Geological Circular 74-1

Depositional - Episodes: Their Relationship to the Quaternary Stratigraphic Framework in the Northwestern Portion of the Gulf Basin

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DAVID E. FRAZIER

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DEPOSITIONAL-EPISODES: THEIR RELATIONSHIP TO THE QUATERNARY STRATIGRAPHIC FRAMEWORK IN THE NORTHWESTERN PORTION OF THE GULF BASIN[1]

by

David E. Frazier[2]

ABSTRACT

The stratigraphic record yields evidence that each episode of clastic silicate deposition has been of limited duration and that each has been preceded and followed by a significant hiatus. Evidence for alternations of deposition and non-deposition is readily apparent in the landward portions of Pleistocene deposits along the Gulf Coast, due to the glacio-eustatic changes in sea level; evidence of alternations, although elusive, exists also in the basinward portions of these deposits. The concept of depositional-episodes explains the significance and relationship of these alternating conditions throughout the basin.

The strata attributed to each depositional-episode are a composite of several discrete facies-sequences and are referred to in this paper as a depositional-complex. Each facies-sequence represents either a single delta lobe within a deltaic progression, or one of the several repetitive sequences deposited in an interdeltaic environment.

Each depositional-complex records and defines a depositional-episode and indicates three phases of development. Deposits of the initial phase record a stillstand of the sea during which each of the several rivers entering the basin prograde a succession of delta lobes and interdeltaic facies-sequences across the shelf. The second phase of development (which is penecontemporaneous with the first) is recorded by the intercalation of clastic and organic flood-plain deposits which accumulate on the newly formed coastal plain, and by the deep-water hemipelagic basin sediments which are secondarily derived from unstable sediments deposited in the outermost shelf and uppermost slope environments. The terminal phase is evidenced by sediments deposited during a period of instability when a marine transgression either continuously or intermittently forces estuarine conditions on the rivers entering the basin. Throughout the terminal transgression, the finite zone of active deposition adjacent to the shoreline is shifted landward. Basinward of this active zone of deposition, hiatal conditions are imposed and at the instant of maximum transgression, when the depositional-episode is terminated, all points on the hiatal surface are synchronous.

The bounding surfaces of depositional-complexes represent natural stratigraphic breaks over the entire basin and are related to hiatal conditions imposed by marine transgressions. Within the Quaternary section, the repetitive alternation of depositional-episodes and significant hiatuses is due to the glacio-eustatic fluctuations of sea level: as a result, worldwide correlations of the Quaternary depositional-complexes and hiatal surfaces may be possible.

INTRODUCTION

The northwest portion of the Gulf basin is a classic example of the deposition of clastic silicates in a slowly subsiding geosyncline with perhaps more than 20,000 feet of sediments in the Quaternary section alone. In addition, much information has been obtained from this area, and compilation of copious data is possible. Within the Quaternary section, however, problems concerned with proper correlations between the deep marine deposits and the shelfal and terrigenous deposits become more and more apparent.

In 1944, Fisk showed the relationship of the Pleistocene terraces on the coastal plain (fig. 1) to mappable units in the subsurface. The bounding surfaces of the mappable units, he pointed out, are the oxidized soil zones. These are hiatal surfaces which correlate with the glacial stages when sea level was low and the shoreline was at the

[1] This paper was presented orally at the S.E.P.M. Symposium, The Marine Quaternary of the Caribbean and Gulf of Mexico Regions, held at the G.C.A.G.S. Convention, Miami Beach, Florida, October 29-November 1, 1969.

[2] Esso Production Research Co.

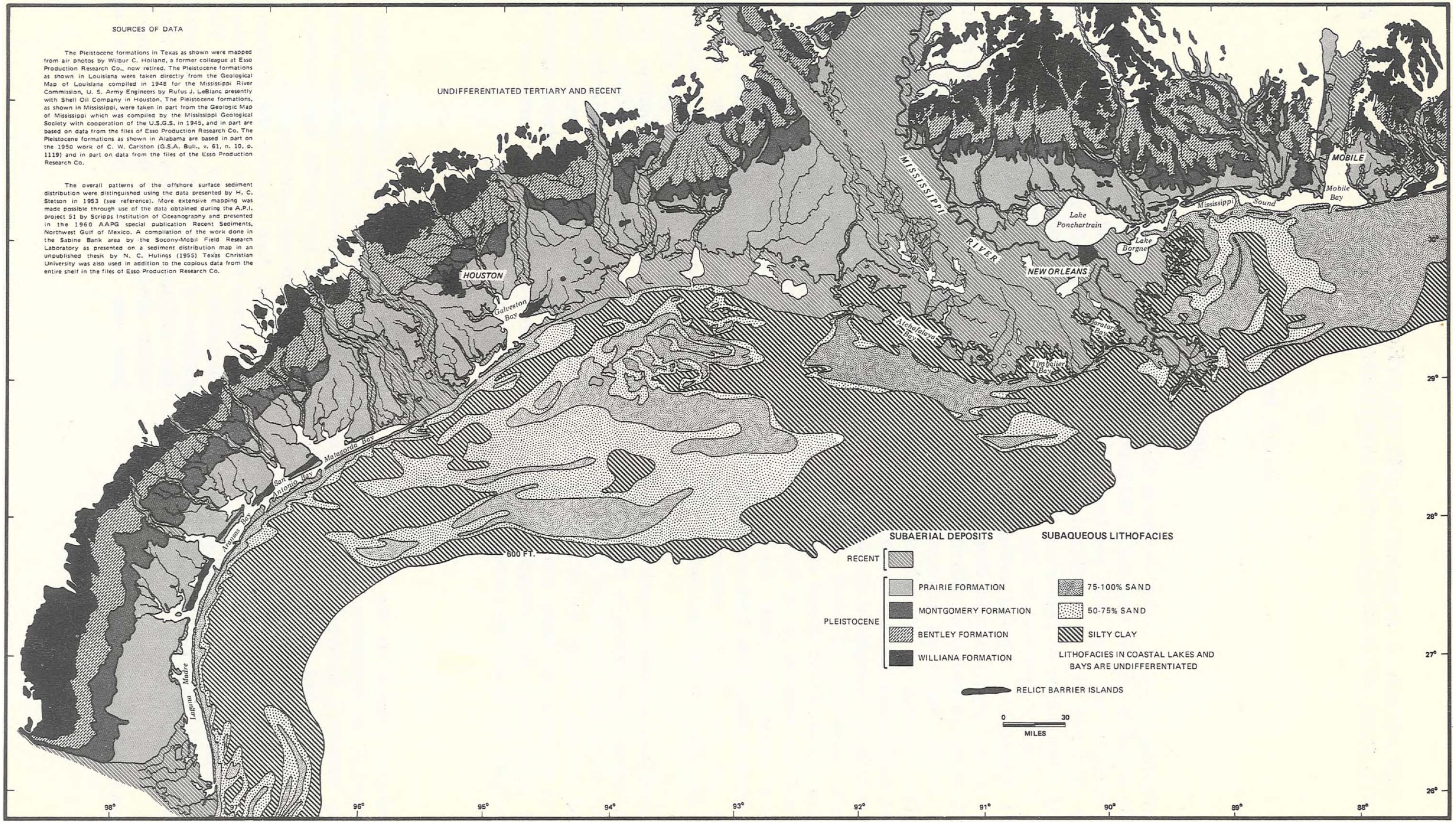

Figure 1.—Exposed Quaternary deposits in the northwest region of the Gulf of Mexico.

basinward limit of the continental shelf. The sediments of the mappable units (between the oxidized zones) were, therefore, deposited during the warm interglacials. The alternation of deposition and nondeposition is readily apparent on the coastal plain: the terraces represent periods of continuous inland uplift and intermittent deposition. Deposition occurred during high stands of the sea, and nondeposition during low stands. Because the oxidized surfaces extend only to the low-sea-level shorelines at the basinward limit of the continental shelf, the mappable units of Fisk are limited to the basin margin.

How then is correlation achieved throughout the basin? During the search for the solution to this stratigraphic problem, the concept of depositional-episodes evolved. It is the intention of this paper to present and document three sedimentologic principles that support the concept of depositional-episodes, to define "depositional-episode" and associated terms, and then to relate these concepts to the stratigraphic framework of the Quaternary silicate sediments in the northwest portion of the Gulf basin, thereby offering a solution to the correlation problem.

SEDIMENTOLOGIC PRINCIPLES

It has been possible to document the three sedimentologic principles stated below in the northwestern Gulf. Until recently, however, their significance went unnoticed, but as the concept of depositional-episodes evolved, their significance became more and more apparent. The discussion which follows relates these principles to the stratigraphic framework of Quaternary deposits in the northwestern Gulf.

1. Clastic silicate sediments are allochthonous and must be brought to the basin margin by rivers.
   - (a) The sources of the bulk of the clastic silicate sediments that fill a basin are the many stream mouths along the basin margin.
   - (b) Basins are filled from the margin toward the center.
2. Basin filling by clastic silicate sediments is achieved by a repetitive alternation of depositional and nondepositional intervals.
   - (a) Deposition does not occur everywhere at any time.
   - (b) Deposition is not continuous anywhere. (Infinitesimal amounts of animal tests, cosmic dust, etc., deposited on a hiatal surface are not considered here.)
   - (c) Hiatal surfaces separate discrete stratigraphic units. (Hiatal surfaces, as used in this report, receive only infinitesimal amounts of sediment, and although measurable deposits may accumulate under these conditions over millions of years, they are insignificant in thickness compared to deposits from terrigenous sources during a depositional-episode.)
3. In a basin being filled by clastic silicate deposition, all points on a hiatal surface do not represent the same duration of time. One instant in time is common to all points, however.
   - (a) The surface upon which a progradational unit of sediments is deposited represents a progressively longer interval of nondeposition in a basinward direction.
   - (b) The surface upon which a transgressive unit of sediments is deposited represents a progressively longer interval of nondeposition in a landward direction.

conversely:

   - (c) The upper surface of a progradational unit represents a progressively longer duration of time in a landward direction.
   - (d) The upper surface of a transgressive unit represents a progressively longer duration of time in a basinward direction.

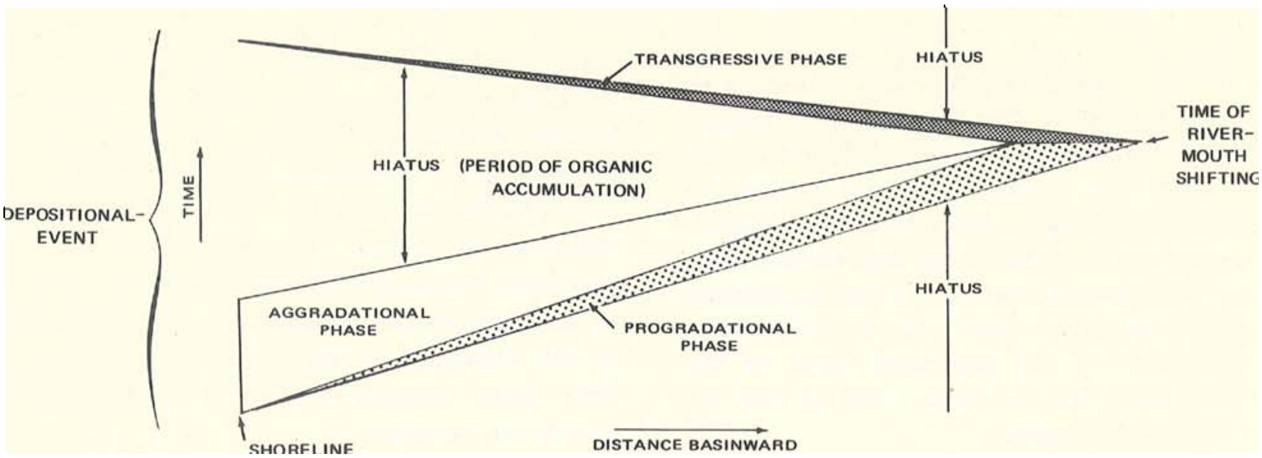

Figure 2.—Depositional-event.

DEPOSITIONAL EPISODES—DEFINITIONS AND DISCUSSION

Depositional-Events and Facies-Sequences

Depositional-events are localized pulses of deposition which are of varying magnitude. In addition, they are separated by hiatuses of varying duration. Depositional-events are not concurrent for all rivers entering a basin; the sporadic shifting of a river course from one depositional site to another leads to the end of one depositional-event and initiates the next at the new site of deposition. For this reason, the temporal extent of any depositional-event cannot be correlated with that of another or with any set unit of time. A depositional-event is depicted on the time-distance diagram of figure 2. The boundaries of the progradational, aggradational, and transgressive phases transgress time as shown. The depositional hiatus that exists between the aggradational and transgressive phases increases in duration landward and is often interrupted by intervals of peat accumulation. The hiatuses that precede and follow the depositional-event increase in duration and merge basinward; they are often marked by an increase in the abundance of marine faunal remains, due to the decrease in clastic deposition and consequent decrease in dilution.

Each depositional-event is recorded and defined by a facies-sequence that reflects an initial progradation, a penecontemporaneous and intermediate aggradation, and a terminal transgression, and occurs within a relatively short interval. All facies within each facies-sequence are genetically related to a common sediment source. The discrete facies-sequence may be similar to an interdeltaic facies-sequence (fig. 4) or to the facies-sequence of a single delta lobe (figs. 5 and 6).

Depositional-Episodes and Depositional-Complexes

Several depositional-events are contained in a depositional-episode, as diagrammed in figure 3. Each depositional-episode is recorded and defined by a depositional-complex that, in turn, consists of several facies-sequences. During each interval of sea-level stability, a basinward progression of facies-sequences occurs. The multiplicity of discrete facies-sequences from any given point source along the basin margin is due to river shifting and the localization of deposition in a fairly restricted area. Consecutive facies-sequences during sea-level stillstands, therefore, are not superposed but are laterally displaced, and hiatal surfaces exist between successive offlapping facies-sequences along any given line of section, as shown on figures 3, 4, 5, and 6. The intervening hiatuses are correlative with portions of the progradational and aggradational phases of the depositional-events which occurred elsewhere when the river shifted. The delta lobes of figures 5 and 6 are numbered on figure 7 in their sequence of development (Frazier, 1967) and demonstrate the shifting of a single river. The interdeltaic facies-sequences of figure 4 are also numbered on figure 7: the numbers correspond to the delta lobes from which the sediments were derived (fig. 8).

Each depositional-complex is a composite of facies-sequences derived from all of the point

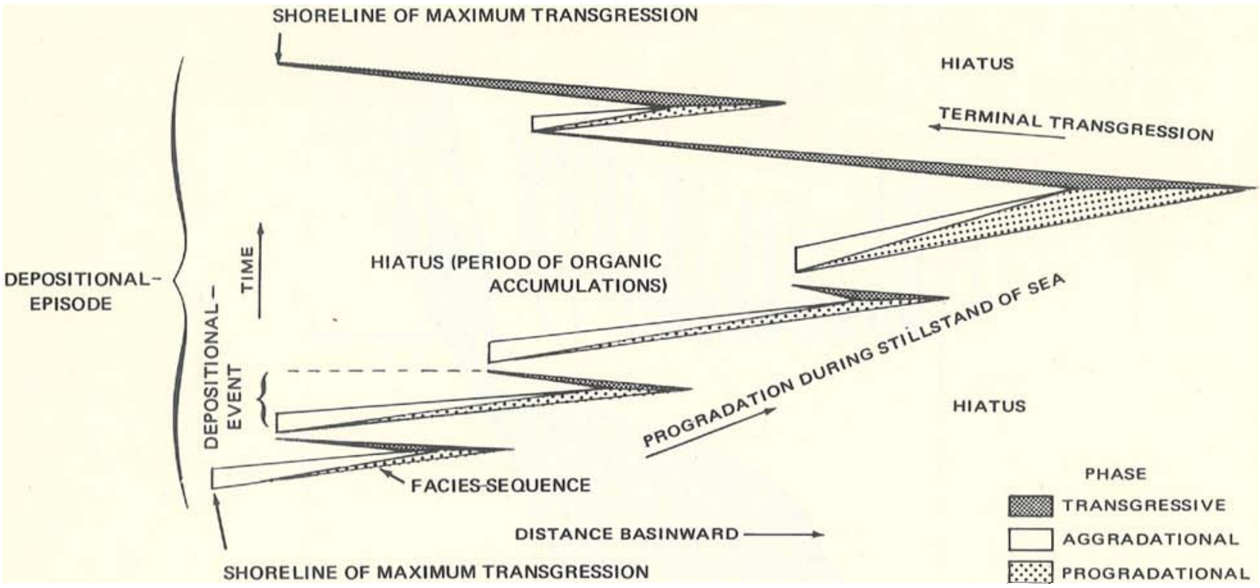

Figure 3.—Depositional-episode.

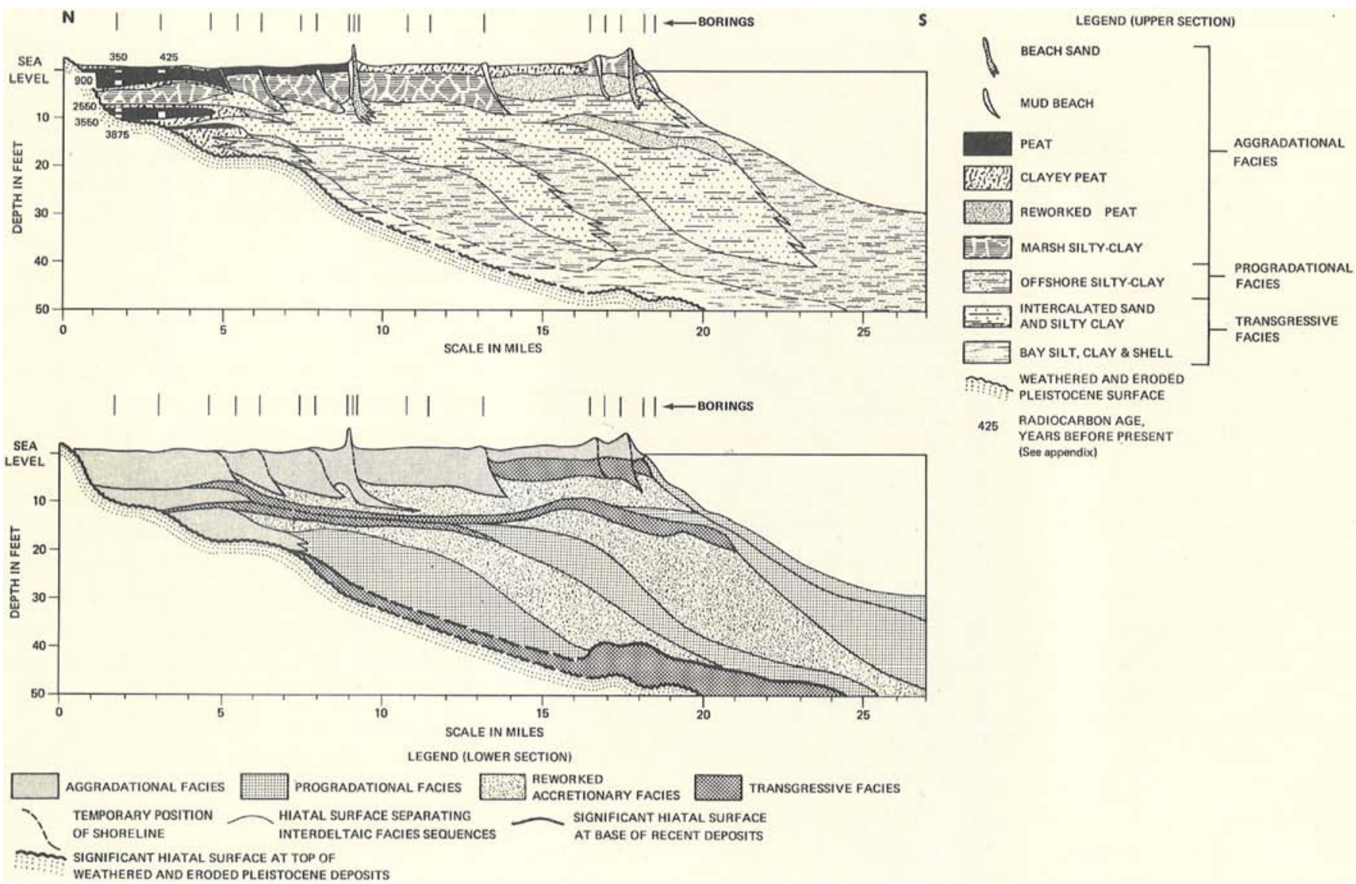

Figure 4.—Interdeltaic offlap: chenier plain.

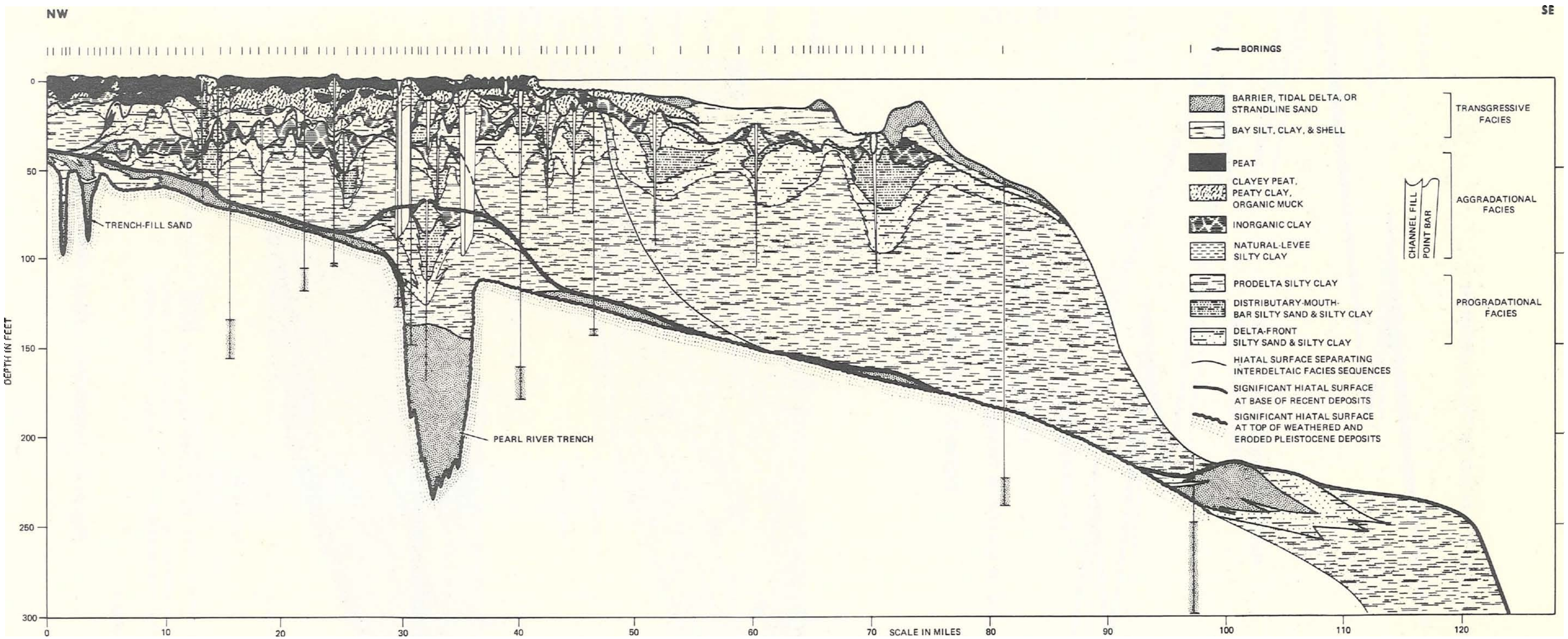

Figure 5.—Deltaic offlap: St. Bernard delta complex.

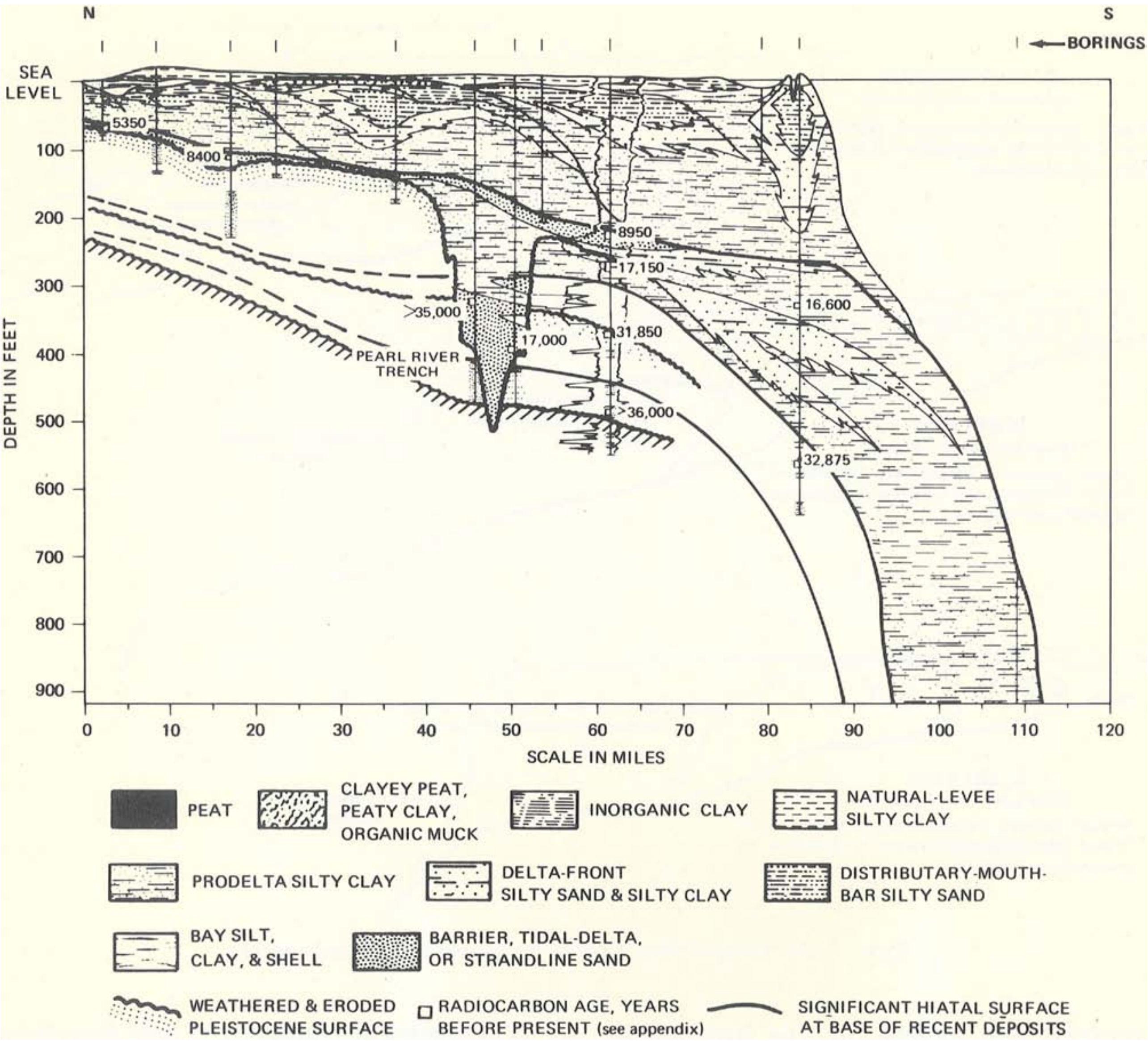

Figure 6.—Deltaic offlap: Plaquemines delta complex.

sources along the basin margin during an interval of sea-level stability. Each depositional-episode is terminated by a major marine transgression which initiates a widespread concurrent transgressive phase of deposition. During the transgression, estuarine conditions are forced on all streams in the transgressed area, and both shoreline deposits and estuarine deposits are laid down. A temporary stillstand during a terminal transgression leads to concurrent progradation along the entire shoreline; the rate of progradation at any given location is, of course, proportional to the distance from a sediment source and the magnitude of that source. Major marine transgressions may be affected by tectonic depression or isostatic subsidence of the basin margin or by a glacio-eustatic rise of sea level. If the transgression is induced by regional tectonism or isostatic adjustment, it too is only regional, and the depositional-episode is terminated only in the affected portion of the basin. After the transgression has reached its maximum landward limit, a new depositional-episode begins along the inundated portion of the basin margin. The construction of a new depositional-complex then proceeds through deposition of a stepwise basinward procession of discrete facies-sequences across the sediments of the older depositional-complex of which it was once a part and which has continued unaffected elsewhere. A depositional-episode, therefore, is not a specified interval of time. Each

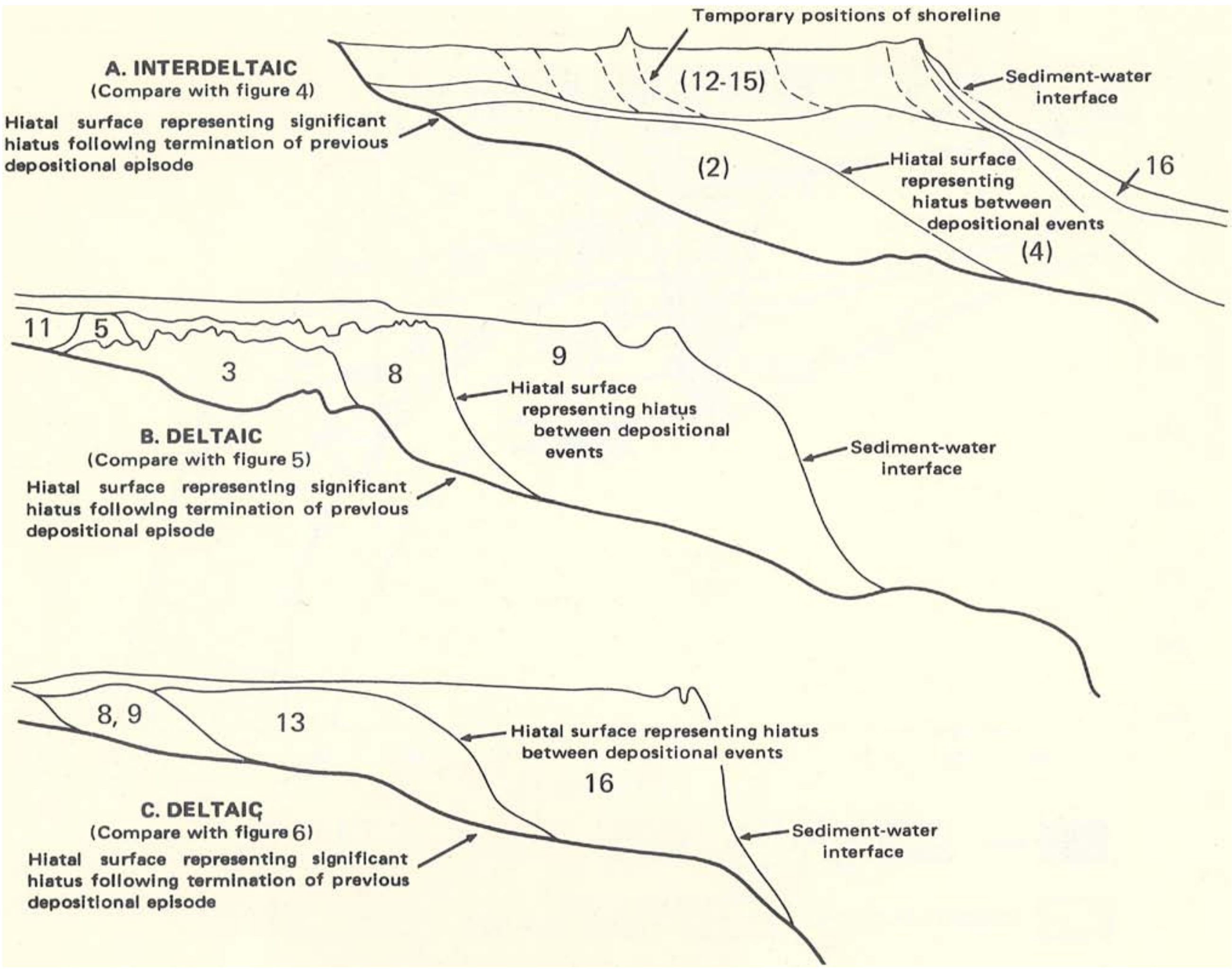

Figure 7.—Stratigraphic framework of the present depositional-complex.

is recorded and defined by the depositional-complex constructed during an interval of sea-level stability. Depositional-complexes may or may not be totally correlative timewise with others within the same basin or those in other basins.

Marine regressions, whether tectonically or isostatically controlled, or glacio-eustatic in nature, move the depositional zone of each river closer to the deeper portion of the basin and increase the competence of all streams by increasing their gradient. The overall effect is an acceleration in the progradational phase of each river's continuing depositional-event. Under these circumstances, the individuality of several facies-sequences may be lost because of their merging, and the multiplicity of sediment sources may be indicated only by local thickenings within a widespread, composite facies-sequence.

The internal configuration of each depositional-complex is dependent on the several discrete facies-sequences of which it is constructed. The minor hiatal surfaces that bound each facies-sequence on figure 7 depict the first-order stratigraphic framework of clastic sediment fill in a basin. Each of the localized hiatal surfaces that separates facies-sequences merges basinward with the widespread hiatal surface that marks the top of the previous depositional-complex and represents the interval of little to no deposition since the end of the previous depositional-episode. [The depositional-complex boundaries are marked by increased numbers of marine organisms due to the lack of sediment dilution.]

Glacially controlled depositional-episodes.—The effects of repetitive glacial cycles best represent the various unstable conditions which regulate clastic deposition along a basin margin. The growth of

Figure 8.—Delta lobes formed by Mississippi River in the past 6,000 years.

	STAGE	RADIOCARBON AGE (YBP)	MATERIAL DATED	DEPTH BELOW PRESENT SEA LEVEL (IN FEET)	SAMPLE LOCATION LAT. N	SAMPLE LOCATION LONG. W	LAB. RUN NO.
WISCONSIN	INTERSTADIAL	35,200±2,400	Beach rock (valves of inner-neritic mollusks and sparry calcite cement)	0 to +3	24°30.0′	97°45 0′	Tx-155
	LATE ALTONIAN ADVANCE	32,500±3,500	Valves of bay pelecypods	48	28°41.2′	97°45 0′	Sh-5578
		31,850±1,800	Valves of inner-neritic pelecypods (immediately beneath transgressive beach sand)	360 to 380	29°16.7′	89°16.3′	L-291L
	FARMDALIAN ADVANCE	29,300±2,000	Valves of inner-neritic pelecypods	147	29°56.4′	90°04.3′	L-291D
		27,000±1,200	Valves of inner-neritic pelecypods	121	29°56.2′	90°05.2′	L-291C
		26,900±1,800	Valves of inner-neritic pelecypods	51	28°50.4′	95°08.0′	Sh-4427
	FARMDALIAN INTERVAL	24,900± 700	Beach rock (valves of inner-neritic mollusks and sparry calcite cement)	0 to +3	24°30.0′	97°45.0′	Tx-156
	IOWAN ADVANCE	24,300±1,400	Valves of bay pelecypods	64	30°15.0′	90°07.5′	0-764
		23,400±1,800	Valves of inner-neritic pelecypods (in shoreface clayey sand)	49 to 51	26°57.8′	97°22.6′	0-630
	INTERSTADIAL	22,440± 800	Beach rock (valves of inner-neritic mollusks and sparry calcite cement)	17 to 18.5	26°58.9′	97°27.4′	0-478
		20,600± 750	Valves of inner-neritic pelecypods (in beach sand)	15.5 to 17.0	26°58.9′	97°27.4′	0-609
	TAZEWELL ADVANCE	19,400± 510	Valves of bay pelecypods	162	28°21.0′	92°49.0′	0-273
		16,940± 680	Valves of inner-neritic mollusks (in beach sand)	288	27°57.1′	95°10.5′	Sh-5576
		16,600± 420	Valves of inner-neritic pelecypods	330-331.8	29°57.9′	89°22.8′	0-1068
		15,575± 500	Valves of inner-neritic pelecypods	348.8-349.8	28°58.5′	89°08.8′	0-1642
	BRADYAN INTERVAL						
	CARY ADVANCE	12,960± 470	Valves of bay pelecypods	189	28°09.4′	94°17.6′	Sh-4894
		11,900± 250	Valves of inner-neritic pelecypods	229	28°57.5′	89°48.8′	0-465
		11,050± 300	Valves of inner-neritic pelecypods	215	29°09.0′	89°59.0′	L-291G
	TWO CREEKIAN INTERVAL	10,700± 220	Valves of inner-neritic mollusks (in beach sand)	167-182	29°39.0′	97°08.3′	0-45
		10,700± 150	Brackish-marsh peat	140	29°12.2′	90°43.8′	L-291X
		10,525± 215	Wood and brackish-marsh peat	115	28°57.6′	93°01.4′	0-1894
	MANKATO ADVANCE	9,250± 210	Wood and brackish-marsh peat	52.5-53.5	29°41.5′	93°19.7′	0-1771
		8,800± 180	Valves of inner-neritic pelecypods	90-100	29°35.2′	90°19.1′	0-73
		8,700± 200	Valves of inner-neritic pelecypods	73	29°57.9′	91°06.1′	L-125G
		8,400± 150	Valves of inner-neritic pelecypods	115.4	29°38.4′	89°57.1′	0-353
		8,150± 180	Valves of bay pelecypods	65.2	29°11.2′	89°53.2′	0-228
POSTGLACIAL		7,240± 160	Brackish-marsh peaty clay	39.5-40.5	29°37.0′	91°32.8′	0-1861
		7,150± 160	Valves of bay pelecypods	49.5	29°01.5′	89°08.8′	0-358
		7,025± 160	Brackish-marsh peat	24-25.5	29°49.2′	91°18.9′	0-1774
		5,650± 140	Brackish-marsh peat	17.8-18.1	29°51.2′	93°03.0′	0-902
		5,600± 140	Valves of bay pelecypods	23	30°12.0′	91°19.3′	0-72
		4,800± 140	Valves of bay pelecypods	20-25	29°59.9′	90°07.8′	0-119
		4,600± 125	Brackish-marsh peat	10	29°39.3′	92°28.2′	0-393
		3,550± 120	Brackish-marsh peat	8.6-9.6	29°43.9′	92°20.2′	0-2214
		3,500± 115	Valves of bay pelecypods	25-29	29°48.8′	90°30.0′	0-1663
		2,550± 110	Brackish-marsh peat	7.5-8.5	29°43.9′	92°20.2′	0-2243
		1,400± 350	Valves of bay pelecypods	12-12.7	29°51.2′	90°01.2′	L-175D
		900± 125	Brackish-marsh peat	1.9-2.9	29°43.9′	92°20.2′	0-2242

Table 1.—Radiocarbon age determinations pertinent to eustatic sea-level fluctuations in the Gulf of Mexico in the past 40,000 years.

continental glaciers leads to a eustatic lowering of sea level and consequent regression; this, of course, is equivalent to a tectonic uplift of the basin margin or to an increase in the volume of the basin while the water volume remains the same. Conversely, the melting of continental glaciers leads to a eustatic rise of sea level and consequent transgression; this, in turn, is equivalent to isostatic or tectonic lowering of the basin margin or to a reduction in the volume of the basin while the water volume remains the same.

The onset and ensuing effects of continental glaciation are depicted in the diagrams of figure 9. The conditions at the end of the Tertiary Period are shown in diagram A. As indicated, progradation had proceeded to the stage at which clastic sediments were deposited directly onto the upper slope. As the regression proceeded in response to the enlargement of continental glaciers (diagram B), the progradational phase of each depositional-event along the basin margin was concurrently accelerated. The regressive deposits became coarser and coarser as the streams, during their adjustment to steepening gradients, eroded first through their previous deltas and later through their upstream meanderbelts. By the time a sea-level low stand was reached (diagram C), the steepened stream gradients had increased the competency of the streams by many times, and depositional rates along the basin margin had increased correspondingly. Although progradation during the low stand further increased the width of the coastal plain, the initial widening had occurred during the lowering of sea level. Within that interval, the deltaic and neritic sediments that had been deposited during the preceding high stand and during the regression were progressively exposed. The subsequent drainage and dessication of these sediments and their consequent shrinkage led to a lowering of the coastal plain, as is shown.

As the climate warmed and the meltwaters from the waning glaciers returned to the sea, a eustatic rise of sea level ensued, and the terminal transgressive phase of the depositional-episode began.

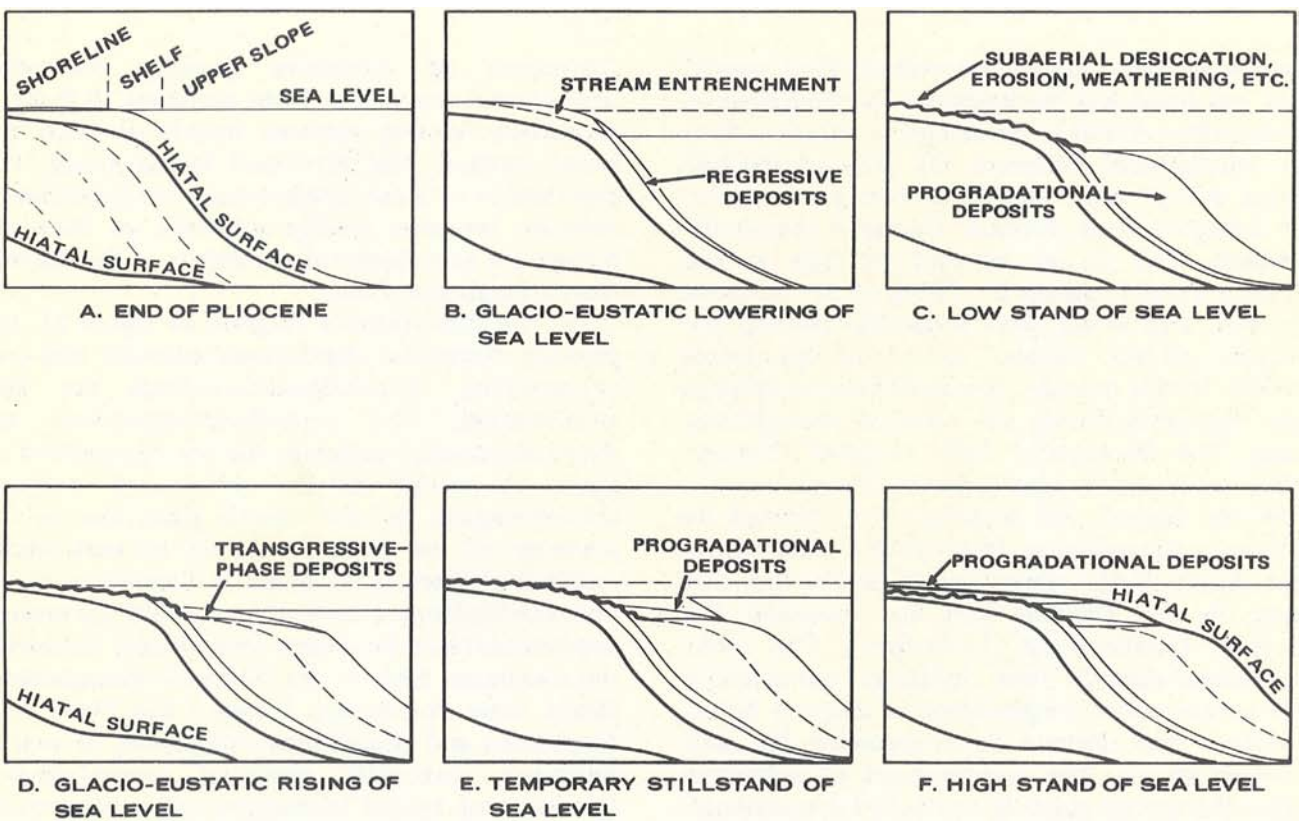

Figure 9.—Deposition during first glacial cycle.

GLACIAL ADVANCE	RADIOCARBON AGE (YBP)	MATERIAL DATED	SAMPLE LOCATION LAT. N	SAMPLE LOCATION LONG. W	LAB. RUN NO.
LATE ALTONIAN	31,800±1,200	Spruce trunk (in till)	42°33′	88°30′	W-638
	30,800±1,000	Spruce log (in outwash sand)	43°03′	88°12′	W-901
FARMDALIAN	29,000±1,000	Wood (in till)	45°56′	92°30′	W-747
	29,000± 900	Spruce log (in outwash gravel	45°45′	88°33′	W-903
	28,700± 800	Peat (beneath till)	46°26′	99°40′	W-1045
"IOWAN" (MORTON LOESS)	25,100± 800	Wood (in loess)	40°40′	89°29′	W-69
	24,600± 800	Wood (in glacial varves above loess)	41°25′	81°34′	W-71
	22,900± 900	Wood (in loess)	40°40′	89°29′	W-68
TAZEWELL OR VASHON	20,500± 800	Wood (in peaty deposit beneath till)	39°45′	87°11′	W-577
	20,500± 600	Moss peat (beneath till)	40°35′	89°16′	W-483
	19,100± 300	Log (in till)	39°25′	84°33′	W-724
	16,100± 850	Log (in till)	40°01′	82°28′	C-893
CARY	13,820± 400	Wood (in forest bed beneath till)	42°02′	94°28′	W-513
	12,200± 500	Wood (in till)	42°04′	93°36′	C-653
	11,952± 500	Wood (in till)	42°04′	93°36′	C-596
MANKATO OR VALDERS	10,856± 410	Spruce log with bark (in till)	44°17′	88°28′	C-800
	10,676± 750	Tree stump (in glacially dammed lake clays)	44°16′	88°20′	C-630

Table 2.—Radiocarbon age determinations pertinent to glacial advances in North America in the past 40,000 years.

The transgression across the near-sea-level coastal plain was rapid, but the landward encroachment of the shoreline slowed as the steeper oxidized surface was encountered (diagram D). The entrenched stream valleys were inundated during this phase, and estuaries were formed. During a temporary stillstand, due to an intermittent halt in the warming trend (diagram E), filling of the estuaries occurred, and deltas were prograded beyond the confines of the alluvial valleys of the larger streams. In this manner, complete facies-sequences were deposited during the terminal transgressive phase. The Maringouin delta complex (Frazier, 1967) was formed during such a stillstand.

At the instant the shoreline had reached its maximum transgression (diagram F), there was a termination of the depositional-episode that had begun in the Pliocene and had breached the Tertiary-Quaternary boundary. The next depositional-episode was initiated immediately with a concurrent progradation of discrete facies-sequences—one discrete facies-sequence for each sediment source. The emplacement of sediments during the second glacially controlled depositional-episode is diagrammed on figure 10, which is presented specifically to show the offlapping configuration of successive glacially controlled depositional-complexes. Note on figure 10 that the subaerially eroded surfaces interfinger with the hiatal surfaces that developed subaqueously. The significance of these subaqueously developed hiatal surfaces becomes readily apparent as they are recognized as widespread separators of the discrete depositional-complexes.

On the time-distance diagram of figure 11, the glacially controlled depositional-episodes and corresponding depositional-complexes are also represented. The depositional-complexes are diagrammatically similar to the one represented on figure 3, except for the consequent erosional unconformities on the coastal plain due to the lowering of sea level. It should be particularly noted that there is no break in deposition at the Pliocene-Pleistocene boundary and that the present depositional-episode began immediately following the maximum limit of the Holocene transgression. Direct time correlation between the Pleistocene formations and depositional-complexes, as well as between subaerial erosional unconformities (glacial) that bound formations and subaqueously developed hiatal surfaces (interglacial) that bound the depositional-complexes, is impossible.

WARM-INTERVAL FORAMINIFERAL ZONE	RADIOCARBON AGE (YBP)	MATERIAL DATED	SAMPLE LOCATION		LAB. RUN. NO.
			LAT. N	LONG. W.	
UPPERMOST ZONE OF *GLOBOROTALIA MENARDII FLEXUOSA*	28,200±2,400	Foram tests	26°53′	92°17′	I-3174
	20,500±1,000	Foram tests	23°40′	92°34′	I-3490
UPPERMOST ZONE OF *GLOBOROTALIA MENARDII MENARDII*	10,430±570	Foram tests	23°40′	92°34′	I-3495
	Modern—presently living in northwestern Gulf				

Table 3.—Radiocarbon age determinations pertinent to warming trends in the Gulf of Mexico in the past 30,000 years.

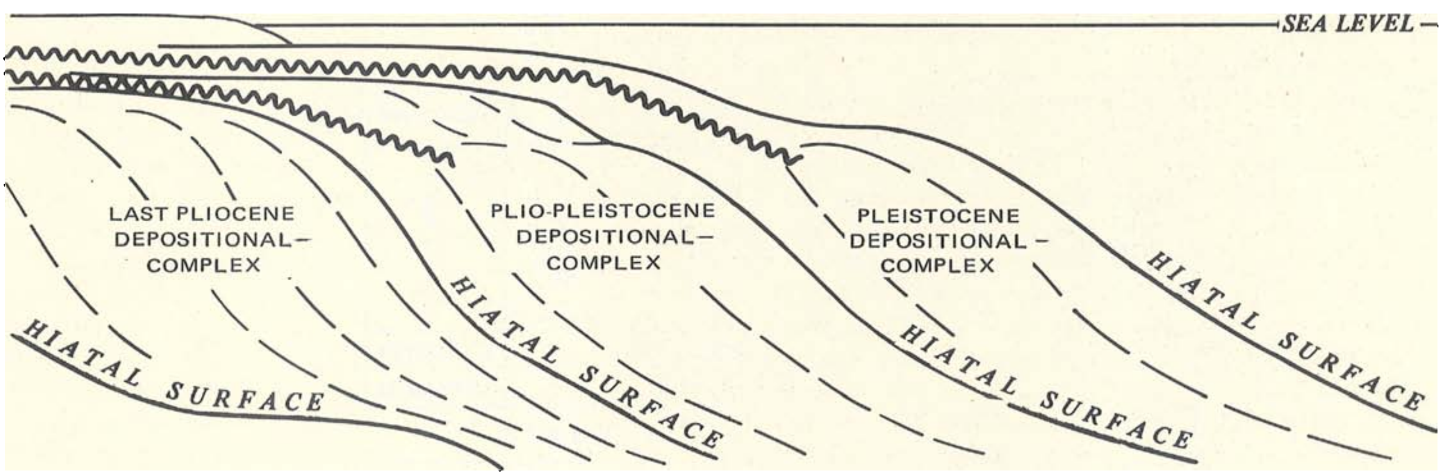

Figure 10.—Succession of offlapping depositional-complexes.

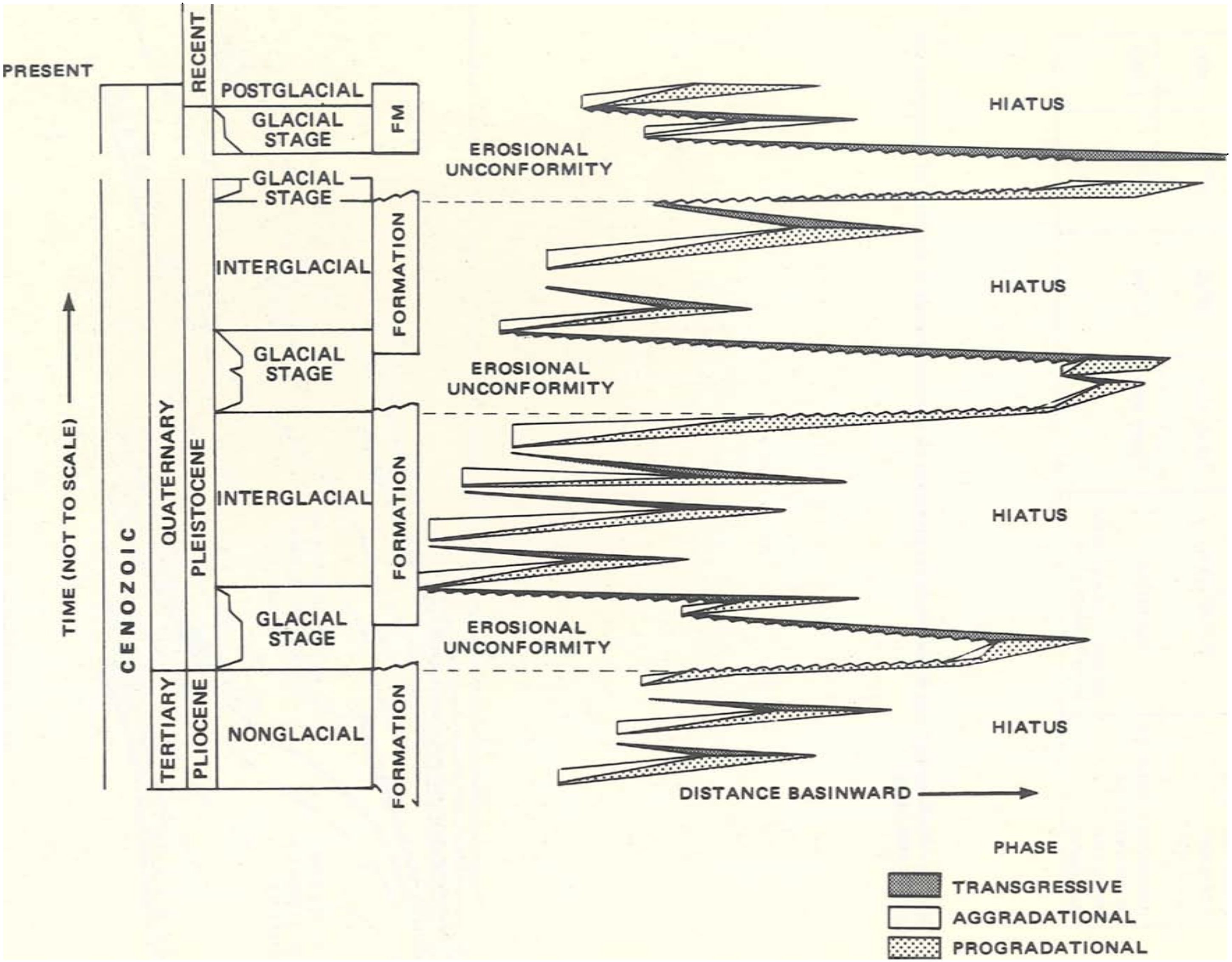

Figure 11.—Glacially controlled depositional-episodes.

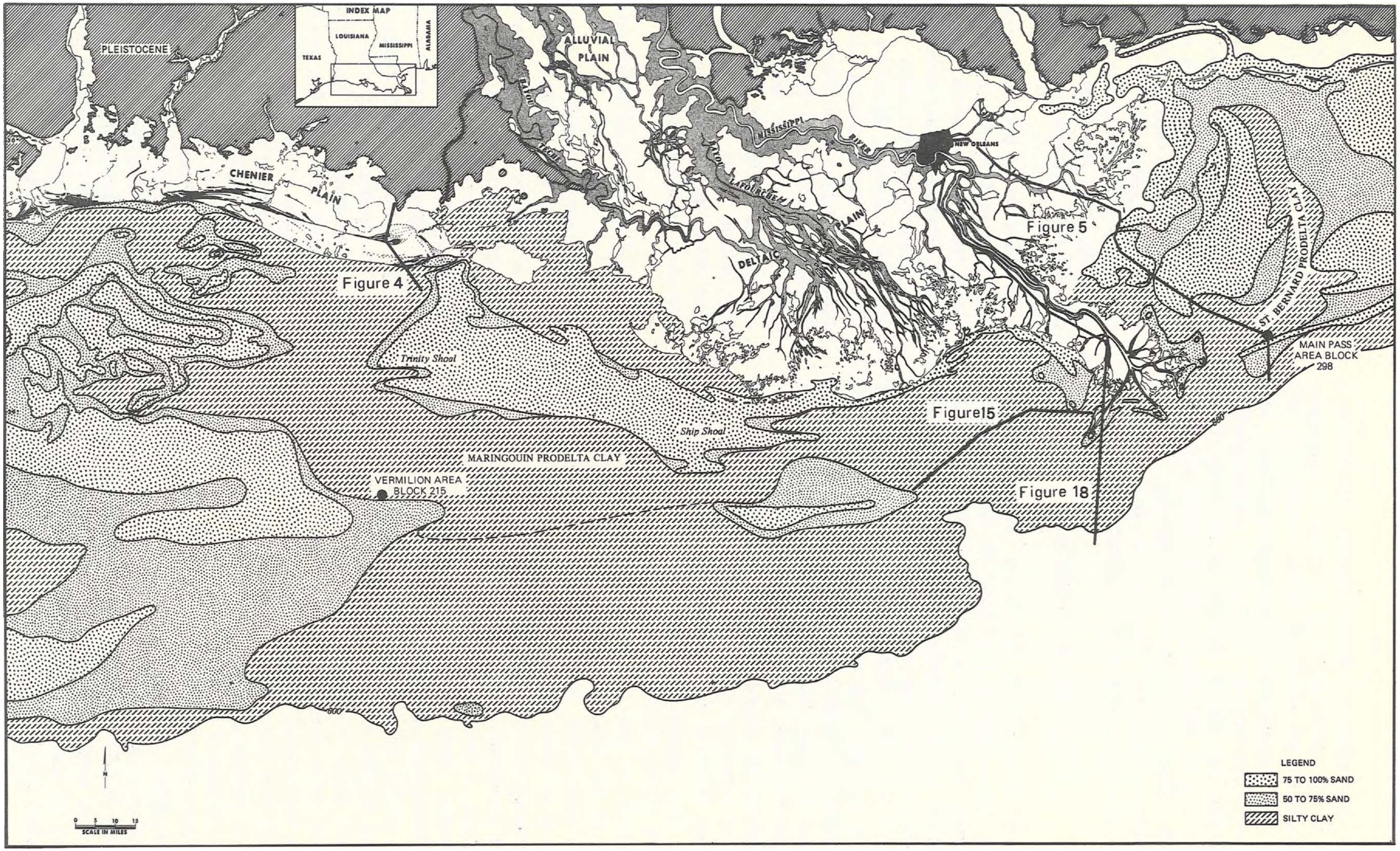

Figure 12.—Sediment distribution on the Gulf floor in the vicinity of the Mississippi delta.

The deposition of clastic silicate particles in a basin is the culmination of a series of processes. The particles are products of the weathering and erosion of continental rocks and sediments, and eventually the majority of them are transported by rivers into the marine portion of the basin. From the vicinity of each stream mouth, these sediments may be immediately reworked laterally to form interdeltaic nearshore deposits, or they may be buried and incorporated into a deltaic mass.

River-Mouth Deposition

There is a natural sorting of particles as the stream debouches its load into the standing body of water. The zone of sand deposition is in the proximity of the stream mouth; the zone of silt deposition from suspension extends slightly basinward of the sands, and that of the clays extends beyond the silts. The zone of clay deposition does not extend indefinitely basinward; it also terminates. Beyond the clay zone very little to no deposition occurs, and a hiatal surface exists.

Evidence to support the hypothesis of a finite zone of prodelta clay deposition is found in the offshore area in the vicinity of the Mississippi delta. The present sediment distribution on that portion of the Gulf floor is shown on figure 12. To the south of the deltaic plain, the transgressive sands of the Maringouin delta complex (Frazier, 1967) form both Ship Shoal and Trinity Shoal. Seaward of these sands are the prodelta clays of the Maringouin complex which terminate abruptly against a deposit of intercalated sands and clays to the south. A boring in Vermilion Area Block 215 penetrated these intercalated sediments beneath the Maringouin prodelta deposits and documents their partial burial by the Maringouin progradations. Similar evidence is found to the east of the birdfoot delta where the prodelta clays of the St. Bernard delta complex have a similar relationship with a well-sorted, calcite-cemented sand. A boring in Main Pass Area Block 298 penetrated the sand beneath 22 feet of the prodelta clay as is shown at the southeastern end of the cross section of figure 5. Located at mile 100 is the abrupt termination of the prodelta clays, which marks the basinward limit of clay deposition during the progradation of the final St. Bernard delta lobe. The surface of the calcareous sand unit, which is exposed beyond, is a portion of the depositional surface across which the lobes of the St. Bernard delta complex prograded; it is also a hiatal surface which represents a progressively longer duration of nondeposition basinward. This hiatus is still in effect basinward of mile 100.

The facies framework shown on the section is also very informative. In the vicinity of each relict distributary channel (most noticeable at miles 52, 60, and 70), the natural size sorting of the clastic particles is apparent. The coarsest grained sediments are adjacent to the once-active distributary channels, and the facies are progressively finer grained with increasing distance away from these channels.

In addition to the hiatal surface at the base of these deltaic deposits, another hiatal surface exists along the top (between miles 0 and 42). Little clastic deposition has occurred in this area since the peats began to form 1,000 years ago. For practical purposes, therefore, the subaerial deltaic plain also represents a depositional hiatus. A similar hiatus occurred during the formation of the Pleistocene soil zone (located beneath the transgressive sands) as sea level lowered in response to continental glaciation. Both the peat deposits on the deltaic plain and the subaerially weathered zone along the Pleistocene surface suggest a longer period of nondeposition in a landward direction.

Initial Deposition of Sediments from Suspension

Initial emplacement of suspended sediment in marine water deeper than 60 feet is less common than is believed; for this reason, initial deposition from suspension merits discussion. Although it is common knowledge that plumes of muddy water flow from the stream mouths along the Gulf Coast, it is not widely known that these plumes remain as discrete units for great distances and do not mix with the underlying marine water. Separate and distinct plumes from the Mississippi River have been traced westward for several tens of miles into water depths of less than 60 feet (Geyer, 1950). Unless the plume's density equals or exceeds 1.02, mixing is negligible and the plume of river water is buoyed upward by denser saline water (Lipsey, 1919), as is shown in figure 13. This condition is also reported by Ouellette (1969). Measurements of the suspended sediment load of the Mississippi River, even during flood stages, have shown concentrations only a tenth as great as those required for a density of 1.02, and similar concentrations have been measured for the Colorado and Brazos

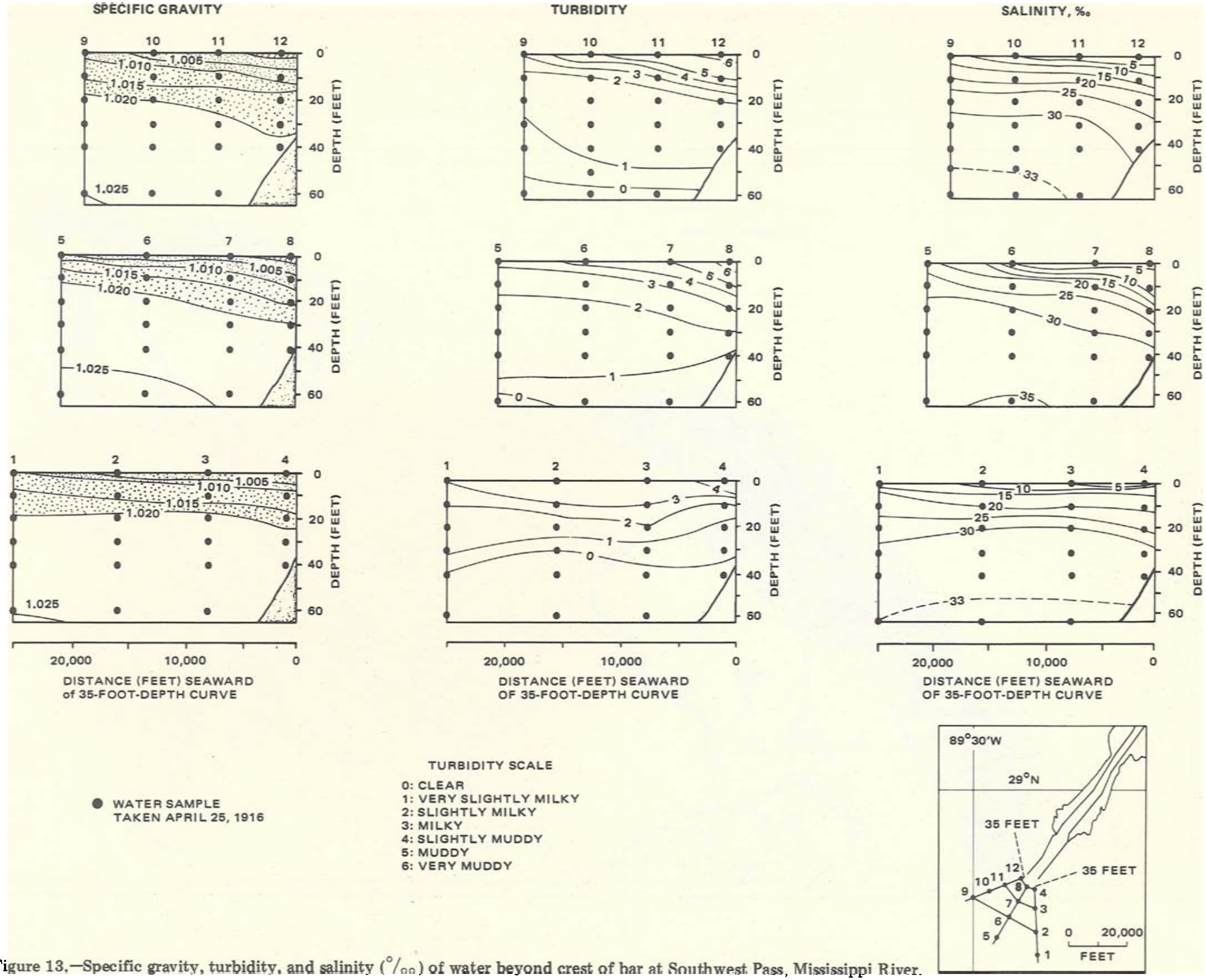

Figure 13.—Specific gravity, turbidity, and salinity (°/₀₀) of water beyond crest of bar at Southwest Pass, Mississippi River.

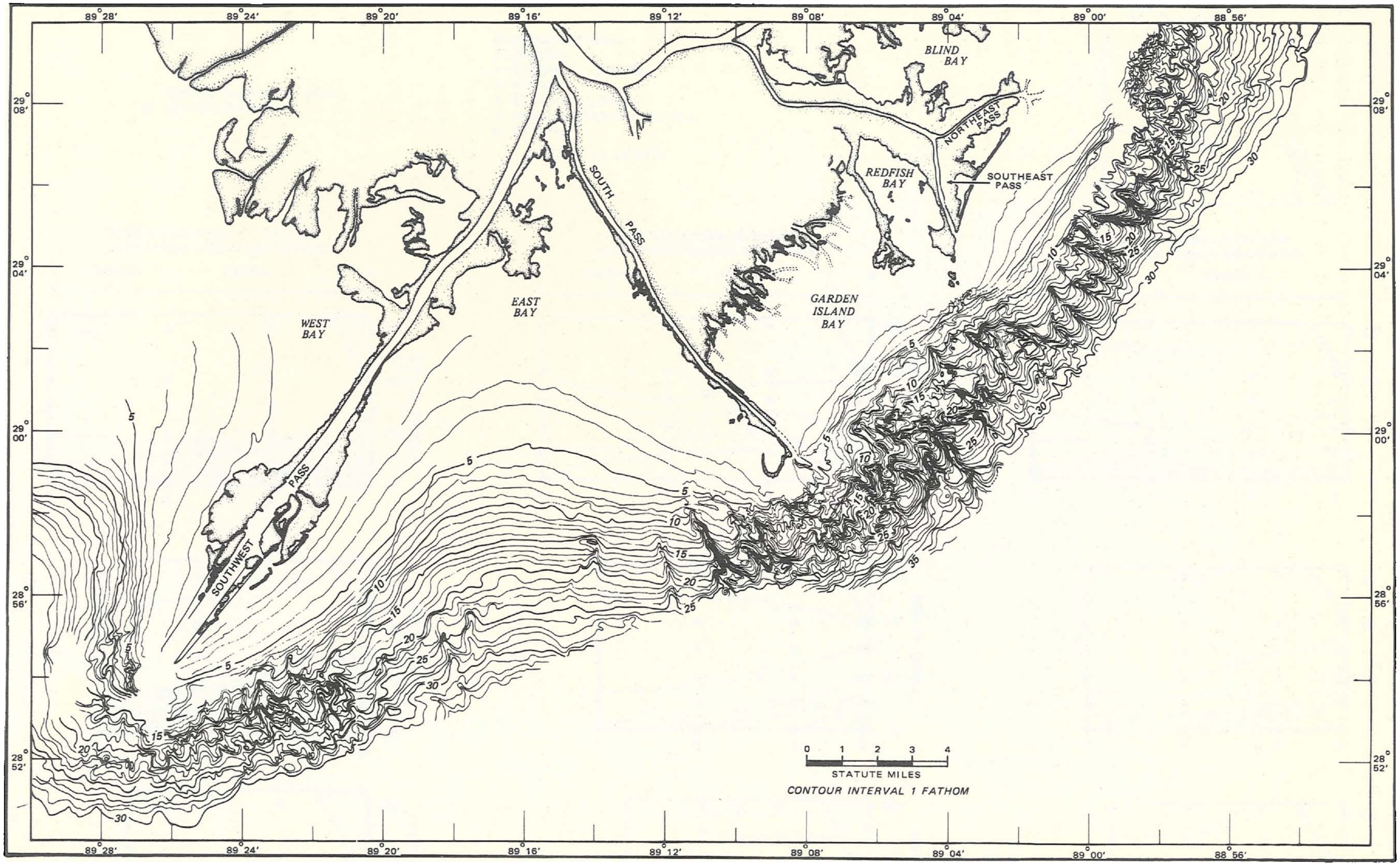

Figure 14.—Submarine topography of the outer shelf off Mississippi birdfoot delta.

Rivers of Texas (Bates, 1953). These data imply that unless a clearly defined river plume comes in contact with the sediment-water interface, no deposition of suspended sediment will occur, for the suspended sediment will remain in the buoyant plume. The distribution of modern marine clays, therefore, is restricted mainly to the nearshore zone around the Gulf, landward of the 60-foot contour where the plumes are driven by surface currents and the suspended sediment in the plumes comes in contact with the bottom and is deposited. In the area of the modern Mississippi birdfoot delta, silty clays of the prodelta facies extend to 380 feet below sea level. The configuration of the delta platform (fig. 14), however, strongly suggests that slumping has occurred below the 48-foot contour as reported by Shepard (1955) and later supported by Terzaghi (1956). The silty clays beyond the zone of contact between the river plumes and the sediment-water interface, therefore, may very well be the result of secondary emplacement by "en masse" creep or slump and not the result of initial deposition from suspension. Indeed, the configuration of prodelta clays as seen on the precision-depth-recorder profile on figure 15 tends to support these contentions. On this section, it is apparent that the relict structure on the left has not received any modern deposition and that a distinct distal boundary exists for the finite zone of sediment deposition from Southwest Pass on the right. Also evident are the modern slump deposits at the base of the delta platform. The basinward limit of the present finite zone of deposition, which is the boundary between Recent and relict sediments, has been determined by analysis of hundreds of surface-sediment samples and is shown on figure 16.

Interdeltaic Deposition

The second facies-sequence cross section (fig. 4) is through interdeltaic nearshore deposits located to the west of the Mississippi River's deltaic progradations (fig. 12). The sediments that formed these interdeltaic deposits were derived from the deltaic deposits to the east and were transported westward along the coast to this location. The silty clays of the interdeltaic offshore and marsh facies are suspended-sediment deposits and are contemporaneous with the prodelta silty clays; they were transported westward by nearshore surface currents during the deltaic progradations. In contrast, the interdeltaic sandy facies were derived from the delta-front sands during destructional transgressive phases; they were transported westward in the longshore drift. The repetitive interdeltaic facies-sequences, therefore, reflect the repetitive progradations of the delta lobes to the east from which the sediments were derived. Their progradational configuration across the hiatal surface at the top of the underlying transgressive deposits is similar to that of the deltaic deposits in figure 5—discrete facies-sequences separated by minor hiatal surfaces. The underlying transgressive deposits in this region overlie the hiatal surface represented by the Pleistocene soil zone as they do in the deltaic region.

The interdeltaic nearshore deposits represent an area of minor sediment influx compared with that of the deltaic region; consequently, progradation of the interdeltaic shoreline is slower than that of the delta. This shoreline configuration is very apparent on the map of the region shown on figure 12.

Major Transgressions and Significant Hiatal Surfaces

Figure 17 shows the surficial sediment distribution on the entire northwestern continental shelf. The sands are relict shoreline deposits which were emplaced during temporary stillstands of the sea and were subsequently transgressed as sea level continued to rise in response to the melting of the last continental ice sheets. The sands are relatively thin with a maximum thickness of a few tens of feet; they contain shells of surf-zone and inner-neritic mollusks, and they are spread landward over silty clays which contain articulated shells of bay pelecypods. Three levels of temporary stillstand are evident; however, there are in reality four. The outermost limit of the relict shoreline deposits is between 280 and 300 feet (48 fathoms) below present sea level; these sands continue landward to the location of another stillstand at present water depths of approximately 174 feet (29 fathoms). The 48-fathom shoreline represents the low stand of the sea during the late Wisconsin continental glaciation. Another higher relict shoreline deposit lies between 130 and 150 feet (23 fathoms) below present sea level, and another is found between 45 and 60 feet (9 fathoms) below present sea level. The present shoreline marks the limit of progradation since the latest stillstand began approximately 2,500 years ago following the last transgression.

The lowermost of these beach sands (48 fathoms below present sea level) consists of coarser sand

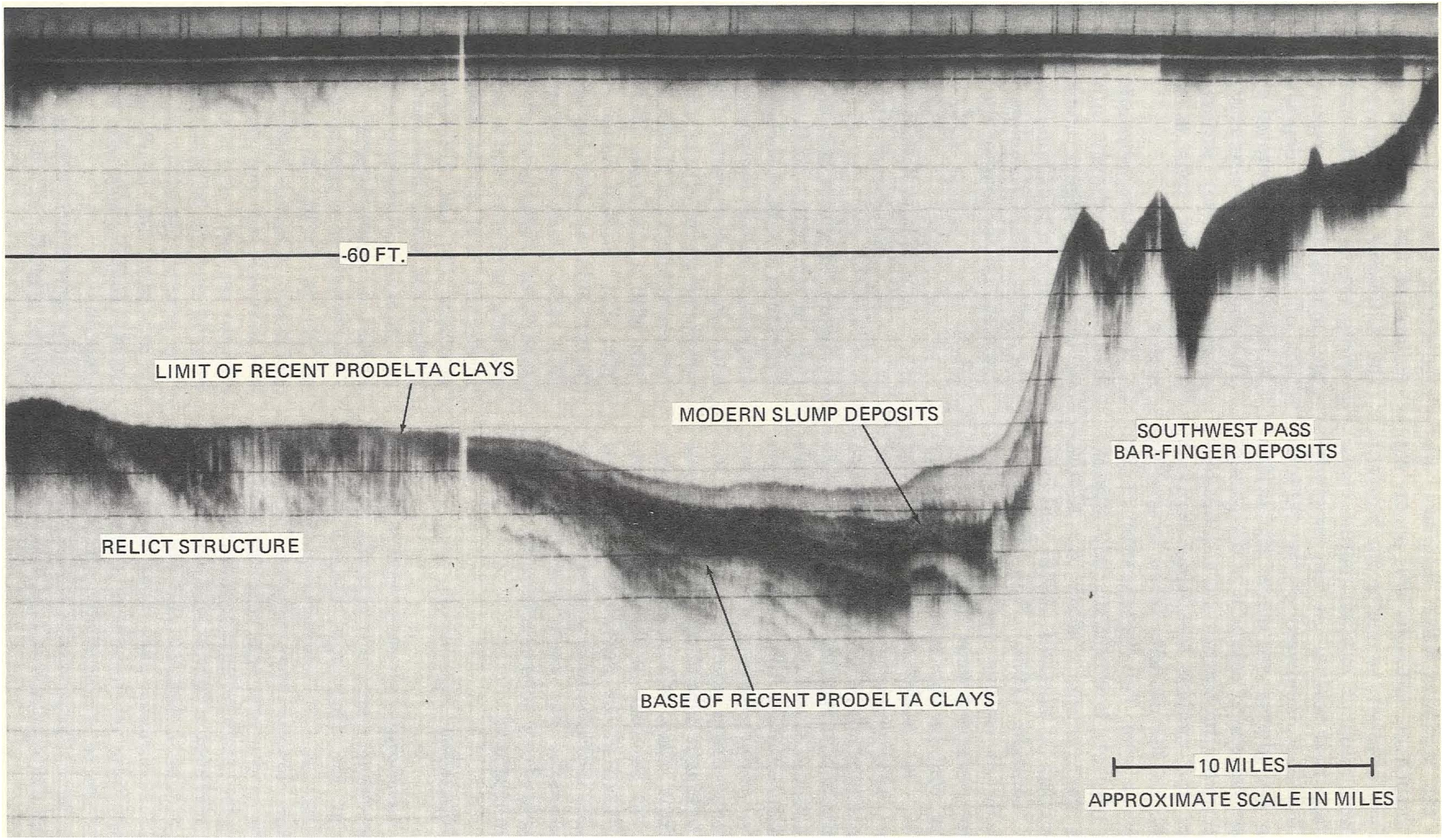

Figure 15.—Precision-depth-recorder section from Southwest Pass of the Mississippi River westward revealing base and lateral extent of Recent prodelta silty clay.

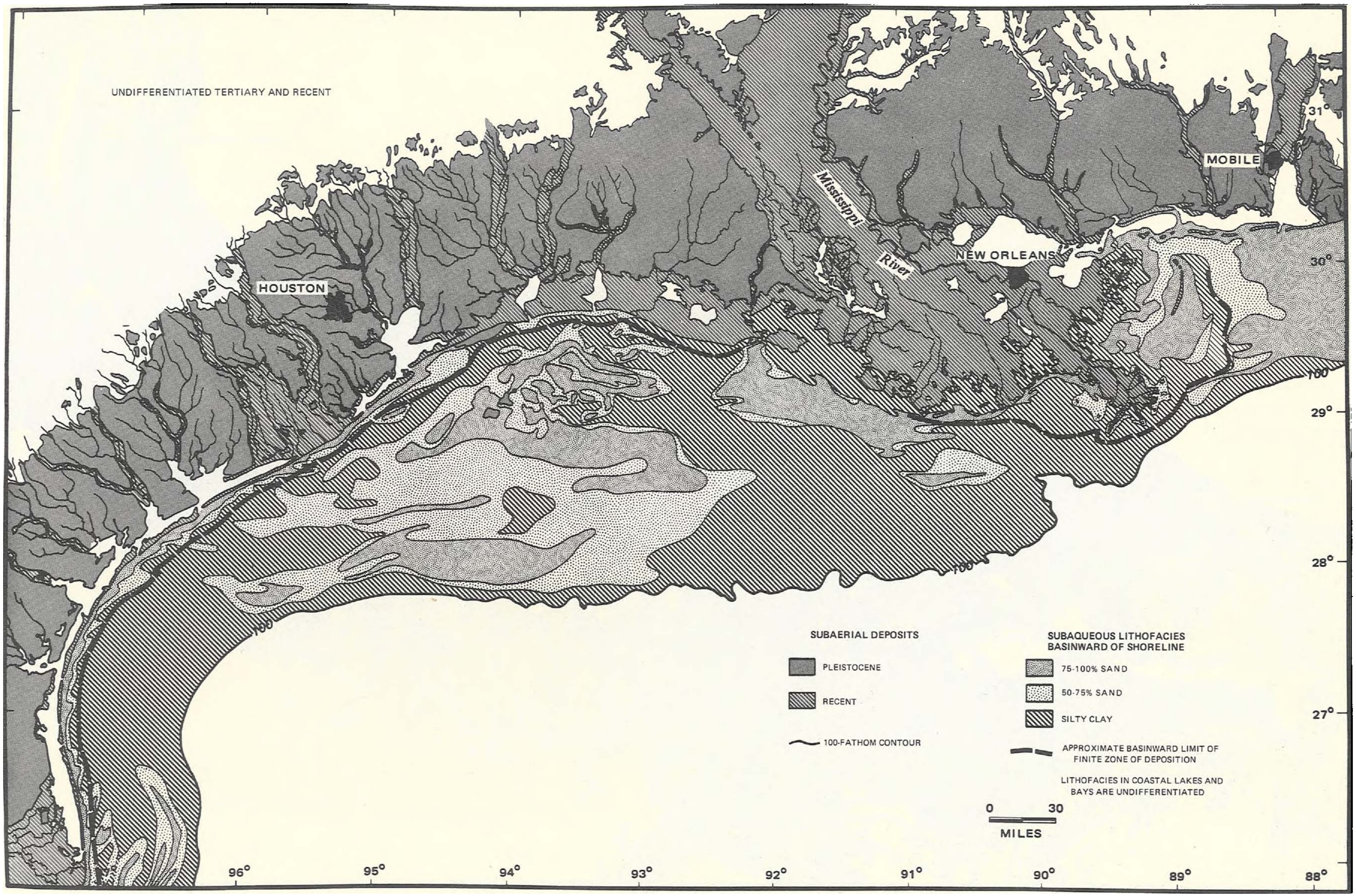

Figure 16.—Basinward limit of Recent clastic deposition on the continental shelf.

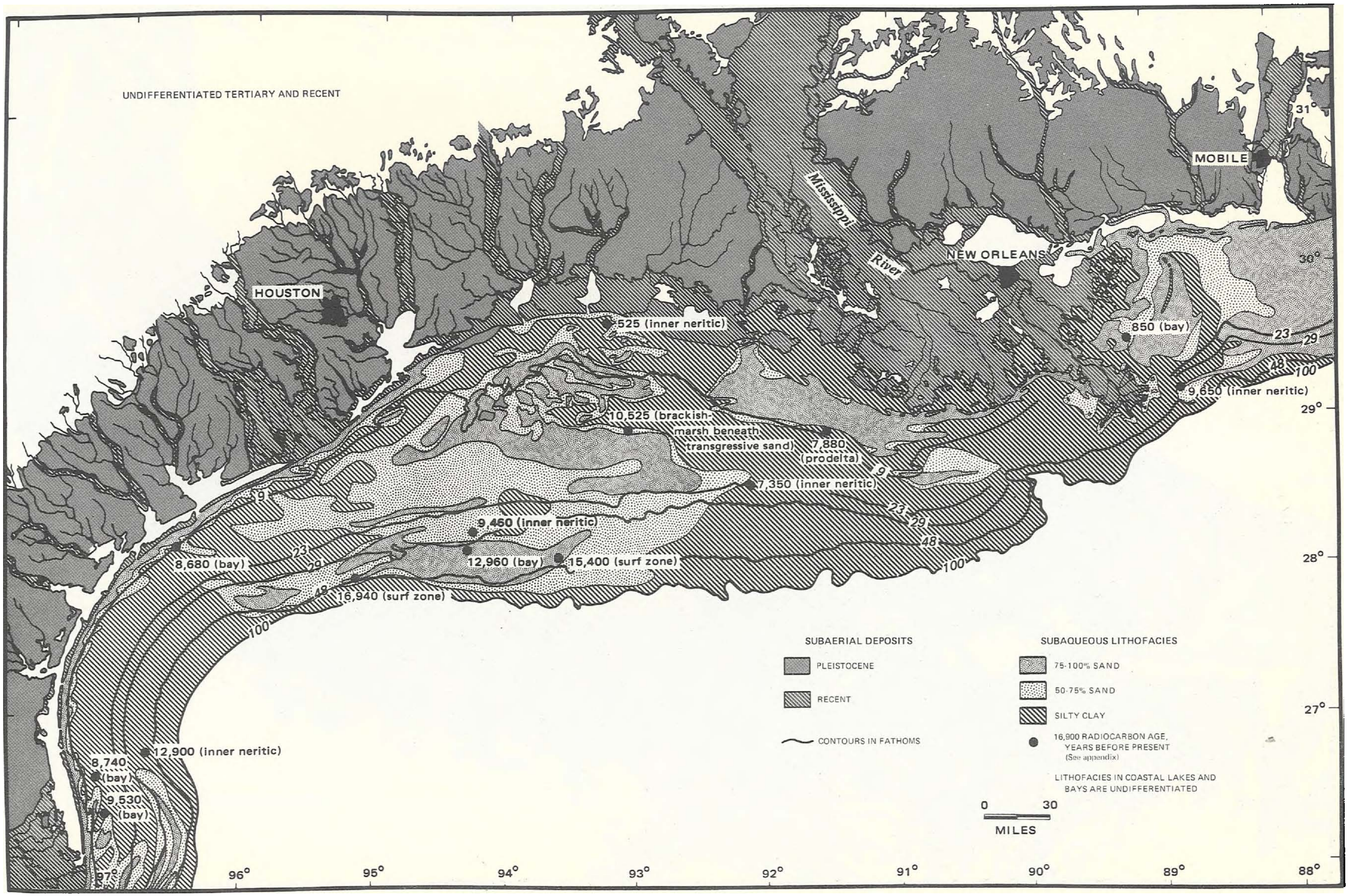

Figure 17.—Relict, transgressive, shoreline sands exposed on the continental shelf, northwestern Gulf of Mexico.

grains than those which were deposited during higher stands of the sea. These coarser sands reflect the greater competency of the steeper gradient streams during the glacially induced low stand of the sea. The median diameters of the 48-fathom beach sands range from 190 to 120 microns; median diameters of the 23-fathom beach sands range from 135 to 84 microns; and those at 9 fathoms range from 109 to 99 microns (Stetson, 1953). Modern beach sands along the northwestern Gulf Coast range in median diameter from 100 to 70 microns.

The successively higher sea-level stillstands which led to the development of these shoreline sands occurred contemporaneously with temporary glacial advances during the discontinuous retreat of the late Wisconsin continental ice sheet. The radiocarbon ages shown on figure 17 were determined from carbon-14 analyses of valves of inner-neritic and bay-type mollusks which were living during the transgression; these fossils were obtained from the surficial deposits, which they consequently date. The oldest shells dated are of inner-neritic, surf-zone mollusks and are approximately 17,000 years old; they mark the most basinward regression of the shoreline during the late Wisconsin glaciation. The remaining radiocarbon ages are progressively younger as the present coast is approached. These dates record the landward progression of the shoreline as sea level rose in response to the melting of the last continental ice sheet.

The sea-level curve for the Gulf of Mexico (fig. 18) is based on radiocarbon dates (appendix) of brackish-water peats and shells of surf-zone pelecypods, supported by dated shells of bay pelecypods and inner-neritic mollusks. The intervals of glacial advance are based on radiocarbon

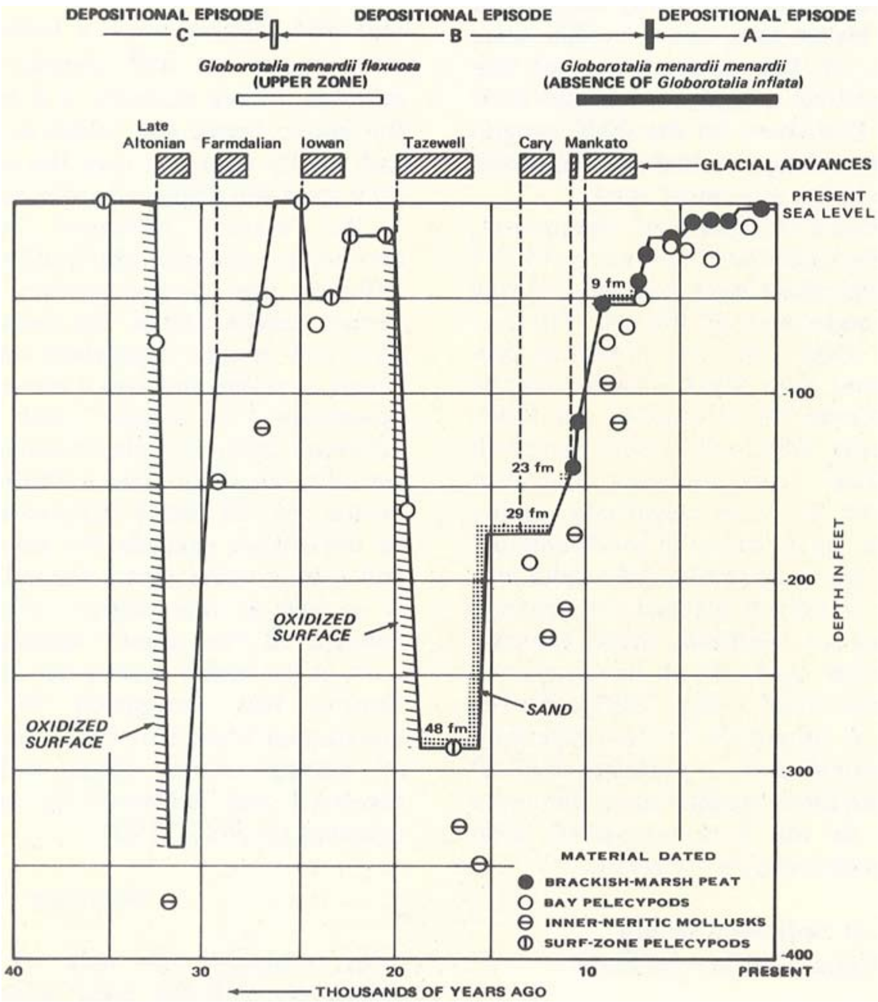

Figure 18.—Sea level during last 40,000 years—Gulf of Mexico.

dates of peat or tree stumps underlying glacial tills and of logs incorporated in the tills. The correlation between continental glaciation and changes in sea level is striking; note the stillstands at 48 fathoms (18,500 to 15,500 BP), 29 fathoms (13,500 to 12,000 BP), 23 fathoms (11,000 to 10,500 BP), and 9 fathoms (10,000 to 7,500 BP). These data add further evidence that the surface sands on the continental shelf are relict shoreline deposits which formed during times of temporary stillstands of the sea (figs. 17 and 18).

Not only do these relict shoreline sands record the landward progression of the last eustatic rise in sea level with its successively higher stillstands, but their age records the duration of the hiatus within each of the different paralic zones on the continental shelf. Wherever these beach and associated facies are exposed on the shelf, no deposition has occurred since their formation. This hiatus has existed for approximately 17,000 years over the major portion of the outer shelf; only seaward of Louisiana has the hiatus been ended in this outer zone as a result of the progradation of the Mississippi River's deltaic complex across the shelf (figs. 16 and 17). Elsewhere on the shelf, progradation is only now ending the hiatus of approximately 8,000 years in the innermost zone.

Progradation across a hiatal surface.—At present, most rivers along the Gulf Coast have not yet filled their estuarine valleys which were inundated during the postglacial transgression of the sea. The few rivers of the Gulf Coast that are presently prograding deltas beyond their filled estuaries are the Rio Grande, the Brazos, the Mississippi, the Pearl, and the Appalachicola. Eventually, other rivers will prograde their deltas across the shelf one after another in the order of their magnitude, as the Mississippi has done. As is evident in the Mississippi birdfoot delta (fig. 6), progradation of a delta into deep water results in a thick platform of prodelta silty clay that encloses localized, thick, elongate distributary-mouth-bar sands which have subsided into the clayey mass (Fisk, 1955, 1961). By the time the shoreline is prograded to the outer shelf margin, several hundred feet of prodelta clay will have lapped onto the hiatal surface along the upper continental slope. As this is accomplished, sediment failure (slump or creep) is expected.

Effects of Sediment Failure Along the Upper Continental Slope

Although sediment failure has occurred along the upper portion of the modern delta platform of the Mississippi River, it is not of the magnitude that would be expected if the Mississippi had already prograded to the upper-slope environment. Modern sediment failures on the slope cannot be documented in the Gulf of Mexico, and documentation must be obtained elsewhere.

Turbidites.—Several sediment failures have occurred off the mouth of the Magdalena River of northern Colombia, which presently is the only river debouching considerable amounts of sediment directly onto the upper continental slope. Cable breaks downslope from the Magdalena delta are related to peak river discharge (Heezen, 1959) and consequent maximum sediment loading of the unstable delta platform. At times of sediment failure, man-made jetties at the mouth of the Magdalena River have disappeared. Because Recent turbidite sands have been cored from the abyssal plain at the base of the continental slope to the north of the delta, it is assumed that the sudden sediment failures place a large mass of sediment into suspension and thereby initiate turbidity currents. These currents, it is believed, rush down the slope, break the cables as they engulf them, and finally dissipate over the abyssal plain, where they spew the displaced sediment.

"En masse" sediment transport.—There is another mechanism which effects displacement of sediment into deeper portions of the basin; this second mechanism is "en masse" sediment transport and is also dependent on sediment failure. Creep or slump involves a movement of sediments downslope "en masse," and both are entirely different from the displacement of sediments by turbidity currents. The initiating sediment failure occurs only in clayey sediments wherever the rate of deposition exceeds the rate of consolidation, and it may occur where the sediment surface slope is as low as one degree (Terzaghi, 1956). The concept of "en masse" submarine sliding is not new; in the earlier part of the century, geologists in Europe had recognized the results of this mechanism in ancient rocks. An excellent account of Silurian rocks which had been repeatedly displaced and deformed by submarine sliding is reported by Jones (1937).

Summary

In summary, the bulk of clastic deposition occurs in a finite zone adjacent to the basin margin, bulging basinward in the vicinity of stream mouths and narrowing landward between them.

The zone slowly shifts basinward during progradations and landward during transgressions. The sediments deposited cover hiatal surfaces as the zone of deposition moves—either in response to eustatic sea-level changes or in response to vertical displacement of the depositional surface in the vicinity of the shoreline.

Progradation to the upper-slope margin leads to sporadic deposition in the deeper portions of the basin. Sediment failures, generated by depositional loading along the upper slope, propagate either "en masse" sediment creep or slump or turbidity currents which effect the transport and secondary emplacement of sediment onto the lower slope and abyssal plain.

STRATIGRAPHIC FRAMEWORK

Questions such as "Where should the base of the Recent sediments be placed?" will be answered differently by those utilizing only terrestrial unconformities and by those studying the overall stratigraphic framework imposed by repetitive depositional-episodes.

Rock-Stratigraphic Units

An understanding of the concepts discussed above leads to an understanding of the stratigraphic framework that is constructed by the processional emplacement of discrete clastic sedimentary units during basin filling. Two ranks of units exist: the facies-sequence laid down during a depositional-event, and the composite depositional-complex constructed of several facies-sequences during a depositional-episode.

According to the 1961 Stratigraphic Code (Article 26), time-stratigraphic units such as a depositional-event and the higher order depositional-episode, should . . . "depend fundamentally for definition on an actual section or sequence of rock . . . [and] record an interval of time that extended from the beginning to the ending of its deposition . . . ." According to Article 27 of the Code, "boundaries may be based on any features thought to be stratigraphically useful . . . they should set the unit apart as representing a significant geologic episode."

The suggested position of the newly proposed units is tabulated below in the stratigraphic hierarchy.

Time-Stratigraphic Units	Rock-Stratigraphic Units
System	
Series	
Stage	
Depositional-Episode	Depositional-Complex
Depositional-Event	Facies-Sequence

Relationship of Depositional-Complexes to Formations

Depositional-episodes extend from the end of one maximum transgression through the succeeding intervals of stability and instability of sea level to the end of the next maximum transgression; the intervals of time during which formations are deposited extend from the end of one maximum progradation through the succeeding intervals of instability and stability of sea level to the end of the next maximum progradation. The interval of time during which stability exists and progradation occurs is common to both stratigraphic units, as are the progradational and aggradational sediments that are deposited during that time interval. The significant difference between formations and depositional-complexes is that the transgressive-phase deposits form the uppermost facies of a depositional-complex, to which they are genetically related, but form the basal unit of a formation, to which they are not genetically related. Their relationship is readily apparent on figures 9, 10, and 11.

The landward limits and exposed progradational and aggradational sediments of both depositional-complexes and formations coincide on the subaerial portion of the present coastal plain (fig. 1). These landward limits are shorelines of maximum transgression and are the turning points of the finite zone of deposition from the terminal transgressive phase of one depositional-episode to the initial progradational phase of the next. Basinward of these common contacts, however, the boundaries of the depositional-complexes and formations are separated by transgressive-phase deposits. The relict transgressive-phase sands and silty clays exposed on the continental shelf (fig. 1) directly overlie the oxidized surface of the Pleistocene Prairie Formation. These same relict transgressive deposits, however, directly underlie the hiatal surface of the preceding depositional-complex.

The basinward limit of the Recent depositional-complex is the basinward limit of the finite zone of deposition shown on figure 16 where sediments are being deposited today. The basinward limit of deposition of the present formation overlying the Prairie Formation is at the basinward limit of the relict transgressive-phase deposits at the edge of the continental shelf which are approximately 17,000 years old.

CONCLUSIONS

It has been documented that modern clastic silicate deposition in the Gulf of Mexico is limited to a finite zone adjacent to the prograding shoreline. Basinward of this zone there is little to no deposition, and hiatal conditions exist. It has also been shown that the relict sand deposits exposed on the continental shelf are transgressive shoreline sands which were initially formed during temporary stillstands of the sea and later reworked landward by marine processes.

The relationship of depositional-episodes to the stratigraphic framework of clastic deposits in the Gulf basin has been demonstrated. This relationship is valid for all basins that have been filled by repetitive progradations from the basin margin in a manner similar to that of the Gulf. Modifications must be applied to the concept, however, for dissimilar basins such as those with borderlands. In these dissimilar basins, the zone of hiatal conditions is landward of an extensive zone of deposition owing to sediment bypassing.

The most significant aspect of the concept of depositional-episodes is that discrete, genetically related stratigraphic units are separated by major hiatal surfaces. Because of this, the depositional-complexes can be mapped and the true stratigraphic framework of the basin can be ascertained.

REFERENCES

American Commission on Stratigraphic Nomenclature, 1961, Code of stratigraphic nomenclature: Am. Assoc. Petroleum Geologists Bull., v. 45, p. 645-665.

Bates, C. C., 1953, Rational theory of delta formation: Am. Assoc. Petroleum Geologists Bull., v. 37, p. 2119-2162.

Fisk, H. N., 1944, Geologic investigation of the alluvial valley of the lower Mississippi River: Vicksburg, Miss., U. S. Army Corps Engineers, Mississippi River Comm., 78 p., 80 figs., 33 pls.

———, 1955, Sand facies of Recent Mississippi delta deposits: World Petroleum Cong., 4th, Rome 1955, Proc., sec. 1, p. 377-398, 8 figs.

———, 1961, Bar-finger sands of Mississippi delta, *in* Geometry of sandstone bodies: Tulsa, Oklahoma, Am. Assoc. Petroleum Geologists, p. 29-52.

Frazier, D. E., 1967, Recent deltaic deposits of the Mississippi River: Their development and chronology: Gulf Coast Assoc. Geol. Soc. Trans., v. 17, p. 287-315.

Geyer, R. A., 1950, The occurrence of pronounced salinity variations in Louisiana coastal waters: Jour. Marine Research, v. 9, no. 2, p. 100-110.

Heezen, B. C., 1959, Modern turbidity currents [abs.]: Eclogae Geol. Helvetiae, v. 51, no. 3, p. 521-522.

Jones, O. T., 1937, On the sliding or slumping of submarine sediments in Denbighshire, North Wales, during the Ludlow Period: Geol. Soc. London Quart. Jour., v. 93, pt. 3, p. 241-283D.

Lipsey, T. E. L., 1919, Currents at or near the mouth, Southwest Pass, Mississippi River: U. S. Eng. School, Prof. Mem., v. 2, no. 55, p. 65-122.

Ouellette, D. J., 1969, Sediment and water characteristics, South Pass, Mississippi River: Louisiana State Univ., Coastal Studies Inst., Bull. 3, p. 29-53.

Shepard, F. P., 1955, Delta front valleys bordering the Mississippi distributaries: Geol. Soc. America Bull., v. 66, p. 1489-1498.

Stetson, H. C., 1953, The sediments of the western Gulf of Mexico: Papers Phys. Oceanography and Meteorology, Massachusetts Inst. Technology and Woods Hole Oceanog. Inst., v. 12, p. 1-45.

Terzaghi, Karl, 1956, Varieties of submarine slope failures, *in* 8th Texas Conf. on Soil Mechanics and Found. Eng., Proc.: Texas Univ. Spec. Pub. 29, 41 p.

APPENDIX

RADIOCARBON AGE DETERMINATIONS

(SHOWN ON FIGURE 4)

YEARS BEFORE PRESENT	MATERIAL DATED	DEPTH BELOW PRESENT SEA LEVEL (IN FEET)	SAMPLE LOCATION LAT. N	SAMPLE LOCATION LONG. W	LABORATORY RUN NUMBER
3,875 ± 125	Brackish-marsh peat	7.5 to 9.0	29°42.8′	92°21.0′	O-1671
3,550 ± 120	Brackish-marsh peat	8.6 to 9.6	29°43.9′	92°20.2′	O-2244
2,550 ± 110	Brackish-marsh peat	7.5 to 8.5	29°43.9′	92°20.2′	O-2243
900 ± 125	Brackish-marsh peat	1.9 to 2.9	29°43.9′	92°20.2′	O-2242
425 ± 105	Brackish-marsh peat	+0.5 to -0.8	29°42.8′	92°21.0′	O-1702
350 ± 100	Brackish-marsh peat	+0.2 to -0.8	29°43.9′	92°20.2′	O-2246

RADIOCARBON AGE DETERMINATIONS WHICH DATE THE SURFICIAL SHELF SEDIMENTS

(SHOWN ON FIGURE 17)

YEARS BEFORE PRESENT	MATERIAL DATED	DEPTH BELOW PRESENT SEA LEVEL (IN FEET)	SAMPLE LOCATION LAT. N	SAMPLE LOCATION LONG. W	LABORATORY RUN NUMBER
16,940 ± 620	Valves of inner-neritic mollusks	288	27°57.1′	95°10.5′	Sh-5576
15,400 ± 510	Valves of inner-neritic pelecypods	228	28°05.9′	93°33.2′	Sh-5582
12,960 ± 470	Valves of *Rangia cuneata* (bay)	189	28°09.4′	94°17.6′	Sh-4894
12,900 ± 400	Valves of *Turritella* sp. (inner neritic)	234	26°46.0′	96°42.0′	Sh-4526
10,525 ± 215	Wood and brackish marsh peat immediately beneath transgressive-phase deposits	115	28°57.6′	93°01.4′	O-1894
9,650 ± 200	Valves of inner-neritic mollusks in calcite-cemented beach sand (on downwarped edge of shelf)	246	29°11.8′	88°47.7′	O-272
9,530 ± 270	Valves of bay pelecypods	100	26°18.9′	96°59.6′	Sh-5585
9,460 ± 310	Valves of inner-neritic pelecypods	162	28°16.5′	94°14.0′	Sh-5584
8,740 ± 260	Valves of bay pelecypods	87	26°34.2′	97°05.3′	Sh-5577
8,680 ± 270	Valves of bay pelecypods	54	28°10.0′	96°31.7′	Sh-4454
7,880 ± 520	Total organic matter (Maringouin prodelta mud)	56	28°59.0′	91°29.0′	SM-366
7,350 ± 160	Valves of middle-neritic mollusks	152	28°35.5′	92°04.6′	O-1997
850 ± 110	Valves of bay pelecypods	16	29°31.2′	89°10.0′	O-469
525 ± 105	Valves of inner-neritic pelecypods	25	29°41.7′	93°08.0′	O-935

RADIOCARBON AGE DETERMINATIONS

(SHOWN ON FIGURE 6)

YEARS BEFORE PRESENT	MATERIAL DATED	DEPTH BELOW PRESENT SEA LEVEL (IN FEET)	SAMPLE LOCATION LAT. N	SAMPLE LOCATION LONG. W	LABORATORY RUN NUMBER
>36,000	Valves of bay mollusks	485 to 490	29°16.7′	89°16.3′	L-291 M
>35,000	Worn, fragmented, pelecypod valves in Pearl River, trench-fill sand	315 to 335	29°23.6′	89°35.8′	O-81
32,875 ± 3,200	Valves of inner-neritic mollusks	557 to 564	28°57.9′	89°22.8′	O-1028
31,850 ± 3,000	Valves of inner-neritic pelecypods (immediately beneath transgressive beach sand)	360 to 380	29°16.7′	89°16.3′	L-291 L
17,150 ± 500	Wood in delta-plain silty clay	260 to 280	29°16.7′	89°16.3′	L-291 K
17,000 ± 500	Detrital wood in Pearl River trench-fill sand and gravel	400 to 420	29°21.5′	89°31.8′	L-291 N
16,600 ± 420	Valves of inner-neritic pelecypods in prodelta silty clay	330 to 332	28°57.9′	89°22.8′	O-1068
8,950 ± 190	Valves of inner-neritic mollusks in transgressive beach sand	216 to 236	29°16.7′	89°16.3′	O-86
8,400 ± 150	Valves of inner-neritic pelecypods in transgressive beach sand	115.4	29°38.4′	89°57.1′	O-353
5,350 ± 550	Valves of inner-neritic mollusks in transgressive beach sand	68.5	29°51.2′	90°01.2′	L-175 E