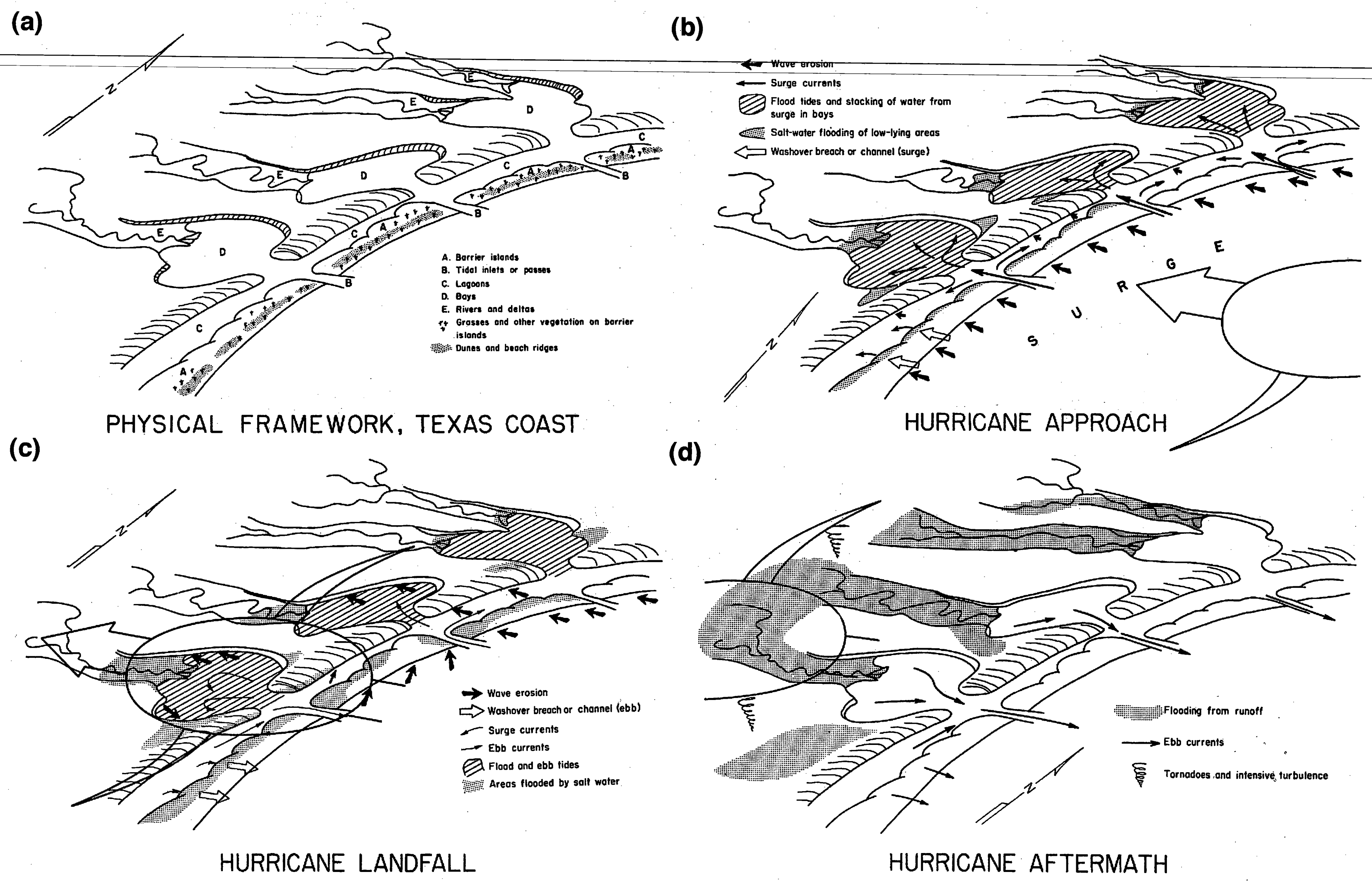


Figure 10. Schematic model of hurricane effects on the Texas Coastal Zone. (a) Physical features characterizing Texas coast. (b) Effect of approaching hurricanes. (c) Effect of hurricanes upon impact with coast. (d) Aftermath effects of hurricanes. After McGowen and others (1970).

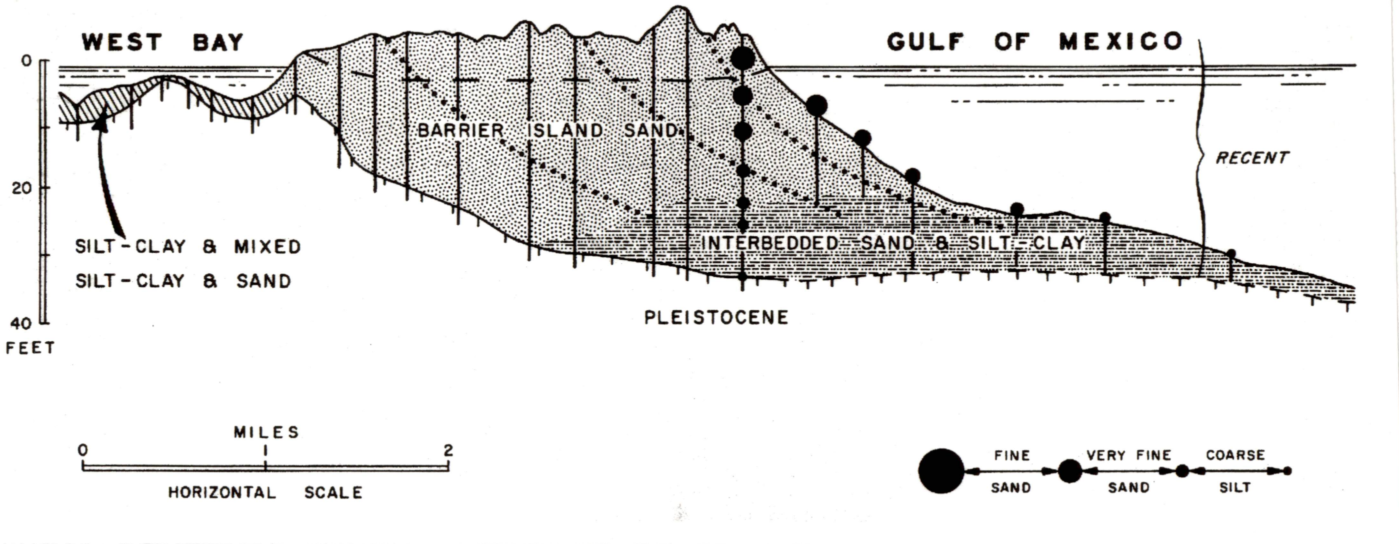


Figure 11. Generalized cross section of Galveston Island along Eight Mile Road. The clean, well-sorted sandstone body offlaps the interbedded sand and silt clay, which are deposited seaward of the offshore break in slope. Dashed line represents approximate boundary between fresh and saline water. Modified from Bernard and others (1970).

present a few feet below the surface. Near the base of the fresh-water lens the water becomes brackish and grades into more saline water with depth.

The amount of fresh water that can be stored in the barrier-island sands is directly proportional to the amount of sand that has accumulated above sea level. Because the maximum ridge height on Galveston Island is only 5 to 10 ft above sea level, very little storage capacity exists for fresh water. Wells that obtain water from the fresh-water lens are commonly only a few feet deep and are used only to supply water for livestock.

Several water wells on Galveston Island penetrate Pleistocene sands known as the Chicot aquifer at depths of about 300 ft.

ACKNOWLEDGMENTS

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