PHYSICAL ATTRIBUTES OF SAND BODIES WITHIN HIGH-FREQUENCY STRATIGRAPHIC SEQUENCES

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INTRODUCTION

Recently the stratigraphic concepts of genetic depositional systems (Fisher and McGowen, 1967), systems tracts (Brown and Fisher, 1977), genetic sequences (Frazier, 1974; Galloway, 1989), and seismic sequences (Vail et al., 1977) have been modified and merged to gain a better understanding of the stratal relationships and spatial distribution of both clastic and carbonate sediments using sequence stratigraphic models (Posamentier et al., 1988). These broad conceptual models have primarily been used to analyze the thick fill of sedimentary basins on passive continental margins and to develop regional stratigraphic frameworks emphasizing the global influence of eustatic sea-level fluctuations. Analytical techniques are just now being developed to combine the mapping of seismic sequences with borehole data (electric logs, sediment velocities, and sediment bulk densities) so that lithologies can be predicted within depositional sequences and potential reservoir-quality sand can be located relative to sequence boundaries.

Correctly interpreting the presence or absence of sand in clastic depositional sequences on the basis of seismic response represents a key element to sequence stratigraphic studies. These concepts are especially important in offshore or other frontier exploration areas far removed from outcrops or other geological information that would allow forward prediction of subsurface lithologies. Geophysicists using convention common-depth-point (CDP) data are able to derive general lithologic properties from relative impedance and seismic velocity (Savit and Wu, 1982). Although these derivation techniques are constantly being improved, the pre-drilling prediction of lithologies still relies heavily on seismic facies interpretations as well as a general understanding of sedimentary processes and geologic history of the basin fill.

A major objective of this research is to correlate lithologies and seismic responses in late Quaternary depositional sequences. Achievement of this objective will greatly enhance the forward prediction of stratal composition using seismic patterns, reflection configurations, and reflection terminations. Successful correlation of geophysical and sedimentological variables in the shallow subsurface offers numerous diverse opportunities to apply the results. For example, depositional models of sand bodies preserved as a result of recent fluctuations in sea level can be applied to geologic framework studies of outcrops or much thicker basin fills. Also, maps of the thickness and distribution of shallow sandy sediments and heavy minerals can be used to locate other sand deposits or concentrations of economically important minerals. Thus the seismic profiles and lithologic control (foundation borings and well logs) can result in stratigraphic models that are applicable both to investigations of the occurrence of hydrocarbons and to assessment of strategic/critical minerals.

Reconstruction of sand bodies on the basis of electric-log patterns commonly superimposes several lithofacies that accumulated during different phases of a depositional cycle. For example, reservoir sandstones deposited in nearshore marine environments are products of progradational (upward shoaling), aggradational (vertical accumulation), and retrogradational (upward deepening) processes, which together may produce a succession of amalgamated sand bodies. Recent work in the Rocky Mountains (Weimer, 1984; Dolson et al., 1991) has focused attention on physical properties of reservoirs and production performance, which are related to the depositional history and juxtaposition of sand bodies with markedly different origins. Recognizing the cross-cutting relationships between entrenched valley fill and barrier island deposits by Weimer (1984), Dolson et al. (1991), and others explains the presence or absence of reservoir rocks, as well as the great variability in fluid production depending on reservoir location within the depositional systems tract.

A primary goal of this study was to test the predictive capabilities of sandstone models derived from seismic sequences and to obtain better agreement between observed and predicted lithologies within depositional sequences. This was achieved by correlating (1) lithologic descriptions of moderately deep foundation borings and (2) discrete stratigraphic sequences mapped on seismic profiles of the Texas continental shelf and upper continental slope. Several late Quaternary sequences beneath the Texas-Louisiana continental shelf have been mapped on the basis of high-resolution seismic profiles (Berryhill and Trippet, 1980, 1981; Suter and Berryhill, 1985; Berryhill, 1987), but the lithologic characteristics of the sequences have not been adequately described because of limited data. Coleman and Roberts (1988) used foundation borings and seismic profiles to construct regional subsurface maps and cross sections of late Quaternary sequences; however, that study was restricted to the Louisiana shelf, which was constructed by ancestral deltas of the Red and Mississippi Rivers. In contrast, the Texas shelf was constructed by smaller coastal plain rivers supplied by much smaller drainage basins (Morton and Price, 1987).

METHODS

A preliminary late Quaternary sequence stratigraphic framework for offshore Texas was established using numerous foundation borings and several grids of high-resolution seismic profiles (figs. 1 and 2). These two principal data bases were integrated to evaluate the relationships between seismic reflection characteristics, subsurface lithologies, and

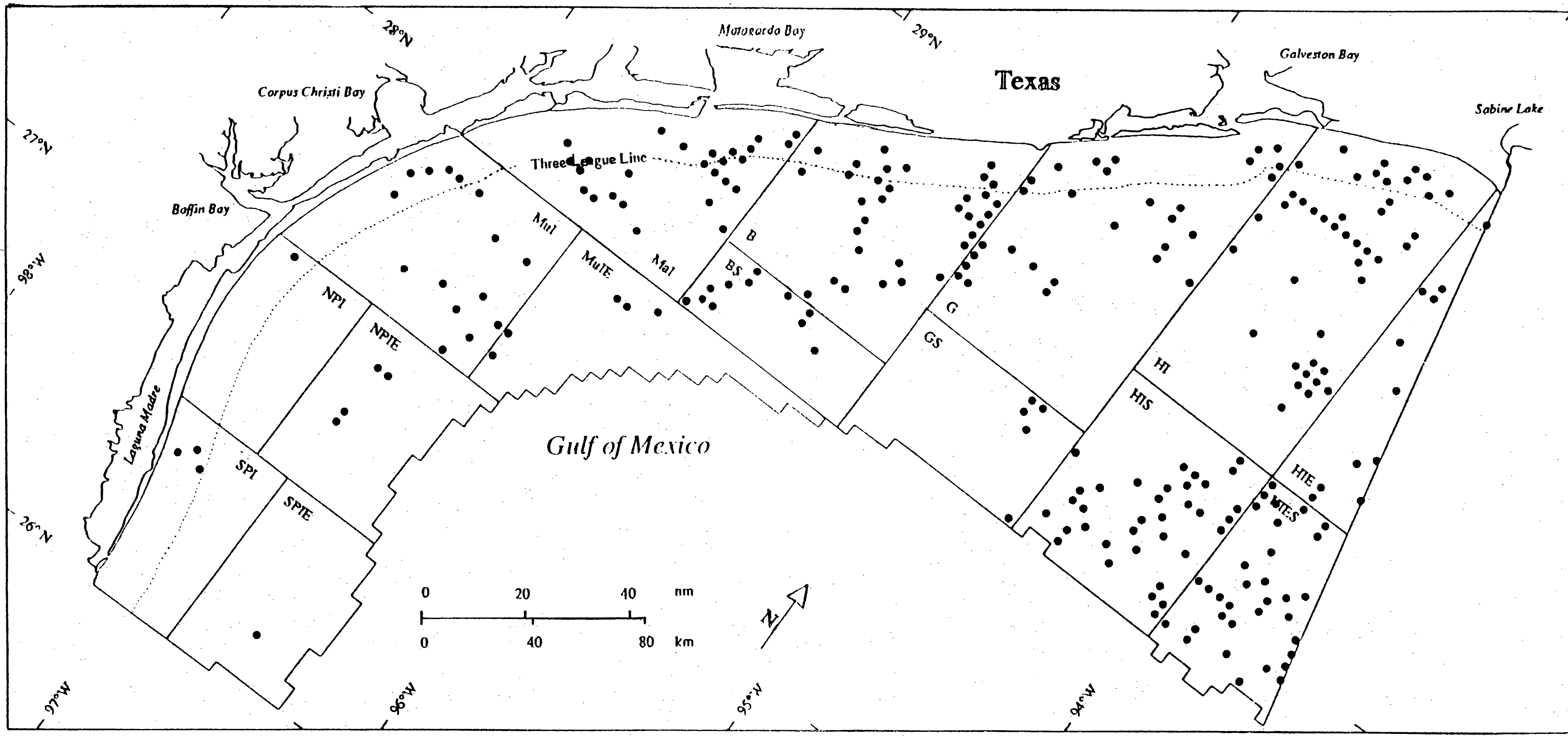

Figure 1. Locations of lease blocks on the Texas continental shelf containing at least one foundation boring.

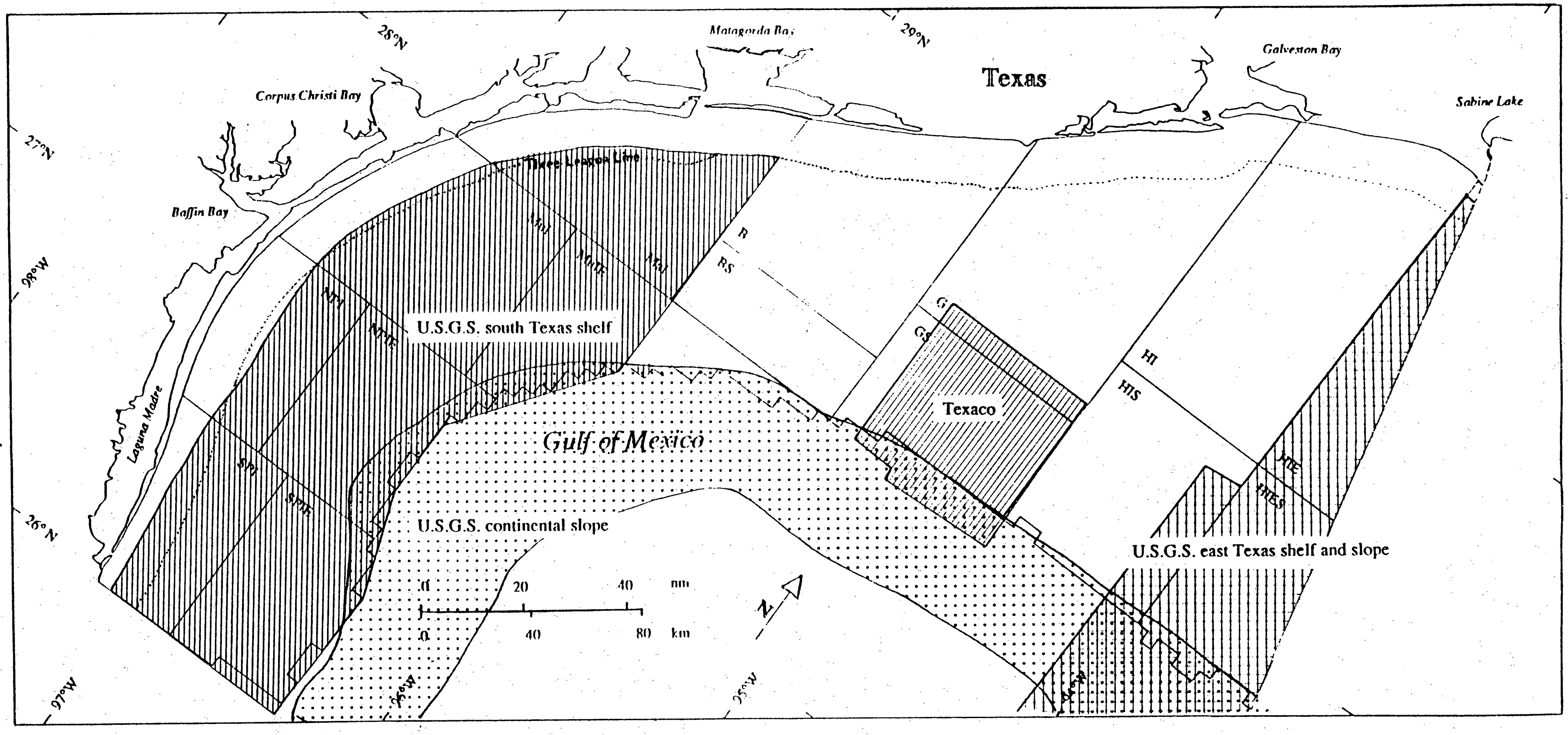

Figure 2. Locations of subregional high-resolution seismic surveys on the Texas continental shelf available for this study. The map does not show small surveys conducted for geological hazards investigations.

stratigraphic sequences. Seismic reflections were interpreted and mapped at selected sites throughout the study area to provide detailed correlations and to determine reflections of subregional significance. Depositional environments and depositional systems tracts were interpreted on the basis of seismic characteristics, lithologic descriptions, stratigraphic position, and paleogeographic location. Emphasis was placed on diverse depositional settings where numerous closely spaced foundation borings near seismic profiles could provide the control necessary to interpret the geologic history of the region and possibly to isolate the influence of global sea-level fluctuations on the construction and preservation of stratigraphic sequences.

Foundation Borings

More than 150 foundation borings were used to determine the lithologic compositions of the sequences and to establish the lithogenetic correlation framework. Most of the borings penetrate more than 90 m (300 ft) below the sea floor and some go as deep as 150 m (500 ft) below the mudline. The boring information was compiled mostly from proprietary sources. Detailed boring descriptions provide information regarding subsurface depth, sediment color, sediment composition, presence and concentrations of accessories (organic matter, shells, calcareous nodules), sediment textures, engineering properties, and other sedimentological attributes that can be used for lithostratigraphic correlations. The borings are well distributed throughout the study area (fig. 1) and many are close enough to the seismic lines that they could be projected into the lines with confidence.

The foundation boring descriptions are prepared by skilled technicians trained to observe and record changes in the cored material. These descriptions, which are based on standard procedures and definitions of physical properties, are supplemented with data from geotechnical tests and grain-size analyses. The final report presents lithologic descriptions and engineering test results in plots corresponding to depth below the sea floor.

Seismic Profiles

More than 1500 trackline miles of high-resolution sparker profiles (fig. 2) were interpreted to establish the regional chronostratigraphic framework and to evaluate the relationships between seismic reflection patterns and lithologies. Styles of sediment accumulation were classified on the basis of stacking patterns of seismic reflections. Uniform and parallel patterns on both strike and dip lines typically indicate aggradation, clinoforms always indicate either progradation or lateral accretion, and low-angle onlapping patterns commonly indicate retrogradation.

Depositional sequences were recognized and interpreted on the basis of internal seismic reflections, reflection terminations along sequence boundaries (Mitchum et al., 1977), lithologies, geometries, and spatial variations in lithofacies and seismic facies. Sequence boundaries identified and selectively illustrated by Berryhill (1987) and by Morton and Price (1987) were verified and used to establish the general stratigraphic framework. Sequence stratigraphic correlation among the foundation borings is possible because the first multiple either occurs below the boring depths or does not obliterate reflections above the boring depths.

Some high-amplitude seismic reflections appear to be sequence boundaries on a single profile, but they do not correspond to equivalent strong reflections on adjacent profiles. These variable amplitude reflections typically occur near the base of a regressive sequence and they eventually merge with clinoform reflections. Such local reflections cannot be correlated throughout the seismic grid and they occur progressively lower in the section toward the direction of progradation.

Seismic Velocities

The speed of sound in late Quaternary clastic sediments can be highly variable depending on such factors as sediment composition, water saturation, degree of consolidation, and presence or absence of gas. Previous geophysical studies have used average acoustical velocities to approximate depths of late Quaternary strata from two-way travel times recorded on seismic profiles. Acoustical velocities used in previous studies range from 1525 m/sec, the speed of sound in seawater (Sidner et al., 1978; Suter et al., 1987), to 1700 m/sec (Bouma et al., 1983). Lehner (1969) used a linear velocity function that incorporated a gradual increase in the speed of sound in sediments with depth. Despite the ability to adjust velocities at depth, the increases in calculated velocities at shallow depths are negligible. Therefore, a constant velocity provides a useful approximation without the effort required to make the more precise, but not necessarily more accurate, adjustments. Local structures probably cause greater errors in projecting foundation boring depths into the seismic lines than do differences in assumed acoustical velocities.

In the present study, an acoustical velocity of 1675 m/sec was used to convert two-way travel time to depth because: (1) previously exposed and overconsolidated Pleistocene sediments should transmit sound faster than does seawater, and (2) 1675 m/sec agrees closely with the velocity Lehner (1969) proposed considering the depth of investigation. Furthermore, lithologic changes observed in foundation borings commonly coincide with unique seismic reflections, indicating that the average velocity selected is reasonably accurate for these shallow sediments.

Stratigraphic Correlations

Seismic reflections, sediment color, and vertical lithologic successions were the principal criteria used to establish the stratigraphic framework of late Quaternary sediments. Seismic sequence boundaries and diagnostic reflection patterns (clinoforms, channels) were transferred to geological cross sections so that lithologic variability within the seismic stratigraphic sequences could be evaluated. Soil horizons, identified by sediment color, were the primary physical evidence from the foundation borings that were also used as correlation markers. Correlations were made on the basis of soil horizons or seismic reflections with equal confidence because they coincided in numerous foundation borings. Secondary correlation criteria included systematic changes in sediment textures (upward-coarsening and upward-fining patterns) within the context of the overall sequence framework.

Holocene Sediments

Only the late Wisconsin-Holocene transgressive sediments were correlated strictly on the basis of lithologic succession. This deviation from established correlation procedures was necessary because the Holocene sediments are so thin that their seismic record is obscured by the broad width of the seismic bubble pulse.

The Holocene transgressive deposits are identified on the basis of sediment composition, induration, and water saturation of the muds. The Holocene deposits are generally very soft to soft and contain as much as 80% water. They are composed mostly of gray clay, silty clay, or sandy clay; less common compositions are silty and clayey sand. All of these lithologies can contain variable amounts of shell fragments and they all can be similar to or different from the underlying lithologies. For example, muddy Holocene sediments can either overlie other muddy sediments or sandy sediments; similarly, the sandy sediments can either overlie other sandy sediments or muddy sediments. Descriptions of a few borings suggest that the Holocene sediments at some sites are composed of both sand and mud. These stacked lithologies represent coastal evolution and migrating depositional environments that produce stacked facies such as sandy beach deposits over muddy coastal plain marsh or lagoonal mud. Sandy beach deposits overlain by offshore mud represent another example of coastal evolution preserved in these young sediments.

Pre-Wisconsin Sediments

There is a substantial correlation discrepancy between pre-Wisconsin sediments identified on the high-resolution sparker profiles (Suter and Berryhill, 1987) and those identified using deeper subsurface data (Morton et al., 1991). Suter and Berryhill (1987) matched transgressions and regressions interpreted on the sparker profiles with the most recent rises and falls of sea level shown on published sea level curves. This "count down" method of chronostratigraphic interpretation suggests that the Sangamon highstand deposits are relatively shallow beneath the continental shelf and should be penetrated by the deepest foundation borings.

Using petroleum industry data, Sangamon interglacial deposits are identified on the basis of faunal assemblages and the extinction of the foraminifera *Globorotalia flexuosa* (Worandt, 1991). Maps and cross sections of this extinction horizon, correlated in well logs and on CDP seismic profiles, suggest that Wisconsin sediments may be more than 2,000 ft thick near the shelf margin (Morton and Jirik, 1989; Morton et al., 1991) and are not encountered in even the deepest foundation borings.

There are at least two explanations for the pre-Wisconsin correlation discrepancy. First, if the "Sangamon" interglacial transgression and highstand mapped in the deeper subsurface is the 130,000 yr B.P. event, then there are more Wisconsin sequences preserved than previously recognized. On the other hand if the subsurface "Sangamon" interglacial event is actually older than 130,000 yr B.P., then the three sequences mapped by Suter and Berryhill (1987) could represent all the sequences of Wisconsin age. Without absolute age dates or indirect evidence such as oxygen isotope data, this discrepancy is not easily resolved and it cannot be resolved just using foundation borings and sparker profiles. Most correlation charts now indicate that the *Globorotalia flexuosa* extinction horizon is older than 130,000 yr B.P.; consequently, the chronostratigraphic relationships established in the High Island Area by Suter and Berryhill (1987) and Berryhill (1987) were maintained for the purposes of this study.

STRUCTURAL FEATURES

Geological structures influencing late Quaternary sediments near the extant shelf margin formed as a result of gravity-driven tectonism, principally tensional stresses and sediment mobilization. These structures are near-surface expressions of deeper Plio-Pleistocene structures that can be seen on multi-channel CDP seismic profiles (Morton et al., 1991). The four main classes of structures are faults, salt diapirs, withdrawal basins, and unconformities. Continuous movement along these structures or reactivation of older structures created the accommodation space for deposition of the youngest sequences.

Faults

Faults that displace the shallow sediments coincide with families of faults that also disrupt the deeper Plio-Pleistocene strata (Morton et al., 1991). They include both major and minor fault zones depending on their continuity and vertical throw. Several stages of faulting and reactivation of older faults are common owing to episodic movement of salt and shifting sites of diapirism. The balance between fault movement and sedimentation is commonly expressed as the presence or absence of fault scarps at the sea floor. Recent movement of regional extension faults in areas of low sedimentation result in offset of the sea floor of as much as 50 ft (15 m) whereas little or no offset occurs where rates of sedimentation are high. The largest fault scarps are located in the southern third of the High Island Area near the extant shelf margin where salt mobilization and basinal extension are still active processes.

En echelon faults that form laterally continuous belts are termed regional or counterregional expansion faults because their syndepositional movement causes increased thickness and vertical separation of Plio-Pleistocene sequences on the downthrown side. Regional expansion faults, which exhibit down-to-the-basin displacement, are subparallel to the paleoshelf margins of the underlying sequences. In contrast, counterregional faults have the opposite throw as a result of late salt migration and formation of large salt withdrawal basins near the shelf margin. The counterregional faults commonly are paired with a primary regional fault that together form the basinward and landward boundaries of salt or shale withdrawal basins.

More localized faults having minor throw and limited lateral extent are either associated with salt diapirs or are antithetic to the major regional or counterregional faults. Faults associated with diapirs are generally closely spaced and radial to the diapir. They typically have complex and opposing displacements that form grabens over the dome crest. Some diapir-related faults join with master faults of the regional or counterregional trends. Antithetic or stress-relief faults have displacements that are opposite to the major fault trends. These secondary faults can only be accurately identified from CDP seismic profiles because their intersection with the master faults occurs far below depths penetrated by the sparker profiles.

Contemporaneous slumping and creation of small intraformational faults occur where the shelf margin is convex-upward and laterally unconfined in a basinward direction. The oversteepened profile and low sediment strength create slope instabilities that promote detachment and downward rotation of large fault blocks. Slope failures range in scale from small rotational slump blocks and slides within a single, relatively thin depositional sequence to extremely large translation and rotation along regional growth faults.

Salt Diapirs

Two classes of salt diapirs influence the thickness and distribution of Late Quaternary sequences in the High Island Area. The first class includes deep-seated diapirs that are manifested as broad, faulted structural highs. The highs are identified by minor to moderate dip of beds on the flanks of the structures. For these structures, salt does not penetrate the shallow section and can only be observed on multichannel CDP seismic lines. An example is the structural high on the upthrown side of the *Trimosina* fault zone (Morton et al., 1991). This high, which is supported by a deep salt ridge, is actually a manifestation of differential subsidence. Subsidence is greater in the salt withdrawal basins and is less where the salt mass remains constant or increases as a result of evacuation from adjacent flanks of the diapir. The second class of salt diapirs is composed of those structures where salt is observed on the sparker profiles and adjacent strata dip steeply away from the diapir.

The spacing, size, and shape of the salt bodies change in a basinward direction. Domes located beneath the continental shelf are narrow, nearly circular spines that are widely spaced, suggesting a mature stage of dome evolution (Woodbury et al., 1973). An intermediate stage of salt evolution is represented on the upper slope by large, nearly continuous massifs that are separated by sediment-filled basins. Many of these diapirs on the continental slope are at or so near the sea floor that they create bathymetric highs. East and West Flower Garden Banks are both reefs growing on shallow salt diapirs that cause large relief of the sea floor.

Unconformities

On seismic lines, regional, subregional, and local unconformities were recognized on the basis of stratal patterns that indicate erosion such as onlap and truncation. The most widespread sequence bounding unconformity is located beneath the South Texas Outer

Continental Shelf probably because this was an interdeltaic setting that only received minor sediment supply during the lowstand in sea level. A subregional unconformity mapped in the High Island Area coincides with a soil horizon; the soil developed on top of deltaic deposits when they were subaerially exposed. Elsewhere, unconformities are either not recognized (seismic reflections are parallel) or the missing section is limited in areal extent, such as around uplifts or at the bases of fluvial channels.

DEPOSITIONAL SYSTEMS, SYSTEMS TRACTS, AND SEQUENCE BOUNDARIES

Depositional Systems

Clastic depositional systems are three-dimensional assemblages of genetically related sedimentary facies (Fisher and McGowen, 1967). They include the common coastal plain and nearshore environments such as rivers, deltas, and barrier-strandplains as well as deeper water environments such as shelves, slopes, and abyssal plains. All of the nearshore systems contributed to construction of the Texas continental margin at some time during the late Quaternary. This report primarily emphasizes fluvial and deltaic systems because they are the principal conduits of sediment transport and loci of sand introduction into the marine environment.

Our current understanding of shelf-margin delta systems is largely based on late Quaternary deltas of the northwestern Gulf of Mexico. Both high resolution sparker and CDP seismic profiles display late Pleistocene shelf-margin deltas that prograded onto the slope and filled former intraslope basins (Lehner, 1969; Suter and Berryhill, 1985; Suter et al., 1987). These delta systems were principally active during falling stages of sea level and subsequent lowstands (lowstand systems tract). In contrast, the modern Mississippi shelf-margin delta was deposited as a result of abundant sediment supply during a recent highstand in sea level (highstand systems tract). Deformation features at this depocenter are concentrated near the shelf edge and include surficial landslides, large-scale slumps, and other evidence of mass movement. Basinward transport of deltaic sediments by creep and gravity flow is caused by gravitational instability along oversteepened slopes of the delta front (Coleman et al., 1983). These rapidly deposited sediments contain abnormally high water volumes and some biogenic gas. The interstitial fluids create high pore pressures and reduce sediment shear strength that together contribute to slope failure.

Slope instabilities near the shelf margin cause different scales of sediment deformation that encompass at least three orders of magnitude. Small-scale surface creep, slumps, and gullies deform sediments a few tens of meters thick (Prior and Coleman, 1982; Coleman et al., 1983; Lindsay et al., 1984). These features normally are contained within the clinoform facies and contribute to progradation of the shelf margin. An intermediate scale of sediment deformation involves large slope failure whereby blocks of sediment slide downslope and bedding is either preserved, distorted, or transformed into debris flows. This type of intraformational slide encompasses a few hundred meters of sediment and occurs above a detachment surface that merges with bedding planes downslope. The largest scale of gravity-induced slope deformation includes deep-seated normal faults and sediment diapirs that grow in response to sediment loading at the shelf margin. This basinal scale of extension and deformation involves strata as much as several thousand meters thick and basinward displacements on the order of tens of kilometers.

Systems Tracts and Sequence Boundaries

Depositional systems tracts are key elements of sequence stratigraphic models used to interpret geological histories of sedimentary basins from stratal patterns and lithologic variability of the basin fill. They were originally defined on the basis of geological processes, sedimentological characteristics, three-dimensional facies relationships, and the resulting stratigraphic successions (Brown and Fisher, 1977). This integrated definition emphasizes coeval spatial relationships including the lateral distribution of interconnected depositional systems and their component facies without any reference to eustatic fluctuations or sea-level position.

Recently systems tracts have been redefined strictly on the basis of interpreted sea-level cycles and inferred segments of the eustatic curve (Posamentier et al., 1988). This redefinition of systems tracts, which is inextricably linked to specific sea level phases by descriptions, illustrations, and terminology, emphasizes temporal relationships especially as they apply to control of stratal stacking patterns presumably by eustatic fluctuations. According to the eustatic sequence stratigraphic models, a sequence is bounded by unconformities or their correlative conformities and is composed of sediments deposited by lowstand, transgressive, and highstand systems tracts (Posamentier et al., 1988).

The two types of unconformities that bound sequences are defined on the basis of the rate of sea level fall relative to the rate of subsidence. A type I unconformity forms when the rate of sea level fall is greater than the rate of subsidence at the shoreline break. Under these conditions the shelf is exposed, rivers incise valleys, and canyon cutting is initiated. These erosional processes produce large volumes of potentially coarse sediment eroded from the shelf and deposited on the slope and basin floor forming lowstand submarine fans (lowstand systems tract). As the rate of sea-level fall declines, river channels aggrade and trap the coarse-grained sediments so that only fine-grained material reaches the upper slope and fills topographic lows created by slumps. When relative sea level gradually begins to rise, deposition is largely confined to shelf-margin deltas and younger submarine fans (lowstand progradational wedge) deposited between the shelf edge and the basin floor submarine fans. A type II unconformity is created when the rate of sea level fall is less than the rate of subsidence (Posamentier et al., 1988). Under these conditions the shelf is only partly exposed, fluvial channels are not incised, and the delta systems prograde only to mid shelf or shelf edge positions. Because sediments are mostly deposited on the shelf there isn't the large-scale transfer of sediment onto the slope as occurs when sea level is rapidly lowered and a type I unconformity is produced.

A rise in relative sea level causes an abrupt landward shift in deposition and coastal onlap (transgressive systems tract). As the continental platform is flooded, coarse sediments are trapped near the shoreline creating sediment starvation on the shelf and formation of a marine condensed section across the shelf and slope. In deep water, only hemipelagic and pelagic mud is deposited as drapes over the submarine fan. During the final stages of rising sea level and highstand (highstand systems tract), large delta systems prograde toward the shelf edge over the transgressive deposits while mud continues to accumulate on the slope (Posamentier and Vail, 1988).

In addition to type I and II unconformities, Nummedal and Swift (1987) defined several other erosion surfaces that occur in some stratigraphic sequences. For the purposes of the present study the most important erosion surfaces are the marine ravinement diastem (Swift, 1968) and the channel diastem. Ravinement surfaces are regional, nearly planar surfaces produced by submarine erosion when the shoreface retreats across the former coastal plain. Owing to the depth of erosion, the ravinement surface may modify or locally replace the sequence boundary where lowstand and transgressive systems tract deposits are thin or absent. The channel diastem is a local erosion surface created by the downcutting or lateral migration of a channel. Channel types are not specified and can be of fluvial, tidal, or submarine origin. In this study, significant channel erosion surfaces are associated with fluvial and distributary channels.

The eustatic sequence stratigraphic and systems tract models, which evolved at Exxon Production Research over the past few decades, are biased toward principal depocenters and extremely large fluctuations in sea level. Sea level fluctuations during the past 130,000 yr were undeniably large and yet the depositional response along the shelf margin was substantially different away from the depocenter because of differences in sediment supply (Morton, 1991).

LATE QUATERNARY SEA-LEVEL HISTORY

Sea level curves for the northwestern Gulf of Mexico (Curray, 1960; Nelson and Bray, 1970; Frazier, 1974) were derived largely from depths and ages of features submerged on the continental shelf. Actual age/depth relationships give conflicting sea-level histories because dated features at different depths have similar ages and vice versa. These differences can be explained by crustal warping and local diapirism, but the curves are drawn as if isostatic and tectonic deformation was negligible. Although details may differ, sea-level curves for coastal clastic provinces, including the Gulf of Mexico, generally have configurations that are comparable to those published for other coastal regions.

Most eustatic curves (fig. 3) show that sea level was at a highstand about 130,000 ka (isotope stage 5). It began falling about 110 ka and reached its lowest position during early Wisconsin glaciation about 70 ka (isotope stage 4). This lowstand was followed by a moderate (about 30 m) rise in sea level that reached mid shelf during the middle Wisconsin interstadial about 50 ka (isotope stage 3). The following late Wisconsin glaciation caused another drop in sea level that lasted until about 18 ka (isotope stage 2) when sea level began a rapid but irregular rise that lasted until about 3,500 ka (fig. 1). Since then, sea level has remained essentially constant at its present highstand (isotope stage 1), although there may have been minor fluctuations having amplitudes of a few meters or less.

During the early and late Wisconsin lowstands in sea level (isotope stages 4 and 2), several shelf margin deltas were constructed where major river systems encountered the ancestral shelf-slope break (Suter and Berryhill, 1985; Suter et al., 1987). These deltas (fig. 4) did not coalesce along the entire shelf margin, rather they remained located near their respective river mouths that were fixed in space by the entrenched fluvial systems. These deltas have been known since detailed bathymetric surveys of the outer shelf and upper slope revealed their lobate geometries and relatively steep gradients along the delta fronts and delta flanks (Curray, 1960). According to Fairbanks (1989), the last lowstand in sea level (isotope stage 2) reached a depth of about 120 m below present sea level. This depth agrees reasonably well with the brows of some lowstand deltas preserved along the Texas continental margin, but it is actually deeper than some progradational shelf-margin deltas even where subsidence rates are high. These discrepancies suggest that either the lowstand in sea level was less than estimated or the response of river systems was different along the shelf margin.

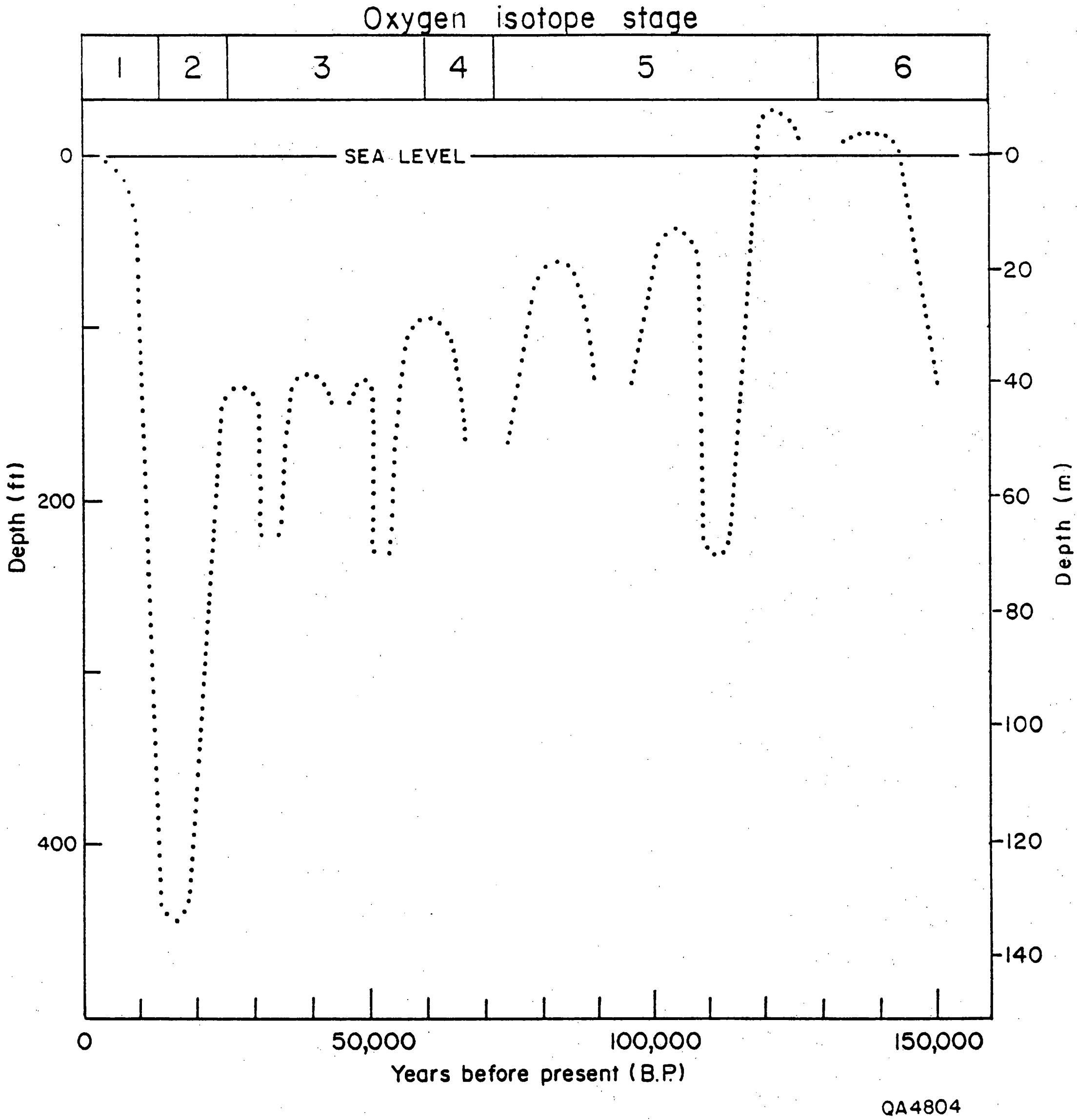

Figure 3. Late Quaternary glacio-eustatic fluctuations and corresponding oxygen isotope stages. Sea-level curve from Moore (1982).

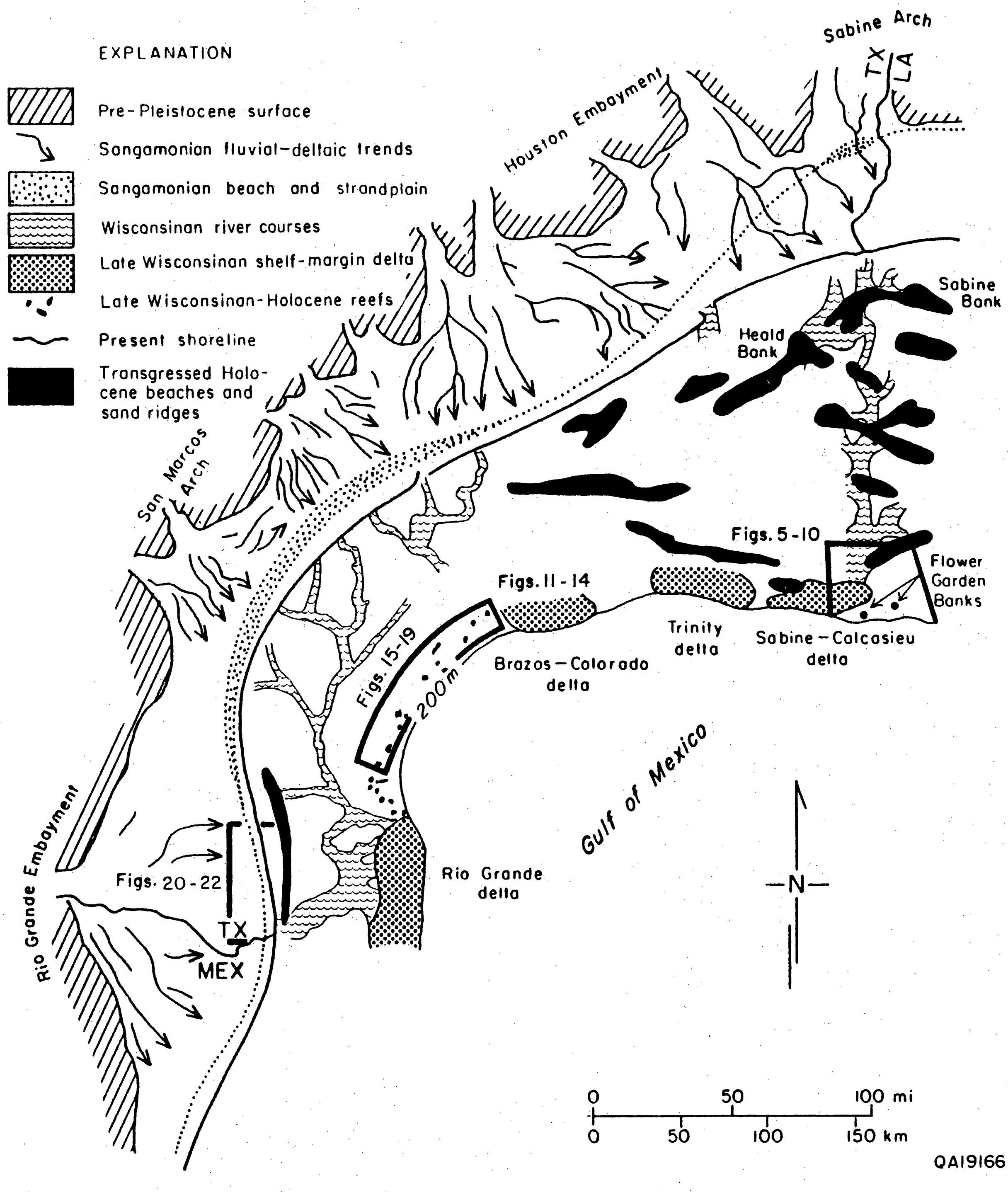

Figure 4. Late Quaternary depositional systems of the northwestern Gulf of Mexico. Modified from Morton and Price (1987).

LATE QUATERNARY SEQUENCE STRATIGRAPHY

Stratigraphic sequences from four different depositional settings (fig. 4) were examined to evaluate seismic responses of different lithologies and to determine how depositional settings influence sand-body attributes (Table 1). Each setting represents a different paleogeographic position with respect to the depocenter and shelf margin and the strata at each site record a different response to sea-level fluctuations and changes in sediment supply. Altogether three deltaic settings and one interdeltaic setting were investigated. The coastal plain rivers and deltas analyzed represent moderately large systems fed by continental drainage southeast of the Rocky Mountains and High Plains, but these systems were not influenced by glacial meltwaters because of their lower latitudes (Morton and Price, 1987).

The principal shelf-margin delta setting (High Island Area) includes a thick wedge of lowstand sediments deposited near the brow of the continental platform. This lowstand delta system was responsible for both aggradation and progradation of the shelf margin into deeper water where the shelf-slope transition was very narrow.

The delta flank setting of another shelf-margin delta system (Brazos South Area) provides insight into the stratigraphic record of laterally reworked sediments that are derived from the delta front and then redeposited along the delta margin.

The platform delta setting (South Padre Island Area) examines highstand and falling sea-level deposits that accumulated as the shorezone systems prograded basinward from the shoreline of maximum transgression. Thus, it records deltaic sedimentation in relatively shallow water and across a broad preexisting continental shelf.

In contrast to the other depositional settings, the interdeltaic setting (Mustang Island Area) emphasizes shelf processes and depositional products away from the depocenter and adjacent shorezone. Sediments were transported to this site by along-shelf and cross-shelf currents that were influenced by geometries and positions of the lowstand shelf-margin delta systems.

Late-Wisconsin Deltaic Depocenter (Southern High Island Area)

Deep foundation borings in the southern High Island Area (figs. 5–9) reveal four late Quaternary sequences that were deposited near the shelf margin on a preexisting continental platform. Sea-level histories and deep subsurface investigations indicate that the platform

Table 1. Geometric characteristics of sand bodies categorized by depositional setting.

Depositional Setting	Sand Body Geometry and Continuity
Wave-dominated shelf delta	Thick, tabular or sheet sand, continuous in strike and dip directions
River-dominated shelf delta	
Distributary-mouth-bar	Thin, lenticular, greater continuity parallel to depositional dip compared to depositional strike
Distributary channel	Lenses of variable thickness and elevation, greater continuity parallel to depositional dip compared to depositional strike
Shelf-margin delta	
Distributary-mouth-bar	Thick, tabular or sheet sand continuous in strike and dip directions
Distributary channel	Lenses of variable thickness, greater continuity parallel to depositional dip compared to depositional strike
Delta-flank barrier	Thin, lenticular, and elongate parallel to depositional strike, narrow lens parallel to depositional dip
Interdeltaic shelf ramp	Sand bodies typically rare in mud-dominated shelf systems. Sand deposits are very thin, patchy, packages of interlaminated sand and mud.

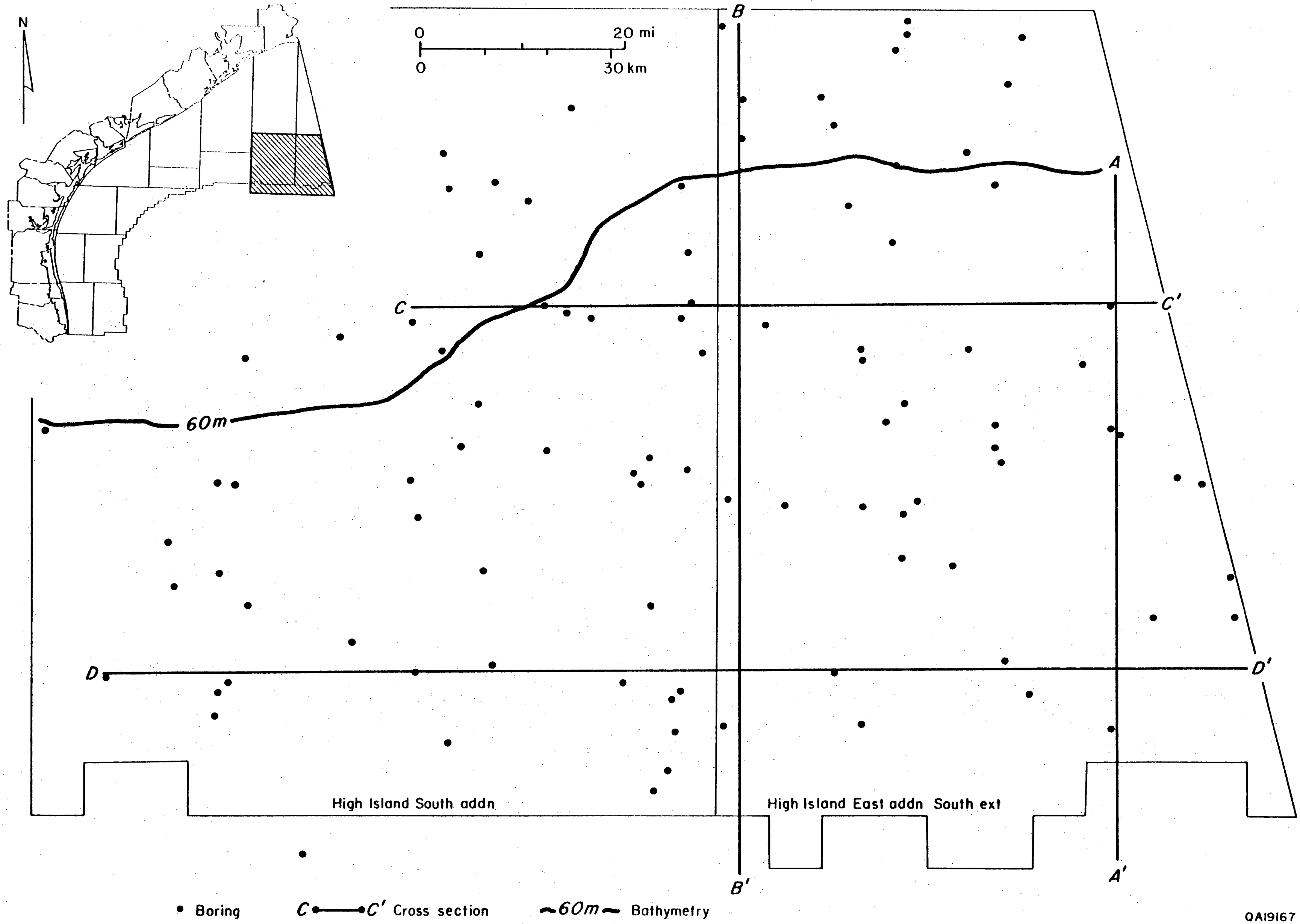

Figure 5. Locations of stratigraphic sections and seismic profiles in the southern High Island Area near late Pleistocene lowstand shelf-margin deltas.

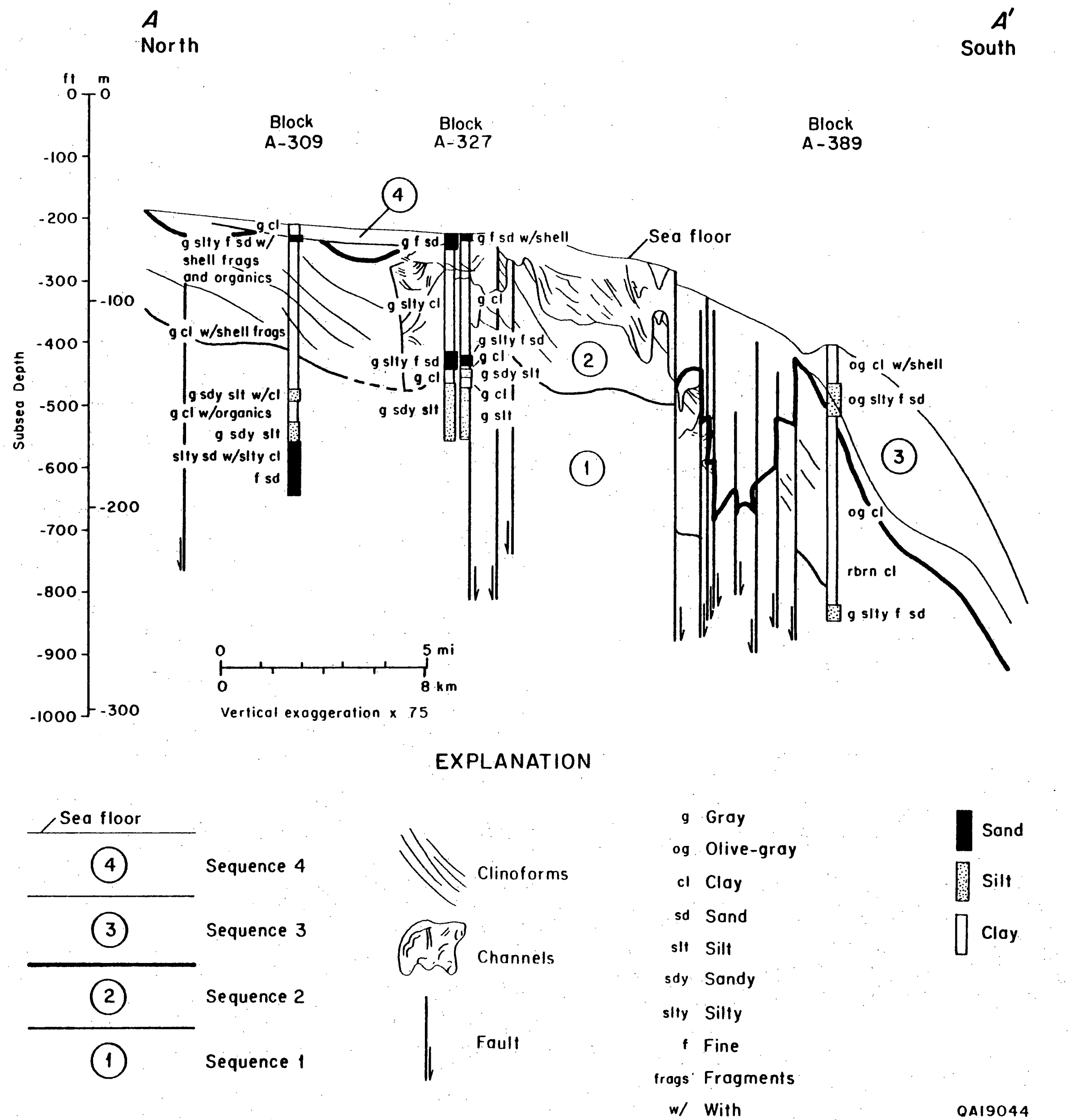

Figure 6. Dip section A-A' in the High Island area showing late Pleistocene and Holocene stratigraphic sequences and recent structural deformation.

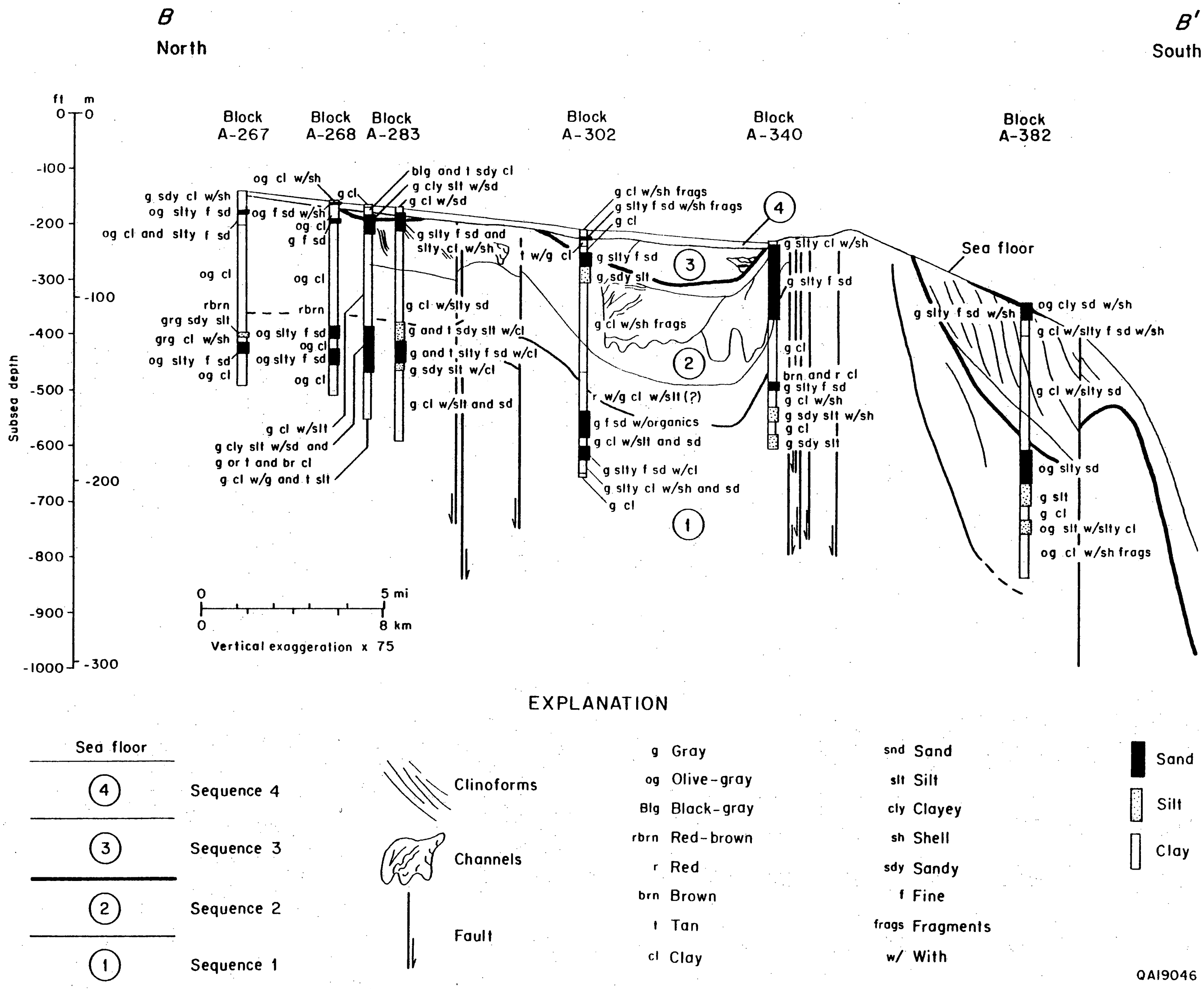

Figure 7. Dip section B-B' in the High Island area showing late Pleistocene and Holocene stratigraphic sequences and recent structural deformation.

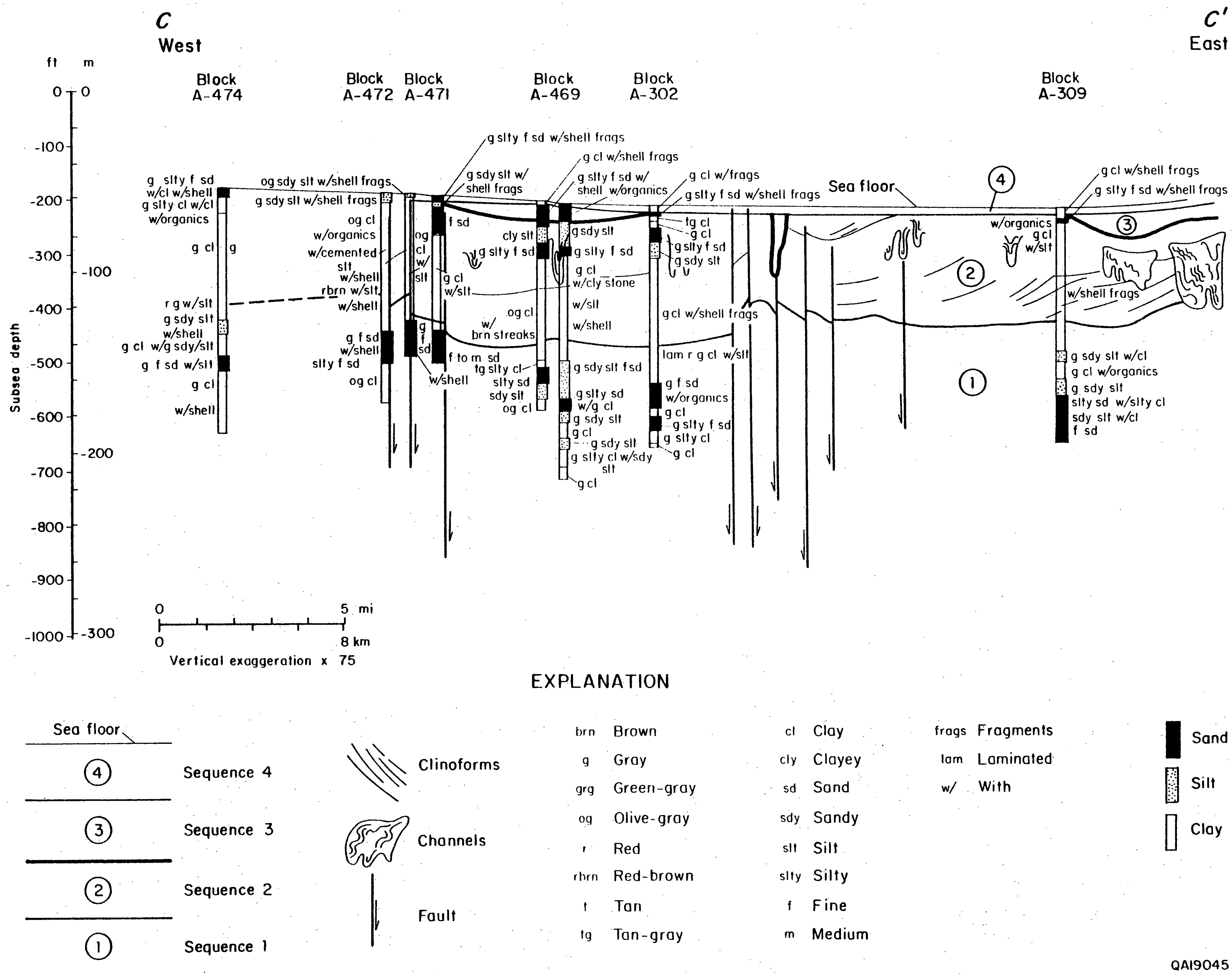

Figure 8. Strike section C-C' in the High Island area showing late Pleistocene and Holocene stratigraphic sequences and recent structural deformation.

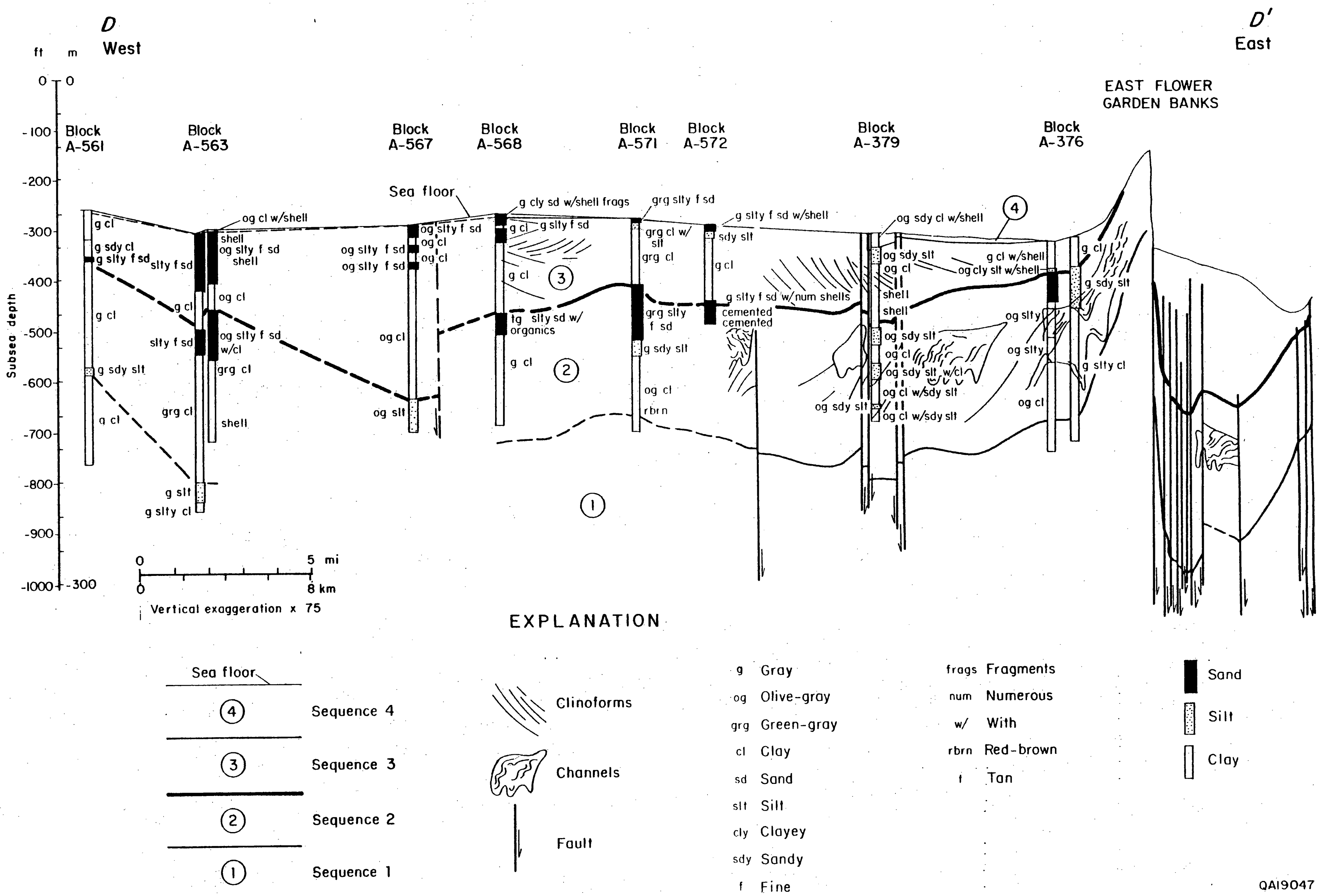

Figure 9. Strike section D-D' in the High Island area showing late Pleistocene and Holocene stratigraphic sequences and recent structural deformation.

was constructed during previous lowstands in sea level when the shelf margins in this area occupied the same average position for at least the past 500,000 yr (Morton et al., 1991).

The three oldest sequences represent upward-coarsening fluvial-deltaic complexes. The deltas were fed by combined drainage systems (Trinity, Sabine/Neches, Calcasieu) as indicated by the trends of incised valleys crossing the continental shelf (fig. 4). The older delta complex prograded in a southerly direction, whereas the youngest progradational wedge exhibits a predominant westerly direction of delta construction. Transgressive marine clay beds, which separate both deltaic wedges, occur at an average depth of about 60 m below the sea floor. Some local structures (faults and salt domes) greatly influenced deltaic deposition and as much as 50 m of section is absent over local structural highs that moved recently (figs. 7 and 9). This implies that short-term rates of subsidence and sedimentation can be significant on a "stable platform" effected by basinal extension and salt migration.

Seismic Characteristics

Late Quaternary shelf margin deltas in the southern High Island Area exhibit all of the common seismic reflection patterns including parallel, divergent, clinoforms, cut-and-fill, hummocky, chaotic, and reflection free (Table 2). The most common patterns are clinoforms and parallel reflections having variable amplitude and variable continuity. The clinoforms typically change shape from low-angle oblique to high-angle sigmoid forms in a basinward direction (figs. 6 and 7). This change in depositional style reflects the effects of progradation into progressively deeper water and the influence of a rise in relative sea-level that eventually overcomes sediment influx (Mitchum et al., 1977).

Channel dimensions, channel shape, and internal reflection characteristics were used to distinguish between principal meandering trunk streams and distributary channels. Wide, deep, and nested channels exhibiting evidence of lateral migration and repeated occupation (figs. 6–10 sequence 2) are interpreted as major alluvial channels that have eroded into the underlying (slightly older) deltaic deposits as a result of prolonged progradation. In contrast, single, narrow channels that do not exhibit evidence of lateral migration (fig. 8, sequence 3) are interpreted as distributary channels.

Sequence Composition

Sequence 1, the oldest sequence, only partly defined because the basal sequence boundary is obscured by multiples or poor data quality. Judging from penetration depths,

Table 2. Associations of high-resolution seismic reflections, lithologies, and depositional environments of late Quaternary shelf and shelf-margin depositional systems.

Reflection Type	Associated Lithologies	Depositional Environments
Parallel	Mud-dominated lithofacies except for thin sandy transgressive shorezone deposits	Flood basins on alluvial plains and delta plains, also shelf platform or ramp morphological setting
Divergent	Predominantly muddy lithofacies	Indicates subtle structural influence or increased water depth rather than a particular depositional environment
Clinoforms	Thick, high-angle sigmoid or oblique clinoforms are mud prone whereas thin high-angle or irregular clinoforms landward of the shelf margin are typically sand prone	Thick clinoforms characterize prodelta and upper slope deposits whereas thin or irregular clinoforms landward of the shelf margin characterize shorezone deposits
Channel or Cut-and-Fill	Large erosional features may be filled with either sand or mud depending on abandonment history, lateral accretion patterns of individual channels suggest meandering and possible deposition of sand-rich point bars	Entrenched-valley fill, fluvial channel fill
Hummocky/Wavy	May be either sand prone or mud prone depending on the original material that was deformed	Indicates minor soft-sediment deformation rather than depositional environment. Typically associated with delta front, prodelta, and slope of unstable shelf-margin deltas
Contorted/Chaotic	May be either sand prone or mud prone depending on the original material that was deformed	Indicates substantial mass movement rather than depositional environment. Typically associated with delta front, prodelta, and slope of unstable shelf-margin deltas
Reflection Free	Sand prone when located immediately above clinoforms, otherwise not indicative of lithology	Not diagnostic of depositional environment, may represent gas content or other physical property that attenuates the acoustical signal

the sequence is best represented by foundation borings located in the northern part of the southern High Island Area (figs. 6–9). Sequence 1 typically consists of about four lithologic units. The basal unit is composed of stiff gray clay and silty clay with rare layers of shell. The overlying unit is composed of olive gray to gray silty fine sand. The sandy lithofacies ranges in thickness from 4.5 to 45 m (15 to 150 ft). Overlying the sand is sandy silt or silty clay (lithofacies 3). The superposition of sandy silt or silty clay over the sand indicates an upward-fining succession. The uppermost lithologic unit is composed of stiff gray clay with thin interlayers of sand and silt and containing some calcareous nodules and shell fragments. The upper boundary of sequence 1 is a soil zone (figs. 6–9) that is 1 to 8 m thick. Colors of the soil zone are red, brown, or yellow and occasionally these colors mixed with gray. At several locations there are two soil horizons about 12 m (40 ft) apart. The repeated development of soil profiles suggests aggradational processes and frequent subaerial exposure.

The vertical facies successions and lithologies indicate that sequence 1 was deposited during a regression that culminated in coastal plain aggradation above sea level and formation of a soil profile. This soil zone was partly preserved even after submergence and truncation by ravinement during the subsequent transgression.

Sequence 2, which is entirely penetrated by numerous borings (figs. 6–9), consists of four lithologic units. The basal unit, which is also the thickest unit, is composed of olive gray clay containing rare shell fragments and some sandy clay. This lithofacies makes up most or all of the sequence at some sites. The second lithologic unit is heterogeneous being composed of several lithologies including sandy silt, sandy clay, clay and sand, and clay and silt. Lithologic unit 3 is composed of sand and silty sand with shell fragments and organic material concentrated near the top of the unit. The uppermost unit (lithologic unit 4) consists of olive gray clay and silty clay with some shell fragments. In a few borings the top of sequence 2 coincides with a thin soil horizon that is composed of tan and gray clay or brown and gray clay.

Sequence 2 is consistently at least 60 m thick except where influenced by diapirism. It is interpreted as principally a regressive fluvial-deltaic sequence on the basis of its upward-coarsening and upward-fining facies architecture. Fluvial channel fills in this sequence are about 38 m (125 ft) thick, which represents a large fluvial system comparable in depth to the modern Mississippi River. Fluvial deposits are mostly composed of fine sand rather than gravel or coarse sand; an indication of size sorting and the extreme downstream location of these deposits. Gravel is more common landward and to the east in southwestern Louisiana (Coleman and Roberts, 1988) where extrabasinal rivers constructed and later filled large channels. Furthermore, the fluvial sands are gray in color indicating a reducing environment of deposition, rather than tan or brown, which would indicate an oxidizing environment.

Sequence 3, which is the youngest complete sequence in the southern High Island Area, can be divided into two depositional settings that were controlled by sea level position, paleogeography, and contemporaneous deformation. Landward of the shelf margin the sequence is thin and onlaps broad depressions and troughs (figs. 6–8) that were created by late salt withdrawal and subsidence. Basinward of the shelf margin, the sequence progressively thickens (figs. 6, 7, and 9) as a result of progradation into deep water where high subsidence rates added new accommodation space. The thickest part of sequence 3 (fig. 10) can be associated with (1) the brow of the delta constructed during maximum progradation, (2) counterregional faults and adjacent bathymetric highs that acted as sediment dams (figs. 6 and 7), or (3) bathymetric lows created by local subsidence in salt withdrawal minibasins.

On the continental platform where sequence 3 is thin and fills broad sags, it is composed mostly of gray clay (fig. 7). Where the sequence is greatly expanded near the contemporaneous shelf margin, it is mostly composed of two lithologic units. The thick lower unit consists of olive gray clay with thin interlayered beds of sand and silt (figs. 7 and 9). The upper unit consists of gray sand and silty sand interlayered with thin beds of clay. In some extreme downdip locations the sand lithofacies is overlain by sandy clay. The clinoform reflections and upward-coarsening facies successions indicate that sequence 3 is a regressive sequence constructed by deltaic progradation during a lowstand in sea level.

Sequence 4 is composed of Holocene sediments, which range in thickness from 0 to 10 m (0–33 ft); however, most of the borings penetrate from 2.4 to 4.5 m (8 to 15 ft) of very soft olive gray clay and fine sand with variable concentrations of shell fragments. Thickness of these young reworked transgressive deposits is also partly controlled by recent deformation. Holocene sediments tend to be thin over recent structural uplifts (topographic highs) and thick in lows created by recent subsidence along reactivated faults. In some areas they are completely absent over topographic highs (figs. 6–9) and they are as much as 10 m thick; however, they are typically 3 to 6 m thick.

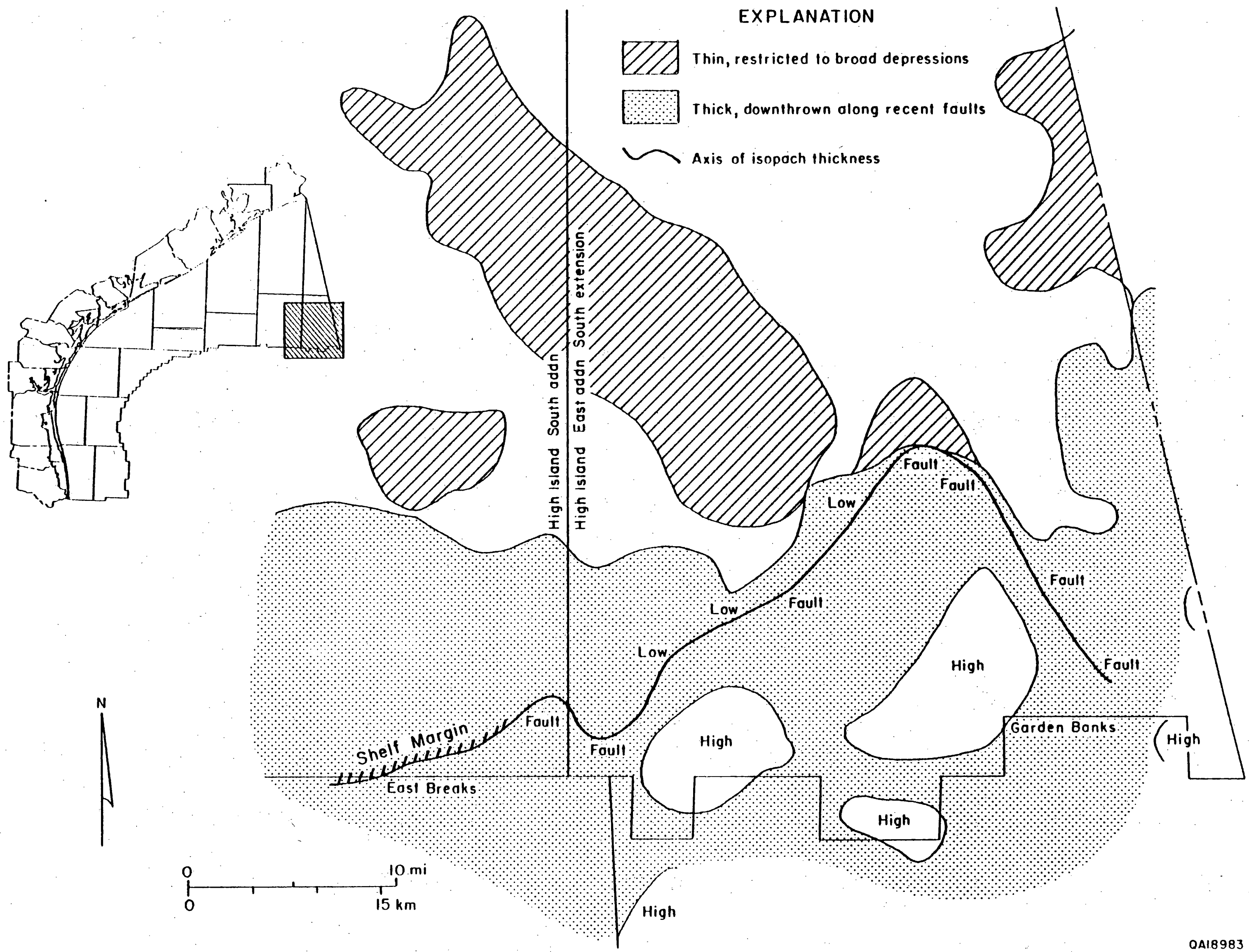

Figure 10. Distribution and relative thickness of the late Wisconsin stratigraphic sequence deposited during the isotope stage 2 lowstand in sea level.

Sequence Boundaries, Stratigraphic Surfaces, and Systems Tracts

The most prominent and continuous seismic reflections in the southern High Island Area separate strata of different ages and are used as sequence boundaries. These boundaries also approximate downlap surfaces (figs. 6–9), which operationally coincide with maximum flooding surfaces and are used to separate transgressive and highstand systems tracts (Posamentier et al., 1988). However in downdip settings the downlap surface may also coincide with the sequence boundary if rates of sedimentation were low during the rising phase and highstand in sea level. Therefore sequences 1–3 are interpreted as having been deposited during lowstands in sea level (lowstand systems tracts) whereas sequence 4, which is an incomplete sequence, represents transgressive systems tract deposits associated with the most recent rise in sea level (post-Wisconsin) and early highstand.

Within the lowstand sequences are numerous stratigraphic surfaces that represent local ravinement surfaces separating delta lobes. Subtle onlap and truncation surfaces within sets of clinoforms are interpreted as minor abandonment surfaces associated with allocyclic shifts in depocenters. Erosional surfaces, which are most prominent in sequence 2, are interpreted as margins of moderately large fluvial channels.

Sand-Body Geometries

In sequence 1, sand bodies occur near the middle and top of the sequence depending on position relative to the systems tract. Sands are thickest and most continuous in updip and middip positions whereas they become thinner and more interbedded downdip (figs. 6–9). Sand bodies display a similar pattern in sequence 2 except that most of the sand is concentrated near the top of the sequence (figs. 6–9). Sand tends to be lower in the sequence where deep channeling is indicated on the seismic profiles. Overall the sand bodies in sequence 2 are discontinuous and have highly variable thicknesses. Sequence 3 contains little sand except were it is expanded downdip of the *Trimosina* fault zone (fig. 9). There the sand bodies are near the top of the sequence and have variable thicknesses that are directly proportional to the rates of subsidence. The thin interbedded and discontinuous sand beds in the prodelta facies may be examples of shingled turbidites described by Vail and Wornardt (1991). In sequence 4, sand bodies are thin, highly discontinuous, and patchy (figs. 6–9). They represent reworked beach sands and possibly inner-shelf shoals that were constructed as the beach eroded.

Late-Wisconsin/Holocene Delta Flank Deposits (Brazos South Area)

Suter and Berryhill (1985) discussed origins and geometries of lowstand shelf-margin deltas in the northwestern Gulf of Mexico. One delta they identified was constructed by the combined Brazos-Colorado drainage systems (fig. 4). Morton and Price (1987) provided lithologic descriptions and interpreted sea-level histories of several sequences penetrated by foundation borings at the site of the Brazos-Colorado lowstand delta. Judging from isopach patterns of delta thickness (fig. 11), the foundation borings and seismic profiles available for this study are actually located updip and on the flank of the depocenter. This proximal delta flank setting provides a contrast to the lowstand delta sequences examined in the High Island Area.

Seismic Characteristics

Seismic reflection patterns of the oldest Brazos-Colorado sequences (<u>1, 2, 3</u> and older) are parallel or slightly divergent having variable amplitudes and continuities. Slight divergence of reflections causes gradual basinward thickening of some sequences (fig. 12A, B, and C). The sequences are also displaced by deep-seated growth faults of middle Miocene age (Morton et al., 1988). The oldest sequence boundaries (boundaries A and B) are onlapped by <u>sequences 2 and 3</u> (fig. 12A, B) whereas uppermost reflections in sequences 2 and 3 terminate by toplap against the upper sequence boundaries (boundaries B and C, figs. 12A and 13).

The youngest Brazos-Colorado sequences exhibit diverse reflection patterns compared to the simple patterns of the older sequences (figs. 12B, C and 14). <u>Sequence 4</u> displays moderate-amplitude, continuous, and parallel reflections that become low-angle clinoforms and downlap the lower sequence boundary near the shelf edge (fig. 12A, B). In contrast to this simple pattern, sequences 5 and 6 exhibit variable types of internal reflections. In <u>sequence 5</u>, continuous parallel reflections become low-angle clinoforms (fig. 12C) that downlap the lower sequence boundary near the shelf edge. However, to the southwest this same sequence is characterized by a broad mound composed of landward-dipping, low, wavy clinoforms (figs. 12B and 14). Broad low-angle clinoforms that are present near the crest of the mound progressively steepen landward and become irregular as the parasequence thins and onlaps the basal sequence boundary (boundary D on figs. 12 and 14). Reflections above the mound either onlap the base of the mound or drape over and are parallel to the mound. Seismic signatures of <u>sequence 6</u> are low-amplitude, discontinuous

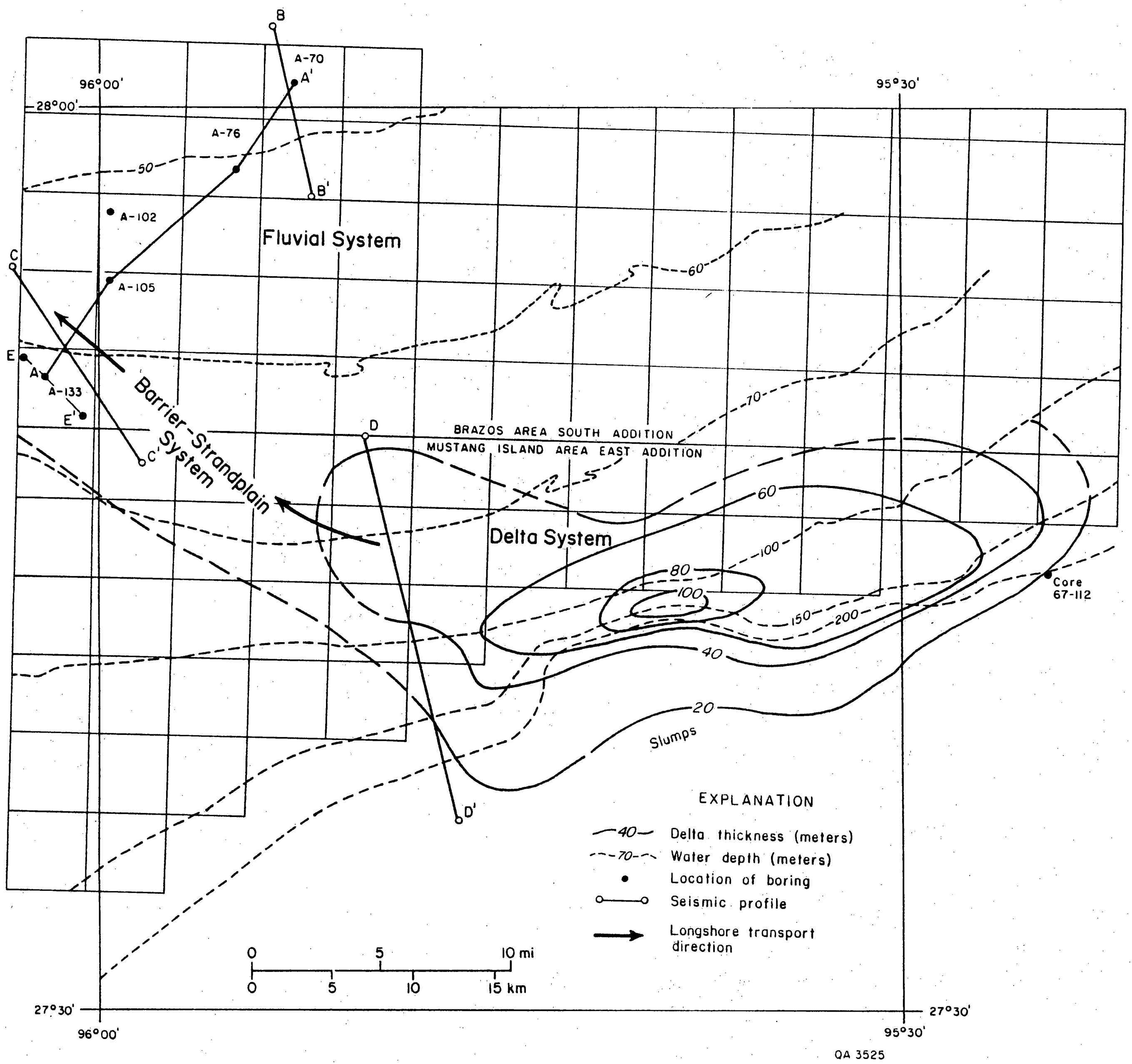

Figure 11. Locations of stratigraphic sections and seismic profiles near the late Pleistocene Brazos-Colorado fluvial-deltaic system. Thickness contours of the delta are from Suter and Berryhill (1985).

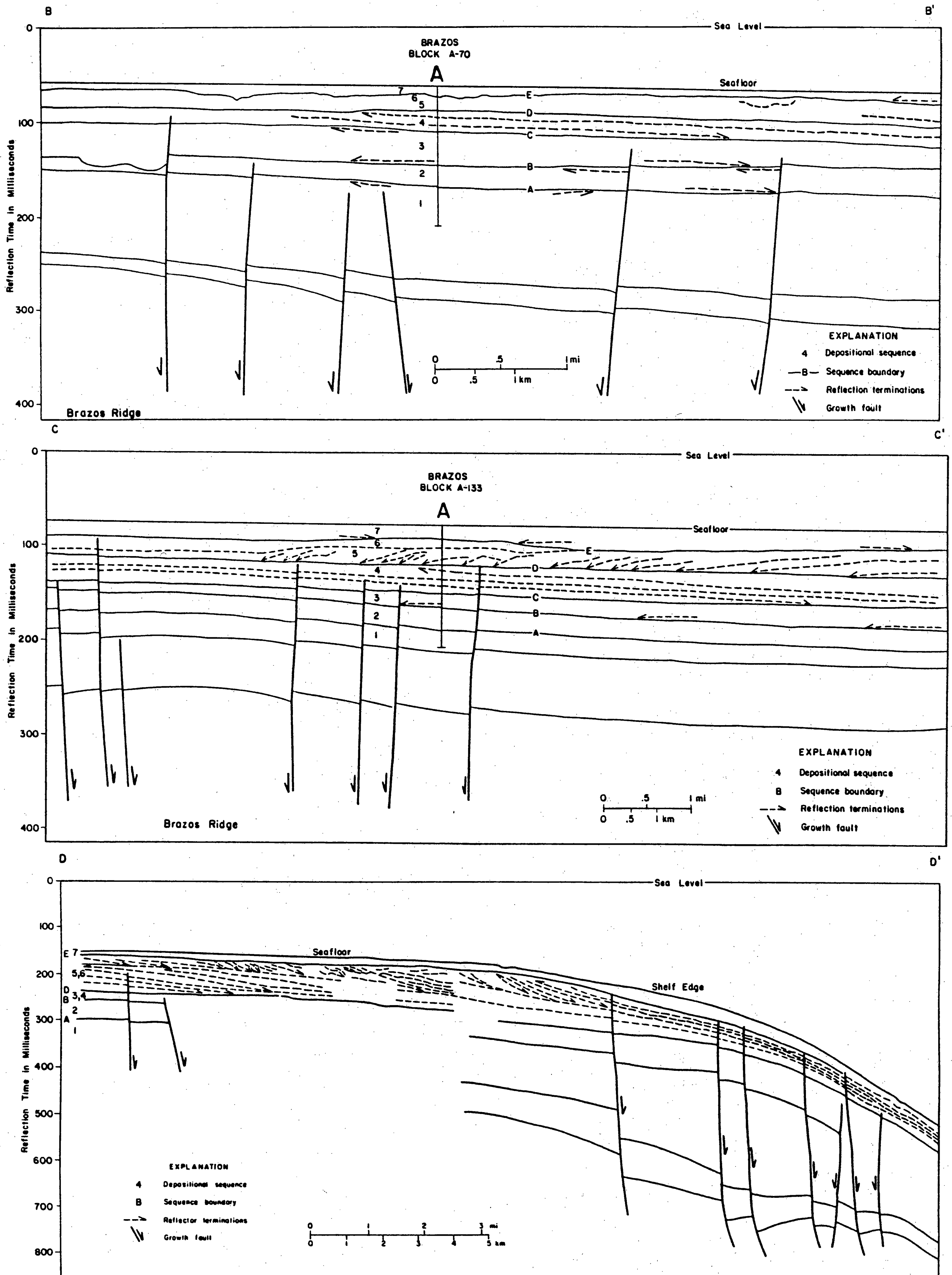

Figure 12. Late Pleistocene stratigraphic sequences near the Brazos-Colorado fluvial-deltaic system. Sequences are illustrated on line drawings derived from seismic profiles. Locations of diagrams are shown on figure 11.

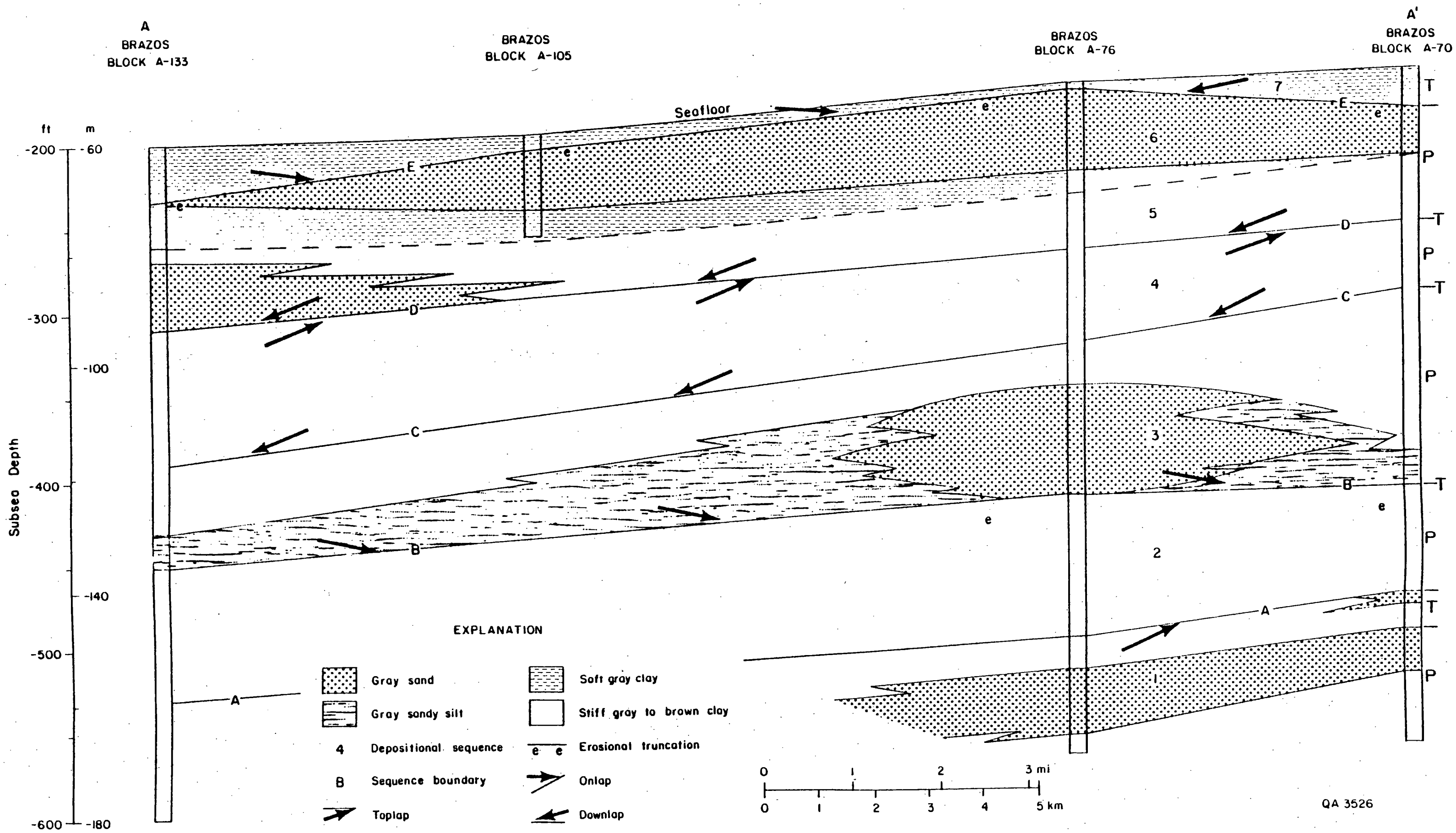

Figure 13. Lithologic composition of late Pleistocene stratigraphic sequences deposited by the Brazos-Colorado fluvial-deltaic system. Location of stratigraphic section is shown on figure 11.

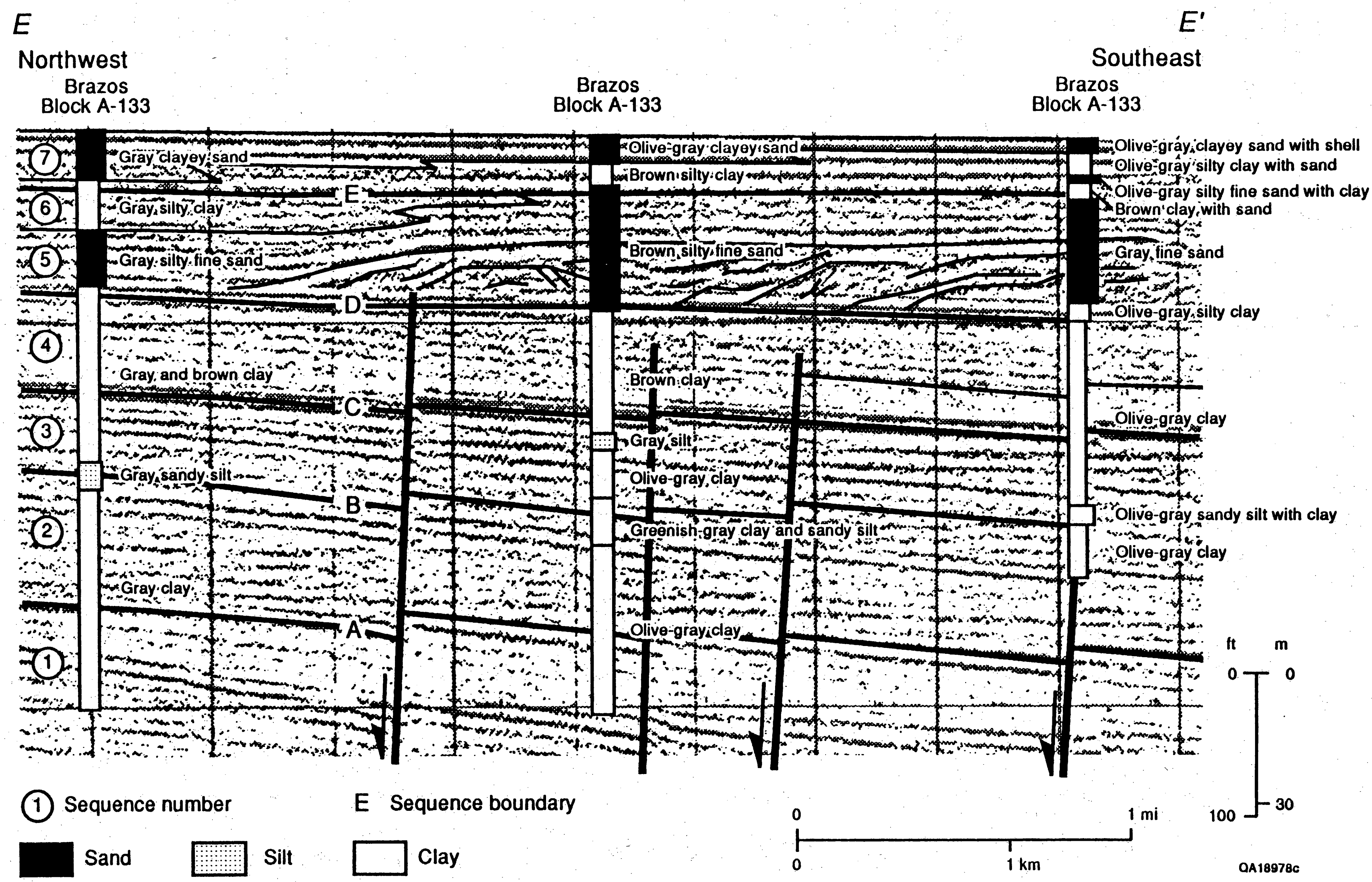

Figure 14. Detailed seismic stratigraphic and lithostratigraphic sections showing late Pleistocene sequences deposited by the Brazos-Colorado fluvial-deltaic system. General location is shown on figure 11.

reflections and wavy, or cut-and-fill patterns near Block A-70 that become oblique and sigmoid clinoform reflections farther downdip (fig. 12C). In all cases the uppermost reflections are truncated by the overlying sequence boundary (boundary E in fig. 12B, C). The top and bottom of sequences 5 and 6 coincide with seismic reflections, but the strong reflection boundary at the top of the mound separating the sequences does not coincide with a change in lithology (fig. 14). Strata within sequence 7 appear on seismic profiles as either reflection free or low- to high-amplitude parallel and continuous reflections that onlap the basal sequence boundary (boundary E, fig. 14).

Sequence Composition

There is only partial coincidence between strong seismic reflections and lithologic changes on the flank of the Brazos-Colorado shelf-margin delta (figs. 13 and 14). Seven lithostratigraphic and seismic stratigraphic sequences are identified from the seismic profiles, but only two lithologic boundaries (base and top of the principal sand unit) coincide with strong seismic reflections. Lithologies of the sequences, from oldest to youngest, are as follows.

Sequence 1 consists of three lithologic units. The lowermost unit is composed of very stiff to hard olive gray clay (figs. 13 and 14) that contains cemented nodules and some thin laminae and lenses of silt. This unit is overlain by gray clayey sand that grades upward into medium to coarse sand containing some gravel and shell fragments (fig. 13). The sand unit, which is about 12 m thick, grades basinward into and is overlain by very stiff to hard gray clay containing sand laminae and shell fragments; the shell fragments are rare and concentrated in thin isolated beds (fig. 13).

Vertical lithologic successions, lateral facies relationships, and seismic patterns suggest that sequence 1 is a regressive-transgressive couplet. Upward-coarsening textures within the lower two units and paleogeographic setting indicate progradation of a deltaic complex during a former lowstand in sea level. The overlying upward-fining unit was deposited during the subsequent transgression that accompanied subsidence and a relative rise in sea level.

Sequence 2 is composed entirely of very stiff gray clay (figs. 13 and 14) containing silt laminae, sand lenses, scattered shell fragments, and cemented nodules. The shell fragments do not occur at a particular stratigraphic position and are not correlatable among foundation borings. The uniform gray color, presence of shell fragments, and sand-filled burrows indicate deposition in a prodelta-shelf environment by a progradational delta system that was located landward of the study site. The regressive sequence is 20–25 m thick in a downdip position, but thins updip (fig. 12A).

Sequence 3 is subdivided into two lithologic units (figs. 13 and 14). The unit immediately overlying the sequence boundary is composed of light gray clayey and shelly fine sand that coarsens upward and grades laterally into gray sandy silt and silty clay containing sand lenses and shell fragments concentrated near the top. This lower unit is from 6 to 19 m thick; maximum thickness coincides with the sand-rich lithofacies. The upper lithologic unit consists of brown to gray silty clay containing sandy silt laminae, shell fragments, and cemented nodules. This upward-fining sequence is capped by a layer of tan or brown clay, which indicates an oxidizing environment and possible soil profile. Abrupt changes in engineering properties also coincide with the change in clay color. This poorly sorted unit also contains shell fragments, which are concentrated in thin beds.

Sequence 3 is a regressive unit exhibiting both upward-coarsening and upward-fining components. The coarser sediments were probably deposited as a distributary-mouth bar that merges downdip with prodelta and shelf deposits. The overlying finer-grained oxidized sediments appear to have a delta plain origin and represent the former base level of deposition. Overall, the deltaic sequence is 18 to 35 m thick. As expected, delta thickness is greatest near the distributary channel and decreases basinward and to the southwest (fig. 13). Absolute age of sequence 3 is uncertain, but its stratigraphic position suggests deposition occurred during the early Wisconsin falling sea level and lowstand (isotope stage 4).

Sequence 4 is composed entirely of very stiff olive gray and brown clay (figs. 13 and 14). The clay also contains cemented nodules and lenses of silty clay. The few thin laminae of shell fragments are associated with layers of sandy silt. The brown colors, coarser sediments, and shell debris are concentrated near the top suggesting deposition in progressively shallower water having higher wave energy. The downlapping seismic configuration, muddy lithology, and presence of shell all indicate a prodelta-shelf environment of a prograding shelf-platform delta. Stratigraphic position of sequence 4 suggests that it may have been deposited during the middle Wisconsin sea-level highstand (isotope stage 3).

Sequence 5 is composed of several lithologies (figs. 13 and 14). To the northeast the sequence is composed of stiff gray clay, which is interbedded with sandy silt and contains shell fragments. The mud-rich unit grades basinward into moderately well sorted gray and brown fine sand and silty sand (fig. 13). The sand unit is 12 to 27 m (40 to 90 ft) thick (fig. 13). The relatively mud-free middle succession is well sorted in contrast to the overlying poorly sorted silty sand. The sand, in turn, grades upward into soft gray clay and silty clay containing shell fragments and sand laminae. Correlative strata in a core from the eastern flank of the depocenter (fig. 11) and in 200 m of water are composed predominantly of mud containing minor amounts of sand in the lower part of the progradational sequence (Sidner et al., 1978).

Sequence 6 is composed primarily of two lithologic units the basal unit contains brown and gray clay and silty clay with thin sand beds (fig. 13). This muddy unit grades upward and landward into olive gray clayey sand with shell fragments and some detrital organic matter (figs. 13 and 14).

Sequences 5 and 6 comprise a single, large depositional complex. The locus of deposition outlined by isopach contours (fig. 11) depict a compound shelf-margin delta that attained a thickness of more than 100 m (Suter and Berryhill, 1985). Geometry, internal reflection patterns, and textures of the lower sand suggest reworking of the delta front and lateral accretion of sand transported northwestward by waves and strong longshore currents (figs. 11, 12B, and 14). Deposition on the flank of the delta was either as a submerged bar system or attached barrier island. A dip-aligned cross section (fig. 14), which essentially parallels the direction of sediment transport, illustrates the stratal configuration and lithofacies changes near the down-current terminus of sand deposition. The absence of compelling evidence indicating subaerial exposure and stratigraphic correlation with the delta-front deposits suggest that the preserved sand shoal was a subaqueous feature. This delta-margin shoreface deposit was subsequently buried by prodelta mud and fluvial-channel or distributary-mouth bar sand of a younger delta lobe. Sand deposited near the channel mouth is as much as 16 m thick (fig. 13, Block A-76).

The mud of unit 5 records a change in depositional environment and a decrease in wave energy associated with continued overall upward-shoaling as delta progradation advanced toward the shelf margin. The mud may represent an interdistributary bay or similar subaqueous delta-plain environment where fine-grained sediment accumulated after the coastal plain flooded.The combined regressive unit (sequences 5 and 6) was probably deposited during a middle Wisconsin falling sea level and subsequent late Wisconsin lowstand (isotope stage 2) that exposed the delta and promoted progradation of the shelf margin. The landward segment of the upper sequence boundary (boundary E, fig.12A) probably represents the subaerial erosional unconformity formed during the most recent sea-level lowstand about 18,000 yr B.P.

Sequence 7, which is the youngest sequence, is mainly composed of very soft gray clay, sandy clay, or poorly sorted clayey sand containing shell fragments and organic matter (figs. 13 and 14). Overall the sequence exhibits an upward decrease in abundance of sand. Sand-filled burrows as well as sand and silt laminae are common even though mud is the predominant sediment type except on the flank of the former delta (fig. 14). The upward-fining succession, poor sorting, and stratigraphic position of sequence 7 all point to transgressive nearshore sediments deposited as sea level rose and the shoreline retreated rapidly.

Sequence Boundaries, Stratigraphic Surfaces, and Systems Tracts

Above and updip of the Brazos Ridge, the upper boundary of sequence 2 is an erosional unconformity (fig. 12A) that, together with the cemented nodules and shell fragments, suggests shoreface retreat during a rising sea level. Thus, the sequence boundary actually represents a submarine erosion surface or ravinement surface (Swift, 1968).

The seismic stratal patterns, lithologies, and vertical facies successions all indicate that most of the sequences penetrated by the borings were deposited by lowstand systems tracts. Strata at the very top of the lowstand deposits probably represent transgressive systems tracts, but these transgressive deposits are very thin either because sediment supply was drastically reduced during the rising phase of sea level or submarine erosion removed some of the early transgressive systems tract deposits. In either case, sequence boundaries and downlap surfaces are the same features.

Thin layers of sandy silt containing shell fragments in a thick succession of mud probably represent local condensed sections. These slightly coarser sediments form when rates of shelf sedimentation are low and the shallow bottom sediments are frequently reworked by strong storm waves and currents.

Regional condensed sections and maximum flooding surfaces are not evident from the foundation boring descriptions and there is a lack of correlatable shell beds at or near the interface between regressive deposits and transgressive surfaces (figs. 13 and 14). One exception is sequence boundary E and overlying sequence 7, which is interpreted to be transgressive deposits of the late Wisconsin-Holocene rise in sea level.

Sand-Body Geometries

Both cross sections (figs. 13 and 14) illustrate thicknesses and geometries of sand bodies in sequences 1, 3, 5, 6, and 7. Sand bodies are absent in sequences 2 and 4 at the study site. Although cross section A-A' is slightly oblique to depositional strike it provides an estimate of the lateral extent of each sand body. None of the sand bodies extends the entire length of the cross section (approximately 24 km) and none maintains uniform thickness between adjacent borings (fig. 13). However, the section is not long enough in either direction to determine the entire lateral extent for each sand body.

The sand in sequence 1 is 6 to 12 m (20 to 40 ft) thick and appears to lose sand from the bottom. The upper surface is at the same stratigraphic position but the base rises to the northeast. The presence of coarse sand and gravel indicate proximity to a river channel or distributary-mouth bar, which is consistent with the sand body contacts and geometry.

The sand body in sequence 3 is 10 to 20 m (35 to 65 ft) thick and extends approximately 19 km (12 mi) along strike. It exhibits upward-coarsening textural trends and it pinches out to the northeast and southwest (figs. 13 and 14). The lenticular geometry, grain-size distribution, and presence of shell are strong evidence of the distributary-mouth-bar origin of this sand. The thickest section of clean sand is penetrated in the Block A-76 boring, which is probably near the river mouth.

The sand in sequence 5 (figs. 13 and 14) is also lenticular in cross section, coarsens upward, and contains shell fragments. However, its paleogeographic position and landward dipping internal reflections illustrate lateral accretion processes onto a surface such that the sand diminishes from the top and to the northwest and northeast. The lower sand body extends no more than 10 km (6 mi) along strike.

Sand body thickness of sequence 6 cannot be determined accurately because its top was eroded during the Holocene transgression. Detailed plots of foundation borings in Blocks A-102 and A-105 show as much as 5 m (17 ft) of relief on the submarine erosion surface. Maximum preserved thickness of the sand body is about 15 m (50 ft). The composite thickness of sand in sequences 5 and 6 is from 6 to 25 m (20 to 85 ft) and represents parasequences that were stacked when the axis of deposition shifted. This allocyclic lobe switching is illustrated in the change in clinoform stratal patterns between sequence 5 and sequence 6 near the delta terminus (fig. 12C). Unlike the sand body in sequence 6, the other sand bodies were protected from erosion by muddy strata (figs. 13 and 14).

The sand body in sequence 7 is thin, patchy, and only locally present in Block A-133 (fig. 14). Characteristics of this transgressive deposit suggests that the rate of shoreline retreat in the area was rapid and supply of coarse sediment was low. Thus, shorezone deposits are only a few meters thick and discontinuous.

Late-Wisconsin/Holocene Interdeltaic Deposits (Mustang Island Area)

Sediments deposited between the lowstand Rio Grande and Brazos-Colorado deltas are recognized on the basis of sea-floor morphology, lithofacies associations, and seismic reflection characteristics. The interdeltaic setting is a broad, low-gradient platform that forms a ramp, which is lower than the adjacent delta lobes (fig. 15). Along the ramp the shelf margin is poorly defined and the continental shelf gradually grades into the continental slope. This gradual basinward tilt is only disrupted by a series of small pinnacle reefs that are partly or completely buried by younger sediments (fig. 16). The carbonate reefs form an arcuate trend that coincides with a zone of reactivated middle Miocene expansion faults (Morton et al., 1988). Natural gas seeps along the faults and reef positions with respect to the faults suggest that reef building may have been stimulated by chemosynthetic processes similar to those described by Roberts et al. (1989).

Seismic Characteristics

Seismic reflections in the interdeltaic deposits are parallel, have variable amplitudes and continuities, and uppermost reflectors are also parallel to the sea floor (fig. 16). The most prominent unconformable reflection in the interdeltaic strata is located about 15 m below the sea floor. This unconformity truncates underlying strata, is onlapped by overlying strata, and merges with the seismic expression of carbonate reefs (fig. 16). This erosional unconformity is interpreted as the late Pleistocene lowstand surface that was modified by submarine erosion during the Holocene transgression.

Most of the seismic reflections below the unconformity are high-amplitude, parallel, and continuous suggesting uniform lithologies. Minor disruptions and offsets in these reflections are caused by small faults that were recently reactivated by basinal extension.

Sequence Composition

The interdeltaic deposits are composed of very soft to soft olive gray clay (figs. 16–19). Water saturation decreases with depth and the softest sediments are typically 7.5 to 12 m (25 to 40 ft) thick. Interdeltaic sediments are uniformly muddy and any sand

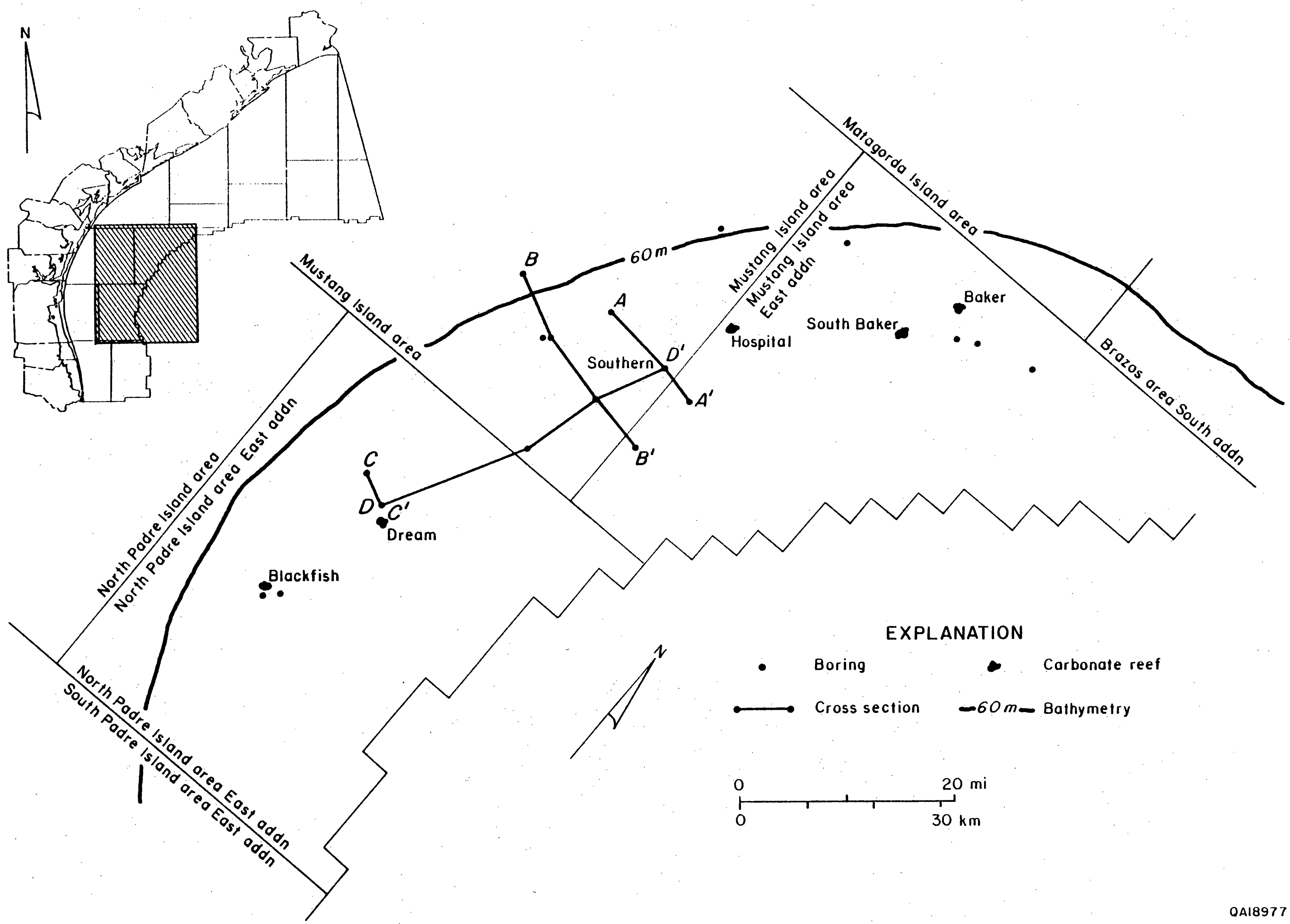

Figure 15. Locations of stratigraphic sections and seismic profile between the Brazos-Colorado and Rio Grande fluvial-deltaic systems.

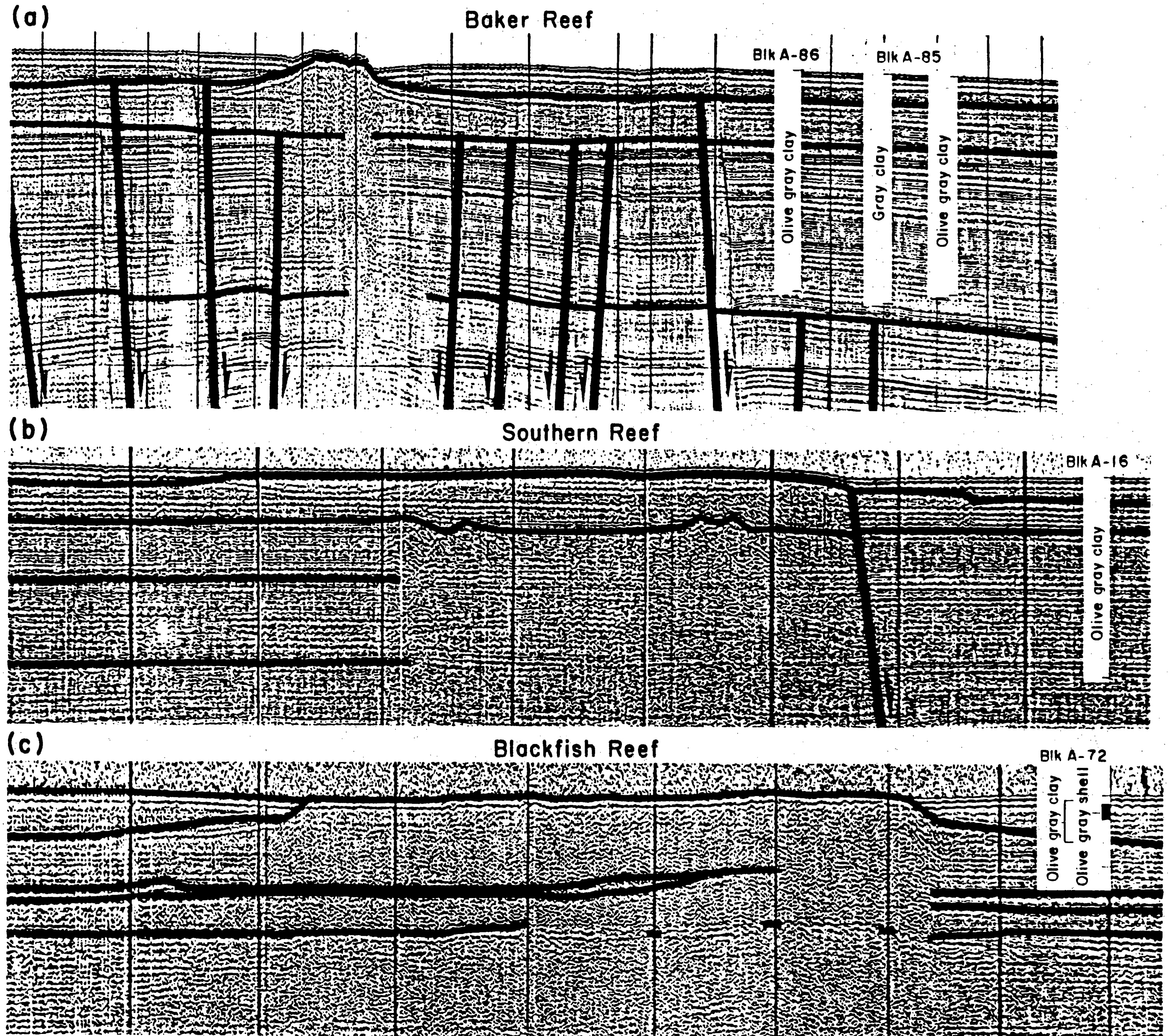

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Figure 16. Stratigraphic sections illustrating seismic stratal patterns and lithologies in the interdeltaic setting near (A) Baker Reef, (B) Southern Reef, and (C) Blackfish Reef. Locations shown on figure 15.

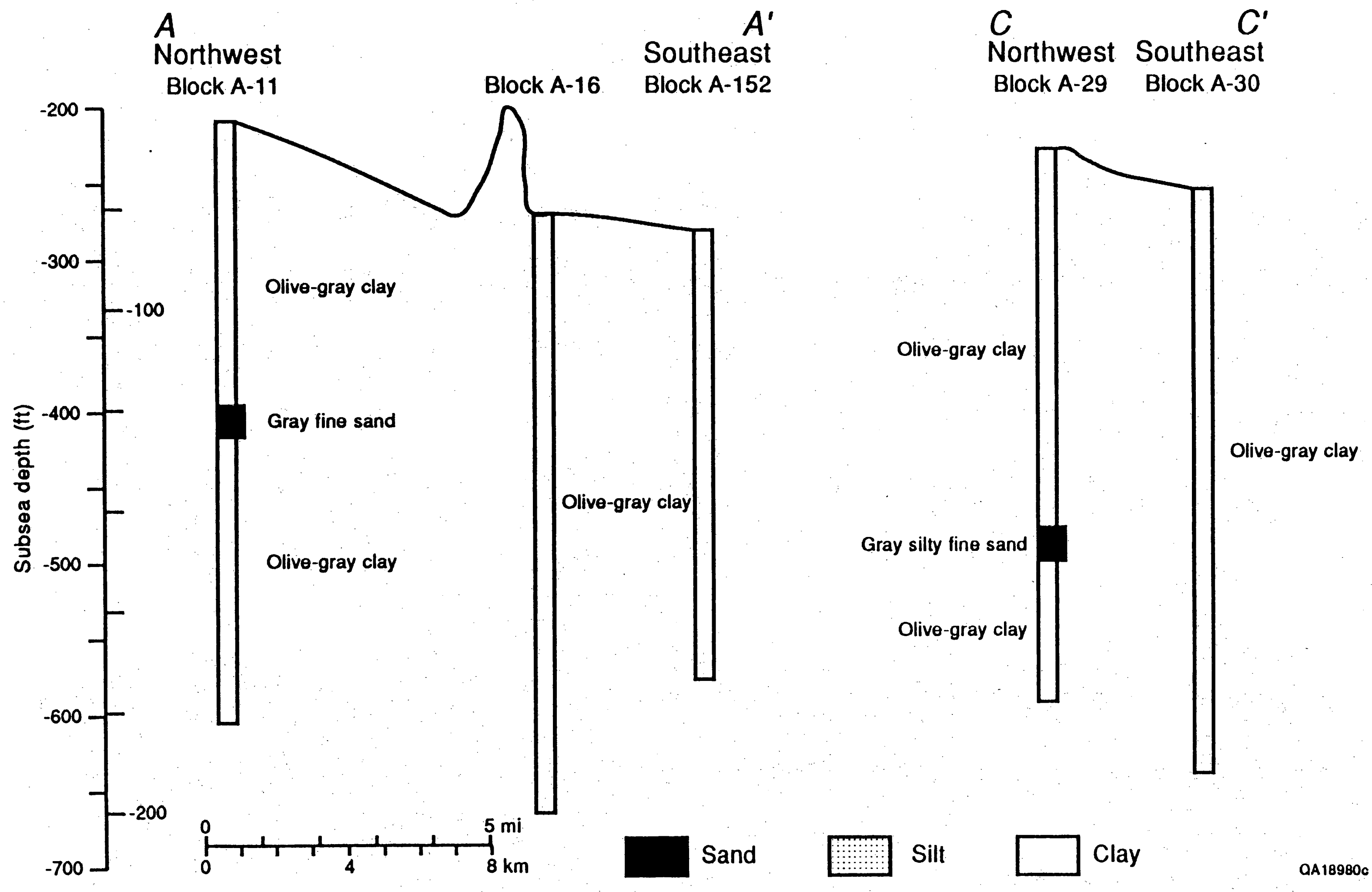

Figure 17. Dip sections A-A' and C-C' showing late Pleistocene and Holocene stratigraphic sequences deposited between the Brazos-Colorado and Rio Grande lowstand deltas. Location shown on figure 15.

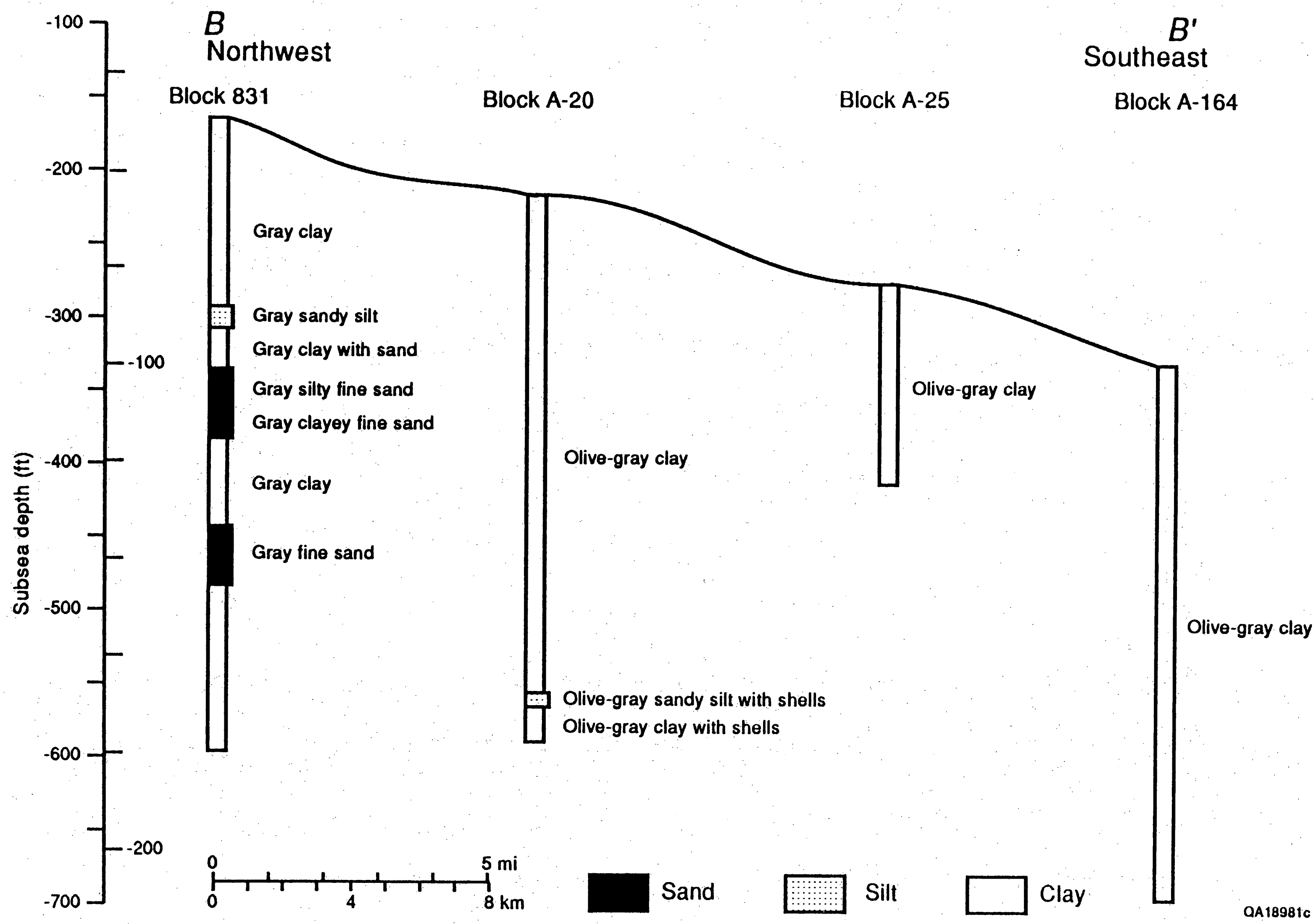

Figure 18. Dip sections B-B' showing late Pleistocene and Holocene stratigraphic sequences deposited between the Brazos-Colorado and Rio Grande lowstand deltas. Location shown on figure 15.

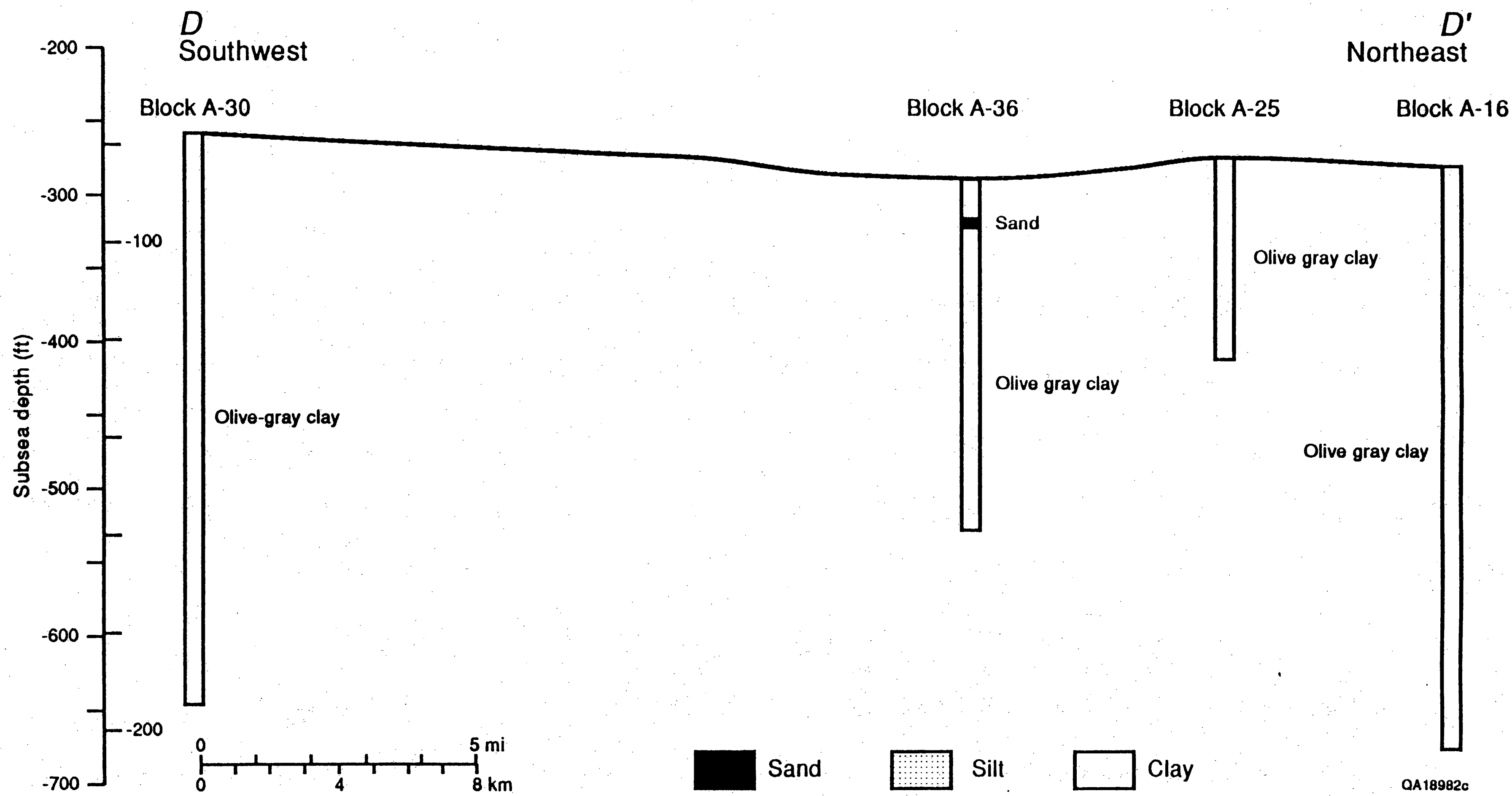

Figure 19. Strike section D-D' showing late Pleistocene and Holocene stratigraphic sequences deposited between the Brazos-Colorado and Rio Grande lowstand deltas. Location shown on figure 15.

beds are rare and discontinuous (figs. 16–19). The muddy deposits contain very little shell except for local shell beds concentrated around the carbonate banks where they are composed of reworked reef detritus (fig. 16C). These reworked carbonate sediments probably represent either storm deposits or condensed intervals formed during a temporary stillstand in sea level when sediment influx was low especially compared to carbonate production associated with reef building. The few shell layers, which are thin and patchy, cannot be correlated between foundation borings with confidence and they do not coincide with a particular seismic reflection.

The thicknesses of pre-Holocene sequences at this site are unknown because they cannot be differentiated on seismic profiles. Judging from the depositional setting, sequences probably are thin and also composed of olive gray clay with rare, thin, and discontinuous layers of sand and shell hash (figs. 17–19).

On the South Texas outer continental shelf, between the late Wisconsin lowstand deltas and landward of the pinnacle reefs, Holocene and modern transgressive sediments are as much as 37 m (120 ft) thick (Berryhill and Trippet, 1980, 1981). These young marine deposits are only a few centimeters to a few meters thick over much of the remaining Texas continental shelf. At least the upper few meters of these deposits are aggradational nearshore sediments composed of intercalated and bioturbated sand and mud. The predominance of either the sand or mud constituent depends mainly on water depth and distance from the shoreline. Percent sand and the number of discrete sand layers are extremely low over the outer shelf; whereas intercalated sand and mud occur near the shoreface and over the ancestral deltas (Berryhill and Trippet, 1980, 1981).

The sedimentologic characteristics and young ages of these uppermost shelf sediments suggest episodic deposition related to storm processes that periodically transported shoreface sands basinward below normal wave base. The substantial thickness of these paralic sediment above the unconformity indicates that accumulation rates of middle and outer shelf strata can be significant during transgressive and highstand phases of sea level.

Sequence Boundaries, Stratigraphic Surfaces, and Systems Tracts

Sequence boundaries, stratigraphic surfaces, and eustatic systems tracts are not easily identified in the interdeltaic deposits. This is because seismic reflections and lithologies are homogeneous and there are no obvious breaks in the physical stratigraphic record except moisture content. This ramp depositional setting of the shelf margin has received only minor amounts of fine-grained sediments during lowstand, transgressive, and highstand phases of sea level, consequently the lithologic and seismic records are incapable of differentiating among sea-level phases of deposition and erosion. Additional independent analyses such as isotopic age dating, high-resolution biostratigraphy, or palynology and paleoecology will be necessary to accurately determine the chronostratigraphy of the mudstones.

Without sequence boundaries to subdivide the depositional cycles, it is impossible to differentiate between transgressive and regressive muddy deposits. Furthermore, the lack of shell bed continuity suggests that condensed sections in these sediments are not represented by carbonate-rich deposits as reported by Coleman and Roberts (1988). Rates of sediment accumulation were undoubtedly lower away from the adjacent deltaic depocenters and therefore condensed sections should be accentuated in the interdeltaic setting rather than suppressed. This apparent difference in geological conditions between Texas and Louisiana probably is not an artifact of the boring descriptions because (1) both studies relied on foundation borings and (2) descriptions of borings in other areas do contain references to numerous shell beds.

The great thickness of muddy interdeltaic sediments, their physical properties, the water depth where they occur, and the magnitude of lowered sea level during the late Wisconsin glacial stage (Fairbanks, 1989) all suggest that mud was subaerially exposed when sea level fell (isotope stages 4–2) and any soil profiles developed during the lowstands were removed by erosion as sea level rose and flooded the former coastal plain. The only lithologic evidence of the unconformity is the contrast between soft and firm mud. Firm mud represents overconsolidated conditions that developed when the sea floor was exposed whereas soft muds represent marine shelf sediments deposited above the ravinement surface by alongshelf currents.

Sand-Body Geometries

In the interdeltaic setting, sand bodies are rare, thin, and either have patchy or lenticular geometries (Table 2). More than 2,200 km^2 (850 mi^2) of the South Texas shelf platform is underlain by at least 120 m (400 ft) of these predominantly muddy sediments (figs. 15–19). Along depositional strike and updip the muddy shelf sediments correlate with sequences containing interbedded mud and sand deposited in shallow-water, high-energy environments (figs. 17 and 18).

The interdeltaic area received a substantial volume of sediment transported along-shelf and across-shelf from adjacent areas and deposited in a broad sediment sink. The area did not receive substantial volumes of sand even during lowstands in sea level. This is an anomalous condition considering that during highstands, the interdeltaic shorelines are commonly characterized by thick well-sorted sand bodies deposited as barrier islands or strandplains (Morton et al., 1983; Morton and Price, 1987).

In this mud-dominated system, shelf deposits are thin interlaminated sand and mud, Each couplet represents a storm deposit that accumulated on the sea floor when the shoreline was slightly landward of its present location. Now the shelf is a depositional site of only fine-grained material transported in the nepheloid layer across-shelf by large oceanic currents (Sahl et al., 1988). Some mud is transported westward by flow from the Mississippi River, but much of it represents resuspension of mud from the shelf northeast and south of the study area.

Late Pleistocene and Holocene Highstand Depocenter (South Padre Island Area)

Foundation borings near the Rio Grande penetrate three late Quaternary stratigraphic sequences (figs. 20–22). Each sequence is composed of upward-coarsening fluvial-deltaic sediments that were deposited during different phases of sea level flux. Comparing the facies distributions, lithologic characteristics, and sand-body geometries of these three fluvial-deltaic systems reveals how differences in nearshore processes, sediment supply, and relative sea-level changes can influence stratal stacking patterns and reservoir continuity within each sequence.

Seismic Characteristics

Two types of seismic reflection patterns are recorded in late Quaternary sequences in the Rio Grande Embayment. Most of the reflections at moderate depths are high-amplitude variable continuity and slightly divergent so that each sequence thickens basinward (fig. 21C). In contrast, the shallowest sequences are recognized by sets of low-angle oblique clinoforms. The clinoforms downlap onto continuous high-amplitude reflections that separate the sequences and represent the sequence boundaries (fig. 21C).

Sequence Composition

Late Pleistocene Sequences - The oldest sediments penetrated by the foundation borings (sequence 1, fig. 21B) are hard greenish gray clays that are at least 11 m thick (35 ft). These muddy sediments are interpreted as prodelta deposits on the basis of composition and associated clinoform reflections (fig. 21B and C). The hard clays are

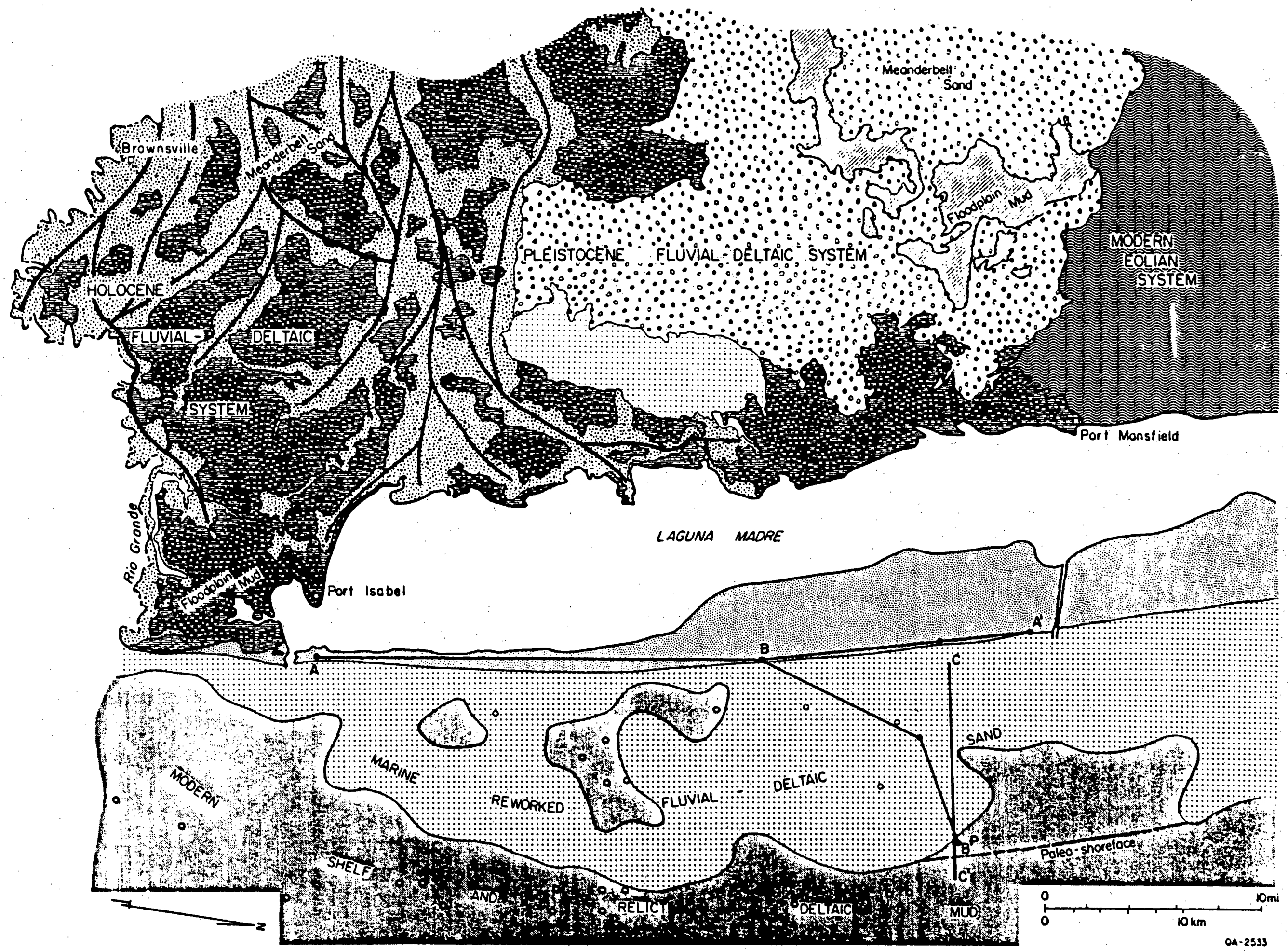

Figure 20. Geologic map showing the distribution of late Pleistocene and Holocene Rio Grande deltas. Locations of stratigraphic sections and seismic profiles are also shown. Onshore depositional environments were modified from Brown et al. (1980).

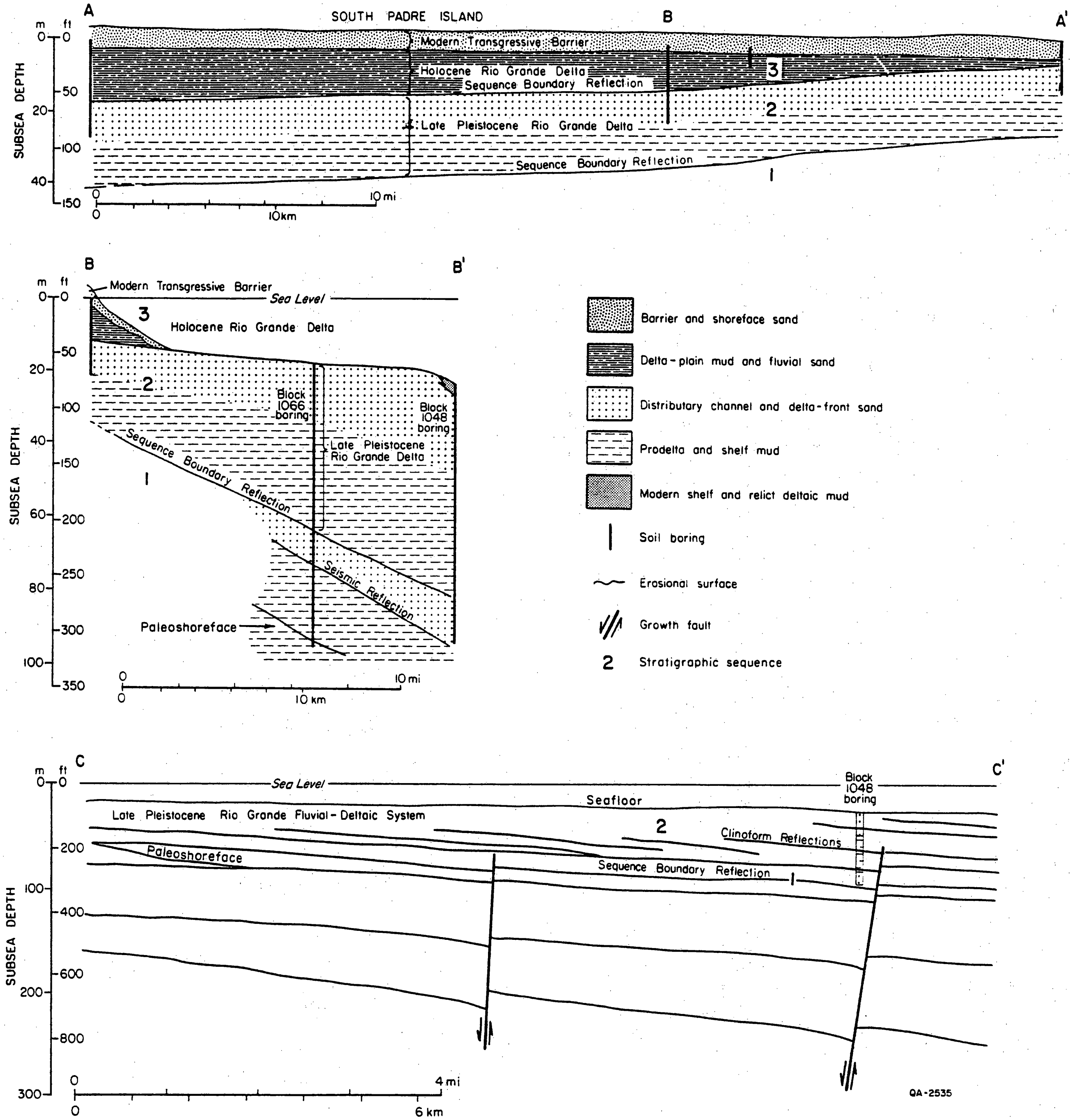

Figure 21. Late Pleistocene and Holocene stratigraphic sequences deposited by the Rio Grande fluvial-deltaic system. Sequences are illustrated on stratigraphic sections (A and B) and on a line drawing derived from a seismic profile (C). Locations of diagrams are shown on figure 20.

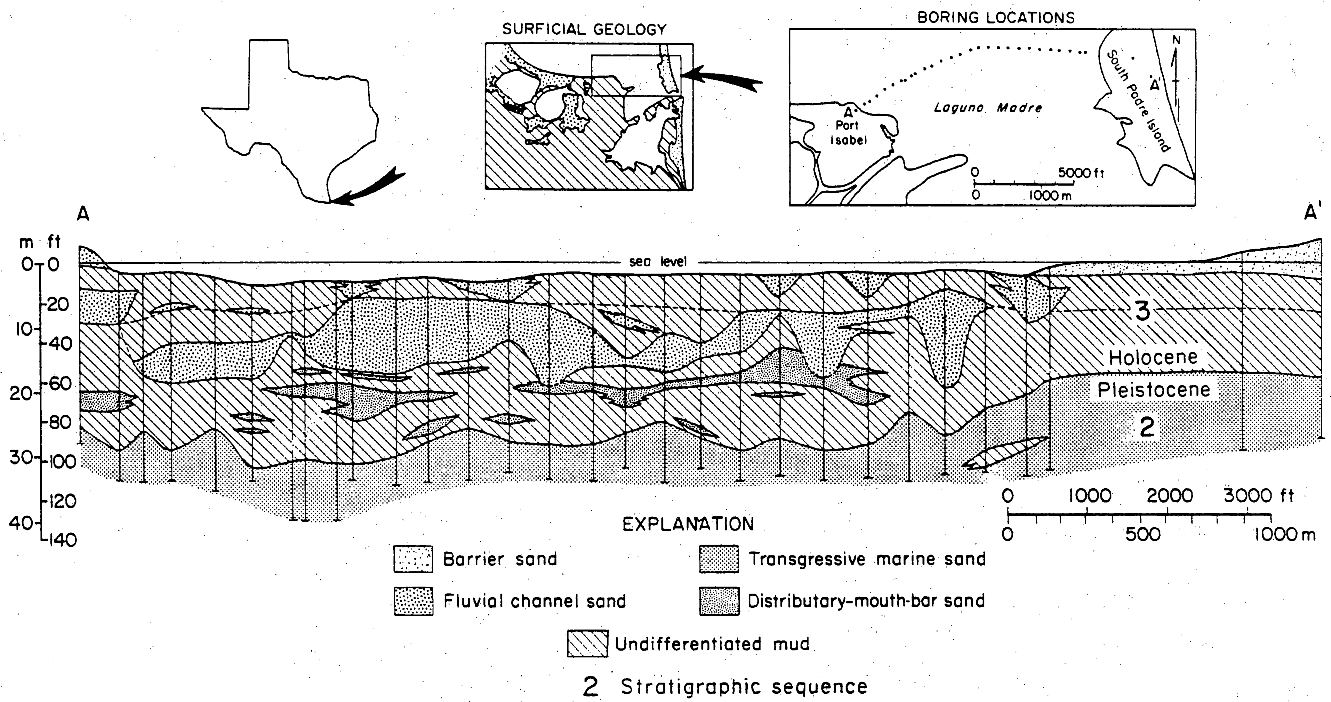

Figure 22. Detailed stratigraphic section illustrating late Pleistocene and Holocene sequences deposited by the Rio Grande fluvial-deltaic system.

overlain by gray silty sand and silty clay that grade upward into gray sand. Total thickness of the sand is 21 m (70 ft) and the top of the sand coincides with the sequence boundary (fig. 21B and C), which is also a high-amplitude seismic reflection.

Downlapping seismic reflections above the sequence boundary, which occur beneath much of the south Texas inner shelf, are interpreted as deltaic sediments (sequence 2) deposited during the last Pleistocene highstand in sea level (isotope stage 5). The basal lithologic unit of sequence 2 is about 30 m (100 ft) thick and is composed of stiff gray silty clay and sandy silt. These prodelta and shelf deposits are recognized by clinoform reflections that dip basinward at 1.5–2.0° (fig. 21C). The prodelta muds are overlain by gray and tan silty fine sands that are about 15 m (50 ft) thick. The sands, which were deposited in delta-front and fluvial-channel environments form the sea floor over much of the adjacent inner shelf (figs. 20 and 21). The coastal plain sequence of thick fluvial-channel and delta-front sands passes landward into floodbasin and delta-plain muds (Morton and McGowen, 1980). Across the South Texas continental shelf, the top of the sand lithofacies also represents a sequence boundary that coincides with a high-amplitude seismic reflection. This reflection marks the Holocene-Pleistocene unconformity that was modified by submarine erosion during the isotope stage 1 transgression.

Holocene-Modern Sequence – The Holocene-Modern Rio Grande delta is one of the two largest delta systems constructed in the northwestern Gulf of Mexico. These young Rio Grande deposits and suprajacent barrier deposits (figs. 21 and 22) can be subdivided into six lithofacies that together constitute a moderately thick regressive sequence (sequence 3). Five of the lithofacies are associated with the delta complex whereas the sixth lithofacies represents the overlying reworked strandline facies.

The basal lithofacies is composed of brown to gray clays and sandy clays that were deposited in a prodelta environment. These prodelta muds, which are 3 to 11 m thick, grade upward into gray thin sands interbedded with fine-grained sediments similar to the prodelta muds. These alternating sands and muds (second lithofacies) comprise the delta-front deposits that are as much as 7.5 m thick (fig. 22). Sediments above the delta-front are relatively thick (11 m) and are composed of either brown to gray sandy clay and organic-rich clay (lithofacies 3) or gray fine sand and sandy silt (lithofacies 4). The sandy and carbonaceous clays were deposited on the delta-plain in natural-levee and flood-basin environments whereas the sand-rich sediments, which are in erosional contact with the underlying delta-front deposits, were deposited in a fluvial-channel environment. Together these two lithofacies account for most of the Holocene fluvial-deltaic system. Strata above the fluvial channels (lithofacies 5) are about 6 m thick and are composed of silty clays containing narrow lenses of sand (fig. 22). Surficial features on the extant coastal plain indicate that these sediments were deposited by a recent delta that prograded into extremely shallow-water (Fulton, 1975). The sixth lithofacies is well-sorted fine sand that is 1.5 to 4.5 m (5 to 15 ft) thick. The sand coincides with South Padre Island, a transgressive barrier that is migrating across the former delta plain as a result of reduced sediment supply and a relative rise in sea level.

Sequence Boundaries, Stratigraphic Surfaces, and Systems Tracts

The age of sequence 3 is uncertain because the countdown method of chronostratigraphic correlation is unreliable for older deposits especially in positions far updip of the shelf margin. Assuming that sequences 1 and 2 were deposited by highstand systems tracts after major interglacial flooding events, then sequence 1 may be isotope stage 7 deposits and the isotope stage 6 regressive sediments are absent because of nondeposition or deposition and subsequent erosion during the isotope stage 5 transgression.

Sequence 2 is interpreted as having been deposited during the Sangamon highstand and earliest Wisconsin falling phases of sea level (late isotope stage 5 and early isotope stage 4). The tan color of these highstand systems tract deposits indicate subaerial exposure as sea level fell. Iron-bearing and oxygen-enriched ground water percolating through the porous sand also may have contributed to the tan sediment color. Furthermore, the uppermost sands may have been reworked by eolian processes when sea level was low (isotope stages 4, 3, and 2). Sediments deposited during the Wisconsin lowstands are confined to the shelf margin (see discussion of High Island Area) or within entrenched valley fills.

The boundary separating sequences 2 and 3 is a composite surface that originated as an erosional unconformity and sediment bypass surface where no lowstand systems tract was deposited. The erosional unconformity exhibits substantial relief (fig. 22) probably because of close proximity to the entrenched valley that was formed by the Rio Grande during the late Wisconsin lowstand in sea level. Although the location of the entrenched valley and its influence on the observed delta thickness are presently unknown, it is clear that the channel axes of both Holocene delta systems were controlled by topography inherited from the erosional unconformity (fig. 22).

As the former coastal plain and entrenched valleys were transgressed, the upper sand surfaces of sequences 1 and 2 were reworked by waves and longshore currents that formed marine erosion surfaces. The generalized lithologic descriptions prevent precise differentiation between the older highstand systems tract sands of sequences 1 and 2 and the transgressive systems tract reworked sands of sequences 2 and 3. The coincidence of the sequence boundaries and the downlap surfaces (figs. 21 and 22) suggests that transgressive systems tract deposits of sequences 2 and 3 are thin and possibly below the level of seismic detection. In the Block 1048 boring, shell fragments reported in the gray clay immediately above the upper boundary of sequence 2 are probably within the transgressive systems tract deposits, but shells are not reported at the same position in the Block 1066 boring. The discontinuous nature of shell deposits suggests that this is not a widespread condensed section and it does not necessarily coincide with the surface of maximum flooding.

The upper boundary of sequence 2 is currently being modified by shoreface and shelf processes that are creating a major ravinement surface where the Holocene Rio Grande delta is being eroded by shoreface retreat (fig. 21B). Another minor flooding and ravinement surface produced by submarine erosion probably separates the early and late Holocene deltaic parasequences (fig. 22).

Sequence 3 is as much as 30 m thick (fig. 21A and 22) and sediment thickness increases away from the overlapping contact with adjacent Pleistocene sediments and toward the river mouth where subsidence was greatest. Sand concentrations in the thickest part of the wedge range from 20 to 50 percent depending on the thickness of channel-fill deposits. Aggradation of upper delta-plain strata and fluvial channels accompanied delta progradation across the inner shelf about 4–7 ka when the rate of sea-level rise diminished. The thin progradational facies (prodelta and delta-front deposits) and thick aggradational facies (channel-fill and delta-plain deposits) of sequence 3 are typical stratal stacking patterns of fluvially-dominated shallow-water-deltas that build onto stable platforms. The older Holocene parasequence is relatively thick because accommodation space was being created by the rise in global sea level and continued subsidence. In contrast the youngest Holocene parasequence is thin because the slow relative rise in sea level limited accommodation space.

The Holocene-Modern fluvial-deltaic sediments (sequence 3) were deposited during the most recent rising phase and highstand in sea level (isotope stage 1). Practical limitations of identifying eustatic systems tracts is revealed by examining the depositional response of a secondary deltaic system where sediment supply has decreased and both progradational and retrogradational parasequences have formed. When sediment supply was abundant, a progradational parasequence was deposited that downlapped onto the sequence boundary. After sediment supply decreased and the delta system continued to subside then the distal delta deposits were eroded and a transgressive barrier-lagoon system formed as the delta flank was reworked. The transgressive surface of erosion, which underlies the transgressive barrier island, continues to retreat during each erosional (storm) event.

The moderately thick progradational parasequence of the delta system could be interpreted as an early highstand systems tract or as a late-stage transgressive systems tract. The overall retrogradational pattern and delta geometry, which is similar to the proximal delta-flank position of the Mississippi delta (Morton, 1991), suggest that the basal progradational parasequence is part of a transgressive systems tract.

Sand-Body Geometries

The late Pleistocene and Holocene coastal plain and continental shelf in the Rio Grande Embayment were constructed by both fluvial-deltaic and marine systems (Brown et al., 1980; Fulton, 1975). Consequently the distribution of sand bodies within the respective sequences are substantially different depending on the predominant influence of rivers or waves.

Sequences 2 and 3 show remarkably different sand-body geometries that are related to paleogeographic settings and sea-level histories. The upward-coarsening prodelta and delta-front succession of sequence 2 is 15 to 80 m thick with thicknesses increasing offshore and toward the fluvial axis (fig. 21). Strike-aligned isolith patterns as well as lateral continuity and abundance of sand strongly indicate that this delta system was dominated by wave energy. The sand lithofacies of this wave-dominated delta system can be traced laterally for more than 50 km (figs. 21 and 22). The relatively steep angle and height of the clinoforms (fig. 21B) indicate that water depths were as much as 50 m. This paleobathymetric relationship suggests that wave power reaching the shoreline was substantially higher when sequence 2 was deposited than today. This would explain the widespread continuity of sand-rich shorezone deposits in sequence 2 (figs. 21 and 22).

Sand bodies of sequence 3 have highly variable thicknesses and continuities (fig. 22). The delta front sands are thin (5 to 15 ft), lenticular, and highly discontinuous extending less than 1.6 km (1 mi) in a dip direction. Fluvial channel sands of the thick progradational wedge range in thickness from 6 to 20 m (20 to 65 ft) and they extend at least 3.2 km (2 mi) in a dip direction. The youngest fluvial channels are associated with crevasse splays that broke through natural levees of the main channels. Sands of the crevasse splays are also thin (3 to 5 m) and discontinuous (fig. 22).

RELATIONSHIP BETWEEN SEISMIC REFLECTIONS AND LITHOLOGIES

Most of the stratigraphic sequences at each depositional setting are composed of several lithologies. Composition of these lithologies was determined mainly by descriptions of borings rather than by seismic reflection patterns (seismic facies) because seismic reflections are a complex response of acoustical energy to various petrophysical properties, not just to lithology. Therefore, there is no direct correlation between reflection patterns and lithologies. Despite this limitation, some seismic facies and facies assemblages can convey generalized lithologic information because they are commonly associated with specific depositional environments, paleogeographic settings, or physical processes (Table 2).

Sequence Boundaries

Sequence boundaries mapped on seismic profiles may coincide with distinct lithologic changes or with abrupt changes in water content and cohesive shear strength, even if lithology is uniform. The decrease in water content and increase in strength are evidence of desiccation and oxidation (subaerial exposure) or overconsolidation (prior burial). Regardless of their origin, abrupt changes in physical properties may indicate the presence of disconformable surfaces (Fisk and McClelland, 1959).

In the High Island, Brazos South, and South Padre Island Areas, relatively deep and widespread sequence boundaries are represented by distinct changes in color of the muddy sediments. The muddy sediments are normally gray because they were deposited in a subaqueous reducing environment. Anomalous sediment colors such as red, orange, brown, tan, or yellow indicate an oxidizing environment and formation of a soil zone. Although soils clearly represent subaerial conditions, they are not diagnostic with respect to sea-level fluctuations because soils can form as a result of subaerial exposure or deposition above base level. Sediments that were deposited subaqueously can be exposed to the atmosphere and weathered when sea level is lowered. The soil horizon formed during the lowstand would be eroded or buried during the subsequent relative rise in sea level and associated transgression. Soil horizons can also form on coastal plain sediments that are deposited above sea level in non-marine environments such as floodbasins, natural levees, delta plains, and alluvial plains. These soil horizons can be preserved by subsidence and renewed sedimentation without a change in eustatic sea level or they can represent pauses during a rise in relative sea level.

The fluvial, deltaic, and strandplain sand bodies examined in this study are concentrated either within the midpart or near the top of seismic sequences. These stratigraphic positions of sandy facies relative to the sequence boundary appear to be different from those reported by Van Wagoner et al. (1990) and Vail and Wornardt (1991), who indicated that thickest sand bodies having the lowest mud content occur immediately above the sequence boundary. This discrepancy in positions of sand bodies relative to sequence boundaries is related to the problem of correlating the erosional unconformity with its correlative conformity in a basinward direction.

Difficulties in establishing sequence boundaries are related to geologic frames of reference and paleogeographic positions where correlations begin. If the sequence boundary correlation is traced downdip from updip control, then emphasis is placed on the erosional unconformity either across the exposed coastal plain and drainage divides or at the base of incised valleys. This erosion surface eventually merges basinward with the base of fluvial channels within the progradational wedge, which is a local surface of erosion. This miscorrelation between a sequence boundary and a facies change or local erosion surface is unavoidable when dealing with fluvial systems. On the other hand, if the sequence boundary correlation is traced updip from downdip control, then the sequence boundary and the downlap surface (maximum flooding surface) are essentially the same. This is because rapid transgression causes sediment starvation on the submerged shelf and precludes deposition of a thick transgressive systems tract over the lowstand systems tract; subsequent deposition of the highstand systems tract is confined to the shoreline, which is far landward of the starting point. In this study, sequence boundaries represent the strongest, and most widely traceable seismic reflections observed so even if these "sequence boundaries" are actually flooding surfaces and the unconformable sequence boundary is at the base of the fluvial channels, then the correlative conformities of that surface cannot be traced throughout the seismic grid with any confidence, thus rendering the unconformity bounded stratigraphic model impractical for these types of seismic profiles and boring data.

Van Wagoner et al. (1990) discussed how channel widths and lateral facies relationships can be used to distinguish between incised valleys and distributary channels so that sequence boundaries can be recognized and correctly located. Because the erosional channels imaged by the seismic profiles are less than 10 km wide and are in contact with underlying prodelta muds and overlying floodbasin muds, they are interpreted as fluvial channels associated with deposition of the progradational wedges and not as valleys incised into much older deposits as a result of lowered sea level. These criteria support placement of the sequence boundary beneath the progradational wedges rather than within the wedges at the base of thalweg scour.

Predominantly Muddy Sediments

Late Quaternary stratigraphic sequences beneath the Texas continental shelf are composed primarily of mud. This is because rivers draining the hinterlands traversed multicycle Tertiary fluvial-deltaic deposits that are also composed primarily of mud. Because the sequences are composed mostly of mud, correlations are highest between muddy lithologies and seismic reflection patterns.

The most common seismic reflection characteristics associated with these mud-rich sediments are parallel, high-amplitude, continuous and discontinuous reflections. The parallelism and gentle divergence of most seismic reflections suggest relatively slow deposition in shallow water characteristic of prodelta, shelf, and floodplain (delta and alluvial) environments (Table 2).

Steep, parallel or slightly divergent clinoform reflections are also commonly associated with muddy prodelta deposits of shelf and shelf-margin deltas. Short, low-angle, divergent reflections can also be associated with muddy levee and overbank deposits of delta systems.

Predominantly Sandy Sediments

Indirect geophysical detection of potentially sand-rich sediments is limited to a few seismic facies (Table 2) such as (1) contorted or hummocky reflections, (2) wavy clinoforms, (3) acoustically opaque patterns, (4) low-angle clinoforms, and (5) channel-fill patterns. Relatively thick sand bodies are commonly characterized by opaque seismic signatures that lack distinct internal bedding; also highly bioturbated mixtures of sand, shell, and mud may be indicated by transparent zones. Distributary channels are favorable sites for sand deposition. These sites can be recognized where distributary channels merge with irregular or wavy clinoforms. The irregular clinoforms are interpreted as migrating locations of distributary-mouth-bars where surficial creep and other mass movement processes transport bar sands into deeper water (Lindsay et al., 1984). The contorted or hummocky-wavy clinoforms typically indicate slumping of sand-rich delta-front deposits along an oversteepened unstable slope, whereas the acoustically opaque patterns commonly occur at the top of high-angle parallel clinoforms. Together the muddy clinoform sediments and overlying sandy transparent sediments constitute an upward-coarsening succession of progradational deltaic deposits.

Although clinoforms typically indicate muddy sediments, some sandy deposits are represented by short, low-angle and curved clinoforms (sequence 5, Brazos South Area, fig. 14). These compact clinoforms almost always delineate nearshore lateral accretion such as spit migration.

CONCLUSIONS

1. Late Quaternary stratigraphic sequences of the Texas continental shelf were deposited primarily by mud-dominated depositional systems.

2. Seismic boundaries separating these stratigraphic sequences coincide with soil horizons and downlap surfaces that are interpreted as being within the transgressive systems tracts of Posamentier et al. (1988).

3. Sequence boundaries of high-frequency glacio-eustatic sequences are the same as flooding surfaces that separate parasequences.

4. Correlative conformities cannot be identified on the basis of high-resolution seismic profiles and foundation borings.

5. In mud-dominated strata, correlations are highest between simple seismic reflections and muddy lithologies. Predicting locations of sand within high-resolution seismic sequences has highest probability near patterns that deviate from the norm (channel-fill, hummocky divergent strata, and tops of clinoforms).

6. Sand is deposited either near the middle or top of a complete stratigraphic sequence not immediately above sequence boundary as indicated by Posamentier et al. (1988).

7. Wave-dominated and river-dominated delta systems can be superimposed if sediment supply, subsidence, and sea level histories significantly reduce accommodation space near the shoreline of maximum transgression from one eustatic cycle to the next.

8. Sand bodies may not be deposited or preserved in interdeltaic deposits even after a fall or lowstand of sea level.

9. Uniform lithologies and lack of seismic reflection terminations make identifying sequence boundaries in interdeltaic deposits extremely difficult if not impossible.

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