SEQUENCE STRATIGRAPHY AND COMPOSITION OF LATE QUATERNARY SHELF-MARGIN DELTAS, HIGH ISLAND AREA, OFFSHORE TEXAS

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ABSTRACT

High-resolution seismic profiles and foundation borings from the northwestern Gulf of Mexico reveal the physical attributes of several late Quaternary depositional sequences that were deposited by wave-modified, river-dominated shelf-margin deltas during successive periods of lowered sea level. Each progressively younger sequence is thinner, and overall they exhibit a systematic decrease in the abundance and concentration of sand, which is attributed to a shift in the axes of trunk streams and greater structural influence through time.

Results of the study show that (1) contemporaneous structural deformation controlled the thickness of each sequence, the oblique directions of delta progradation at the shelf margin, and the axes of major fluvial channels; (2) a soil zone capping the oldest sequence is a regressive surface of subaerial exposure that was later preserved during marine transgression; (3) the downlap surfaces are not true surfaces but zones of parallel reflections that become progressively higher and younger in the direction of progradation; (4) the downlap zones are composed of marine muds that do not contain high concentrations of shell debris as would be expected in condensed sections; (5) evidence of submarine erosion and reworking of the delta surface during transgression (ravinement surface) is not widely observed probably because rapid subsidence coupled with rapid eustatic sea-level rise quickly submerged the delta plain below the depth of effective wave reworking; (6) no evidence exists that incised valleys or submarine canyons formed along the paleoshelf margin, even though moderately large rivers were present and sea-level curves indicate several periods of rapid sea level fall; and (7) boundaries of these high-frequency type 1 eustatic sequences are flooding surfaces that occupy the same stratigraphic position as boundaries separating parasequences.

INTRODUCTION

During years eight and nine of the Minerals Management Service Continental Margins Program, the Bureau of Economic Geology conducted detailed studies of sand-body continuity, primarily within shelf-margin lowstand systems tracts and transgressive systems tracts deposited by moderately large fluvial-deltaic systems. These same types of shelf-margin deposits are the preferred hydrocarbon exploration targets in the northern Gulf of Mexico and in other petroliferous basins worldwide. This report summarizes the results of the continental margins research, which has potential applications to geologic framework and petroleum-related studies in other parts of the world.

Shelf-margin deltaic sedimentation is the principal mechanism by which most continental margins prograde and sedimentary basins fill. Because shelf-margin deltas are an important element of basin-fill processes, several depositional models have been proposed that relate basin energy to delta morphology and sediment distribution. Genetic models of shelf-margin deltas are based on morphological characteristics and inferred physical processes (tides, waves, river discharge), whereas sequence stratigraphic models rely on subsurface data such as high-resolution seismic records (Lehner, 1969; Suter and Berryhill, 1985; Berryhill and others, 1987) or well logs and maps (Winker and Edwards, 1983).

Shelf-margin delta models also convey physical attributes regarding physiographic position, potential excavation of submarine canyons, and transport of sediment onto the adjacent continental slope. Some depositional models of shelf-margin deltas may be biased because much of the work has been from very large delta systems, such as the Mississippi or Niger, where sediment supply is extremely high and the deltas have reached the shelf-slope break during the most recent highstand in sea level.

The current study examines the lithologic variability and stratal characteristics of several late Quaternary shelf-margin delta systems in the High Island area of offshore Texas. These discrete, moderately large deltas were not fed by continental drainage systems or by rivers carrying glacial meltwaters during the warming phases of the glacio-eustatic cycles. They are thus similar to other ancient basin-filling deltas that have no large depocenter fixed in space for long periods of time.

A primary goal of this study was to test the predictive capabilities of depositional models derived from seismic sequences and to compare observed and predicted lithologies and stratal patterns within depositional sequences. This was achieved by integrating the lithologies of deep borings into the stratigraphic sequences mapped on high-resolution seismic profiles of the Texas continental shelf and upper slope (fig. 1).

Another objective of the study was to examine the mechanisms of sediment transport on the outer shelf and upper slope so that the extent of sand deposited by the lowstand shelf-margin deltas could be evaluated. Results of the study revealed the physical attributes (grain size, thickness, heterogeneity) and continuity of the shelf and slope sand bodies and their implications with respect to potential hydrocarbon reservoirs that formed as a result of similar sea-level fluctuations and shelf-margin positions. These same types of shelf-margin deposits are the preferred hydrocarbon exploration targets in the deep-water Gulf of Mexico Flex trend (Morton and others, 1991) and in other petroliferous basins around the world.

Several late Quaternary shelf-margin deltas of the Texas–Louisiana continental shelf have been mapped on the basis of high-resolution seismic profiles (Winker and Edwards, 1983; Lewis, 1984; Suter and Berryhill, 1985; Berryhill and others, 1987; Morton and Price, 1987), but the lithologic characteristics of the deltas have been inadequately described. Coleman and Roberts (1988) used foundation borings and seismic profiles to map late Quaternary sequences of the Louisiana shelf; however, that study did not specifically examine shelf-margin deltas or the factors controlling the distribution of sand bodies within deltaic sequences. Although the chronostratigraphic relationships among the shelf-margin deltas in the northwestern Gulf of Mexico are imprecisely known, it is clear that these young deltas constructed several relatively thick parasequences within each sequence as a result of rapid deposition in moderately shallow water.

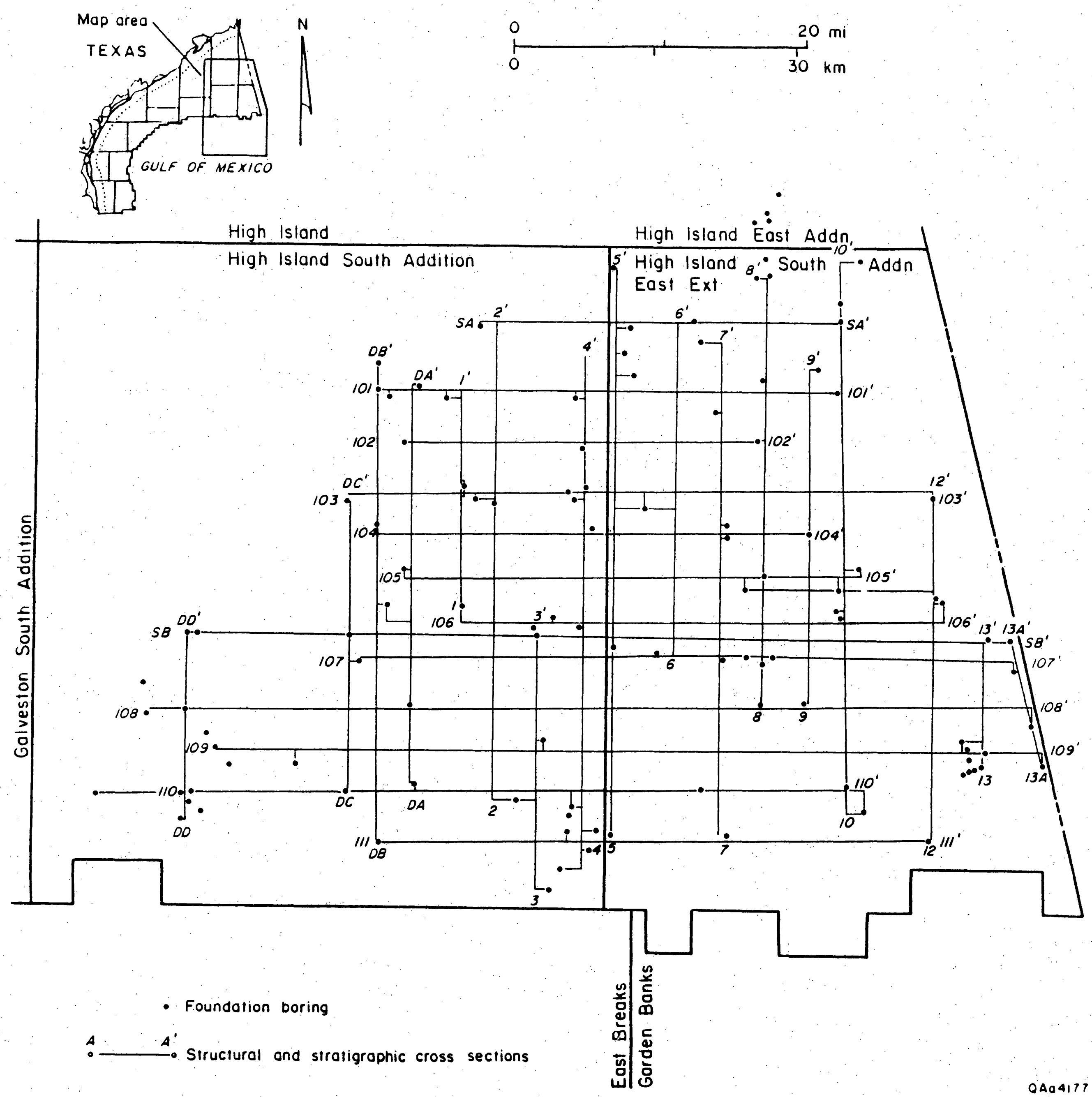

Figure 1. Index map showing study area and locations of foundation borings. Numbered cross sections coincide with grid of high-resolution sparker profiles that provide continuous coverage of the area.

The present study focuses on approximately 7,500 km^2 of the Texas outer continental shelf (fig. 1). The general morphology and seismic characteristics of the shelf-margin deltas have been previously reported (Lehner, 1969; Suter and Berryhill, 1985; Berryhill, 1987), but little information was available regarding composition of the deltaic sediments or the contemporaneous influence of sea-level fluctuations and diapiric structures on the distribution of sedimentary facies. The scarcity of evidence regarding the origins of these deltas led Winker (1991) to speculate that during lowstands in sea level, the shelf-margin deltas might have been wave dominated. An objective of this study was to evaluate the primary controls on shelf-margin deposition and sand distribution.

STRATIGRAPHIC FRAMEWORK

The stratigraphic framework of the late Quaternary shelf-margin deltaic sequences was established using foundation borings and a grid of single-channel high-resolution seismic profiles (fig. 1). These principal data bases were integrated to evaluate the relationships among seismic reflection characteristics, subsurface lithologies, and sequence boundaries. Seismic reflections were interpreted and mapped throughout the study area to provide detailed correlations and to determine which reflections could be traced subregionally. Depositional environments and depositional systems tracts of each sequence were interpreted on the basis of seismic characteristics, lithologic descriptions, stratigraphic position, and paleogeographic location. Numerous closely spaced foundation borings near the seismic profiles provided the control necessary to interpret the geologic history of the region. The seismic and lithologic control also made it possible to isolate the influence of global sea-level fluctuations on the construction and preservation of these high-frequency stratigraphic sequences.

Foundation Borings

More than 100 deep foundation borings (fig. 1) were used to determine the lithologic composition of each stratigraphic sequence and to establish the lithogenetic correlation framework. Most of the borings penetrate more than 90 m below the sea floor, and some are as deep as 150 m. Detailed boring descriptions provide information on subsurface depth, sediment color, sediment composition, presence and concentration of accessories (organic matter, shells, calcareous nodules), sediment textures, engineering properties, and other sedimentological attributes that can be used in lithostratigraphic correlations and interpreting depositional environments.

Seismic Profiles

More than 1,500 trackline miles of high-resolution sparker profiles were interpreted to establish the regional chronostratigraphic framework and to evaluate the relationship between seismic reflection patterns and lithologies. Late Quaternary shelf-margin deltas in the southern High Island area exhibit all common seismic reflection patterns including parallel, divergent, cut-and-fill, hummocky, and chaotic reflections; clinoforms; and reflection free patterns (table 1). The commonest patterns are clinoforms and parallel reflections having variable amplitudes and variable continuity. Inferred 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, and low-angle onlapping patterns commonly indicate retrogradation, whereas clinoforms always indicate either progradation or lateral accretion.

Depositional sequences were recognized and interpreted on the basis of internal seismic reflections, reflection terminations along sequence boundaries (Mitchum and others, 1977), lithologies, stratal geometries, and spatial variations in lithofacies and seismic facies. Sequence boundaries identified and illustrated by Suter and Berryhill (1985) and Berryhill and others

Table 1. 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

(1987) were verified on the seismic profiles and were used to establish the general stratigraphic framework. Sequence stratigraphic correlation among the foundation borings is possible with the aid of seismic profiles because the first seafloor multiple either occurs below the boring depths or does not obliterate reflections above the boring depths.

Seismic Velocities

The speed of sound in late Quaternary clastic sediments can be highly variable depending on sediment composition, water saturation, degree of compaction, and presence or absence of gas. Previous geophysical studies have estimated depths of late Quaternary strata using two-way travel times recorded on seismic profiles and average acoustical velocities that range from 1,525 m/sec (Sidner and others, 1978; Suter and others, 1987) to 1,700 m/sec (Bouma and others, 1983).

Two-way travel time was converted to depth using an acoustical velocity of 1,675 m/sec. This velocity was selected because the overconsolidated Pleistocene sediments should transmit sound faster than does sea water (1,525 m/sec). Also, 1,675 m/sec agrees closely with the velocity Lehner (1969) proposed for sediment types and depths similar to those of the present study. Abrupt 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, and parasequence boundaries) were transferred to structural cross sections so that lithologic variability within the seismic stratigraphic sequences could be analyzed and mapped. Soil zones were the primary physical evidence from the foundation borings that were used as lithostratigraphic correlation markers. The soil zones were identified by sediment color and the presence of carbonate or iron concretions. Stratigraphic correlations were made on the basis of either soil zones or seismic reflections because they coincided in numerous foundation borings. Another criterion used in correlating was systematic changes in sediment textures (upward-coarsening and upward-fining textural patterns) within the context of the overall sequence framework.

Coleman and Roberts (1988) reported that they were able to correlate individual carbonate zones (shell beds) over a broad region. The carbonate zones were interpreted as condensed sections representing sediment starvation of the continental shelf during regional transgressions. Our data indicate that the shell beds are evidence of reworking and shoreface retreat, but they are highly discontinuous and provide no basis for lithologic correlation independent of continuous seismic profiles. In fact, correlating most shell beds would lead to erroneous stratigraphic correlations because the several shell layers within each sequence do not carry a uniquely diagnostic signature that can be used in chronostratigraphic correlation.

Sand-Body Continuity

Each structural cross section was converted to a sequence stratigraphic cross section using the upper boundary of each sequence as a horizontal datum. These restored sections remove the distortions caused by postdepositional structural deformation and illustrate the stratigraphic framework, geometry (table 2), and lateral continuity of sand bodies within each sequence. Sand bodies were reconstructed three-dimensionally by integrating all the foundation borings that penetrated a sequence. Sand bodies within each sequence were correlated by projecting control from the cross-section network at the tie points between cross sections. Each sand body was assigned a relative chronostratigraphic age, which was determined by sand-body position with respect to other sand bodies and to the sequence boundaries. The relative

Table 2. Geometries 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.

chronostratigraphic ages were then used to map the positions of successive delta lobes and to interpret the environments of deposition.

LATE QUATERNARY SEA-LEVEL HISTORY

Sea-level curves of the northwestern Gulf of Mexico (Curray, 1960; Nelson and Bray, 1970; Frazier, 1974) generally have configurations that are comparable to those of other passive margin coasts. Most of the eustatic curves (fig. 2) show that sea level was at a lowstand about 160,000 ka (isotope stage 6) and then 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 midshelf during the middle Wisconsin interstadial about 50 ka (isotope stage 3). The subsequent 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. 2). Since then, sea level has remained essentially at its present highstand (isotope stage 1), although minor fluctuations of a few meters or less in amplitude may have occurred.

During the pre-Wisconsin, early Wisconsin, and late Wisconsin lowstands in sea level (isotope stages 6, 4, and 2), several shelf-margin deltas were constructed where river systems encountered the ancestral shelf-slope break (Suter and Berryhill, 1985; Suter and others, 1987). Instead of coalescing along the entire shelf margin, these deltas remained near their respective river mouths that had become fixed in space by the entrenched fluvial systems. Positions of the lowstand shelf-margin 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

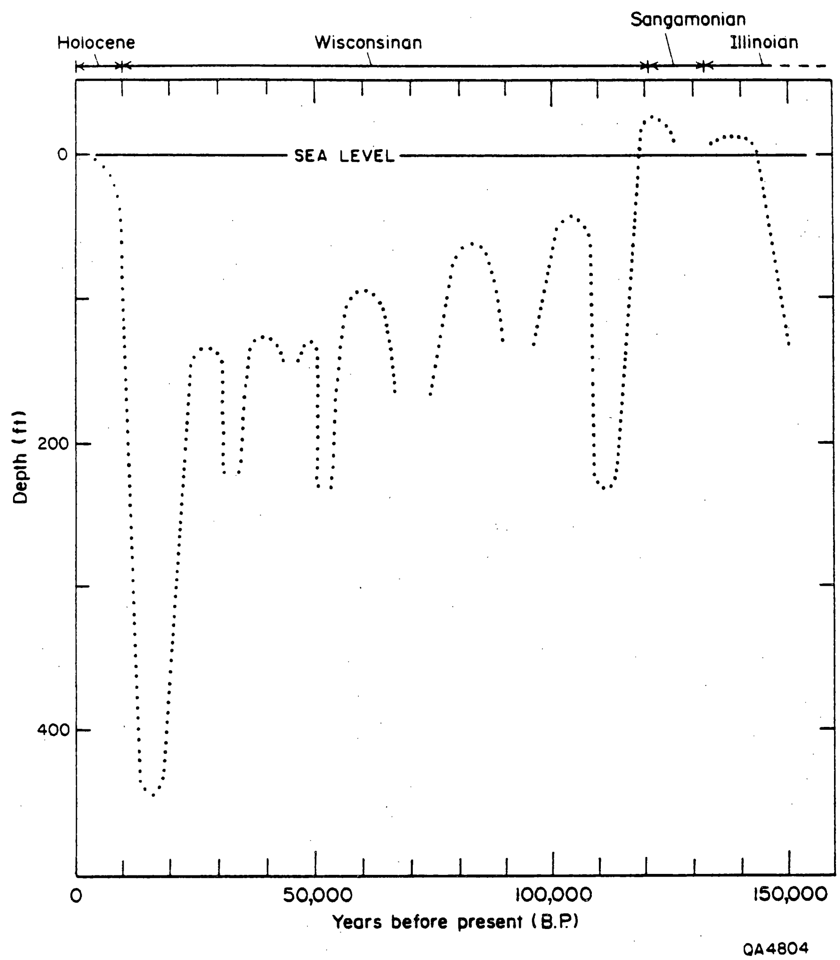

Figure 2. Late Quaternary sea-level fluctuations and corresponding O^{18} isotope stages. Modified from Moore (1982).

Texas continental margin, but it is actually lower than that of 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.

SYNDEPOSITIONAL STRUCTURES

Geological structures influencing late Quaternary sediments near the extant shelf margin (fig. 3) formed as a result of gravity-driven tectonism involving tensional stresses and sediment mobilization. These structures are near-surface expressions of deeper Plio-Pleistocene structures observed on multichannel common-depth-point (CDP) seismic profiles (Morton and others, 1991): faults, salt diapirs, withdrawal basins, and unconformities. Continuous movement of these structures or reactivation of older structures created the accommodation space for deposition of the youngest sequences.

Faults

Faults that displace the late Quaternary sediments coincide with families of faults that also disrupt the deeper Plio-Pleistocene strata (Morton and others, 1991). 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 expressed as the presence or absence of fault scarps at the seafloor. Recent movement of regional extension faults in areas of low sedimentation resulted in offset of the seafloor of as much as 15 m, whereas little or no offset has occurred where rates of sedimentation were high. The largest fault scarps are located in the southern third of the High Island area near the extant shelf margin (fig. 3), where salt mobilization and basinal extension are still active. Curray (1960) attributed the irregular bathymetry of the Texas outer continental shelf to erosional channels and depositional ridges. However, seismic profiles clearly show that some of the features are

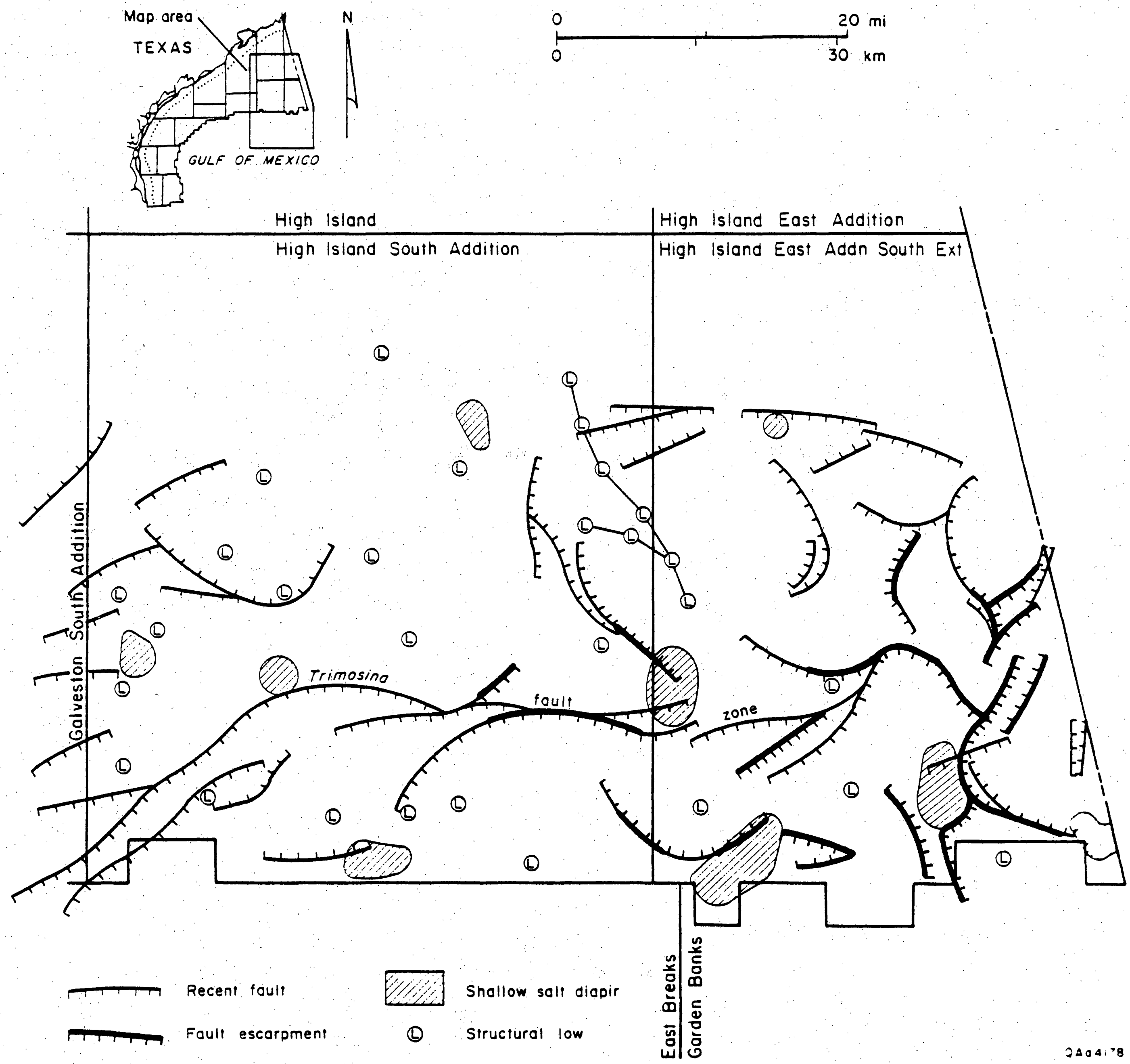

Figure 3. Generalized structural features showing principal faults, withdrawal basins, and salt diapirs.

escarpments formed by recent fault movement that caused rapid subsidence and created new accommodation space.

En echelon faults that form laterally continuous belts are referred to as 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 are subparallel to the paleoshelf margins of the underlying sequences and exhibit down-to-the-basin displacement. 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 are paired commonly with a primary regional fault that together form the basinward and landward boundaries of salt or shale withdrawal basins (grabens).

The *Trimosina* fault zone (fig. 3) has lowered the early Wisconsin sequence from 45 to 115 m and juxtaposed two stratigraphic sequences of different ages. Without the benefit of seismic control, sequence miscorrelation across the *Trimosina* fault zone is almost certain because the vertical displacement of the sequences is so large and the facies architecture of older and younger sequences is similar. Even high-resolution biostratigraphy might be unable to differentiate the juxtaposed sequences. This is because the sedimentary facies of both sequences are similar and the time elapsed between early Wisconsin and late Wisconsin depositional events was insufficient for significant faunal evolution.

Local faults having minor throw and limited lateral extent are either associated with salt diapirs, or they are antithetic to the major regional or counterregional faults. Faults associated with the 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 master faults of the regional or counterregional trends. The antithetic or stress-relief faults have displacements that are opposite to the major fault trends. These secondary faults can only be identified accurately from multichannel CDP seismic profiles because their intersection with the master faults occurs far below sparker-profile penetration.

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. Both scales of fault displacement are observed on the seismic profiles. Slumping as a result of oversteepening occurs at the shelf margin in the early Wisconsin and late Wisconsin sequences. Because the seismic profiles do not penetrate deep enough to image the former shelf margin of the pre-Wisconsin sediments, it is unknown how much sand in that sequence was transported downslope by slumping of the delta front.

Salt Diapirs

Two classes of salt diapirs affected the thickness and distribution of late Quaternary sequences in the High Island area. One is composed of those structures in which shallow salt is observed on the sparker profiles and adjacent strata dip steeply away from the diapir. The second class includes deep-seated diapirs that are manifested as broad, faulted structural highs and are identified by slight to moderate dips of beds on the flanks of the structures. In these deep-seated 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 and others, 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 lesser where the salt mass remains constant or increases as a result of evacuation from adjacent flanks of the diapir.

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

Withdrawal Basins

Withdrawal basins form where deep salt is evacuated and the overlying sediments subside. This subsidence creates new accommodation space that is filled if rates of sedimentation are high relative to the rates of subsidence.

Areas of greatest subsidence are outlined by isopach thicks of the early and late Wisconsin sequences (fig. 4). The *Trimosina* fault zone (fig. 3) overlies a salt wall of a salt withdrawal trough that controlled shelf-margin deposition of the early and late Wisconsin sequences. Thickness of the pre-Wisconsin sequence is similar across the principal faults, indicating that it was not greatly influenced by differential subsidence and increased thickness on the downthrown side, as is characteristic of most Gulf Coast extensional faults.

Unconformities

Regional, subregional, and local unconformities observed on seismic lines were recognized on the basis of stratal patterns that indicate erosion, such as onlap and truncation. A subregional unconformity mapped in the southern High Island area coincides with a soil horizon, which developed on top of deltaic deposits when they were subaerially exposed. Elsewhere, unconformities are either unrecognized (seismic reflections are parallel) or the missing section is limited in areal extent, such as around uplifts or at the bases of fluvial channels.

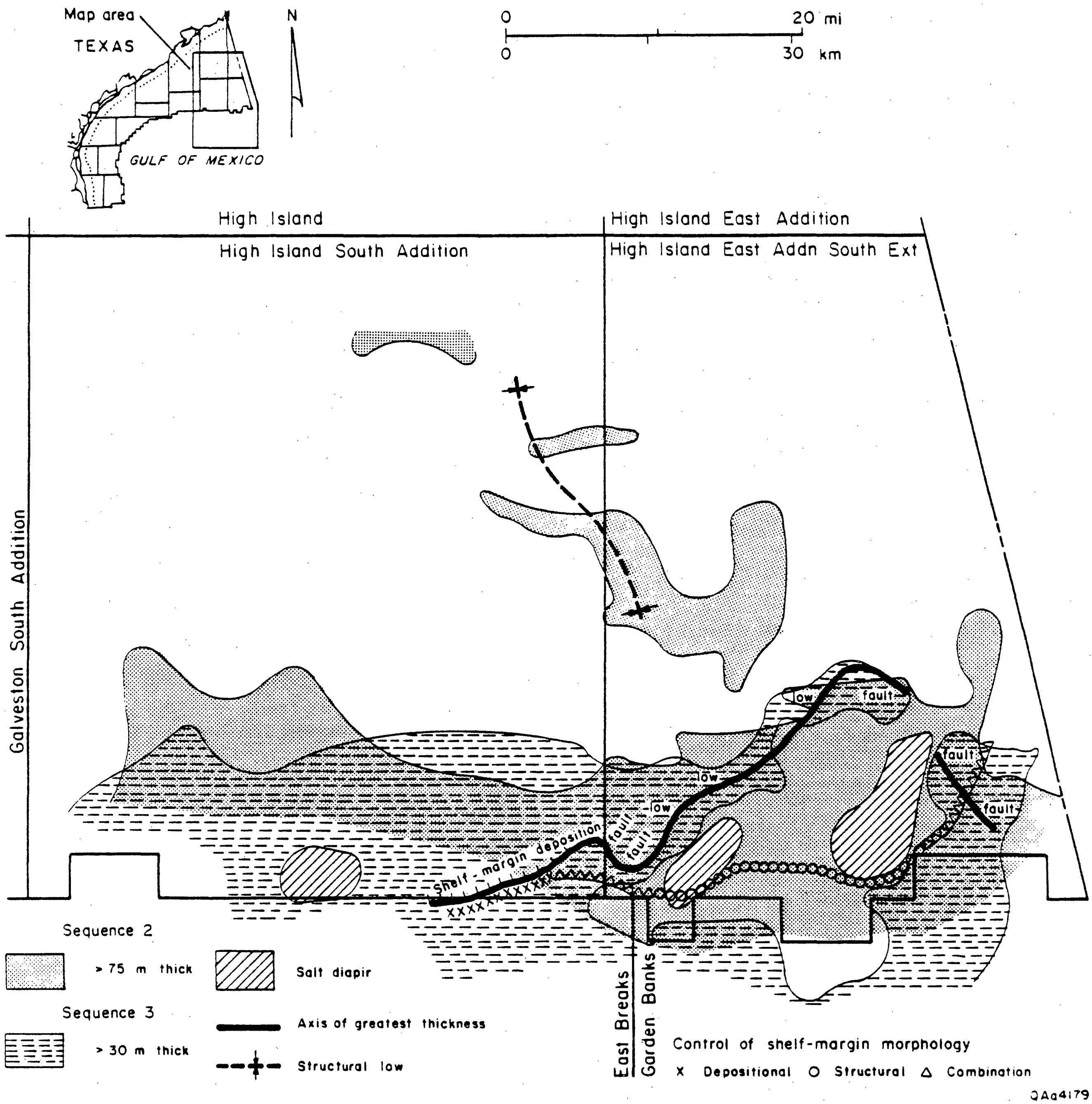

Figure 4. Areas of maximum subsidence as outlined by thickness of the early Wisconsin and late Wisconsin sequences.

Influence of Structures on Sequence Thickness

The active faults and salt diapirs near the shelf-slope break can greatly influence the thickness of each sequence. If the sequence is unaltered by active structures, then the depositional brow of the delta coincides with the thickest part of the sequence (fig. 5a). Shelf-margin deltas that are constructed on stable passive margins have this configuration. Although the depositional brow configuration is illustrated in all the shelf-margin delta models, it is atypical of the deltas in this study because structurally unaffected delta deposits constitute less than 20 percent of the entire width of the shelf margin investigated (fig. 4). Most of the shelf margin is partly controlled by either faults or structural highs. At most sites the shelf-slope break is unrelated to the thickest part of the sequence because the axis of deposition coincides with a withdrawal low, regional fault, or counterregional fault (fig. 5b, 5c, and 5d). In all three of these examples the physiographic shelf-slope break is basinward of the thickest part of the sequence (fig. 4).

SEQUENCE BOUNDARIES AND STRATIGRAPHIC SURFACES

Successful subsurface stratigraphic correlations require establishing sequence boundaries that can be either observed on the seismic profiles or recognized from the descriptions of the foundation borings. Ideal sequence boundaries have both diagnostic seismic images and distinguishing lithologic characteristics. Difficulties in establishing sequence boundaries relate to geologic frames of reference and paleogeographic positions where the stratigraphic correlations are initiated. If the sequence boundary correlation is traced downdip from updip control, then erosional unconformity is emphasized either across the exposed coastal plain and drainage divides or at the base of incised valleys (fig. 6). This nonmarine erosion surface eventually merges basinward, the base of fluvial channels within the progradational wedge, which is a local surface of erosion and not a regional unconformity. Miscorrelating a sequence boundary and a lithofacies change or local erosion surface is unavoidable when dealing with fluvial systems. On

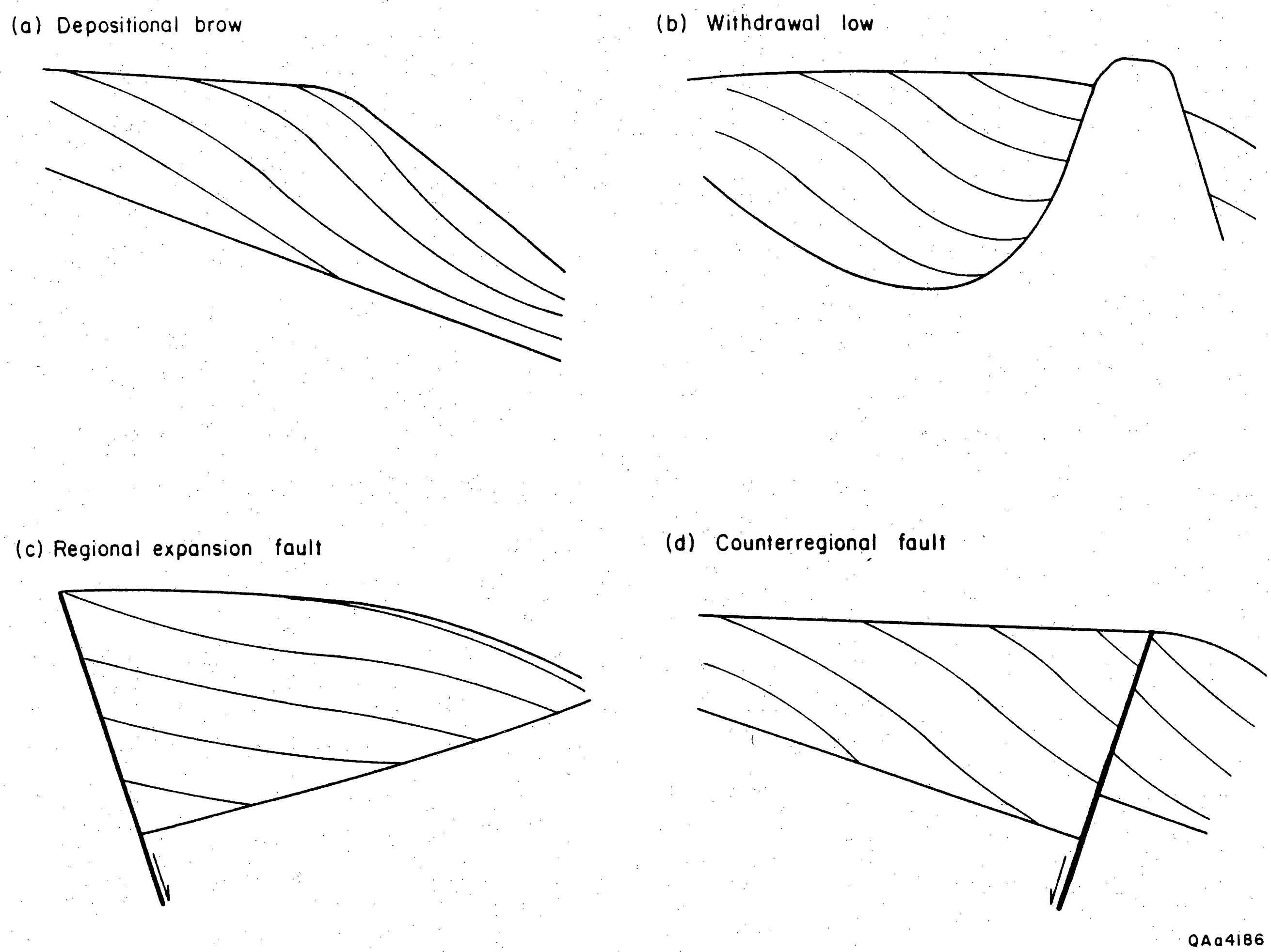

Figure 5. Schematic diagrams illustrating possible relationships between axes of greatest sequence thickness and physiographic shelf-slope break.

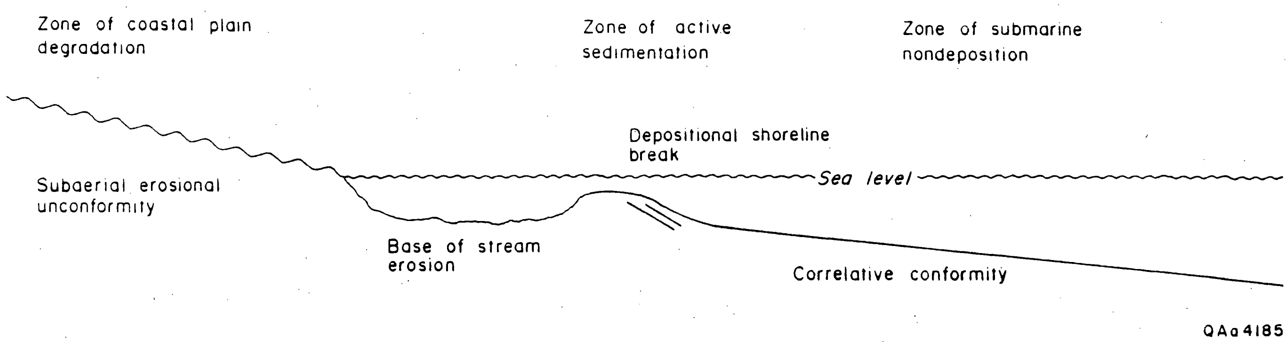

Figure 6. Relationship of sequence boundary to zones of erosion, deposition, and nondeposition.

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) can be essentially the same surface. This is because rapid transgression commonly 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 correlation starting point.

Seismic Sequence Boundaries

The most prominent and continuous seismic reflections in the southern High Island area separate strata of different ages, and they are used as sequence boundaries. These boundaries also approximate downlap surfaces (figs. 7 through 14), which operationally coincide with maximum flooding surfaces and which are used to separate transgressive and highstand systems tracts (Posamentier and others, 1988). As previously noted, 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. Even if these "sequence boundaries" are actually flooding surfaces and the unconformable sequence boundary is at the base of the overlying 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 in these types of seismic profiles and foundation borings.

The same sequences and surfaces mapped on the high-resolution seismic profiles can also be seen on multichannel CDP profiles, although at a much smaller scale. Despite lack of detail in the seismic reflections, the sequence boundaries mapped on the CDP lines (fig. 15) are the same downlap surfaces as those identified on the high-resolution profiles.

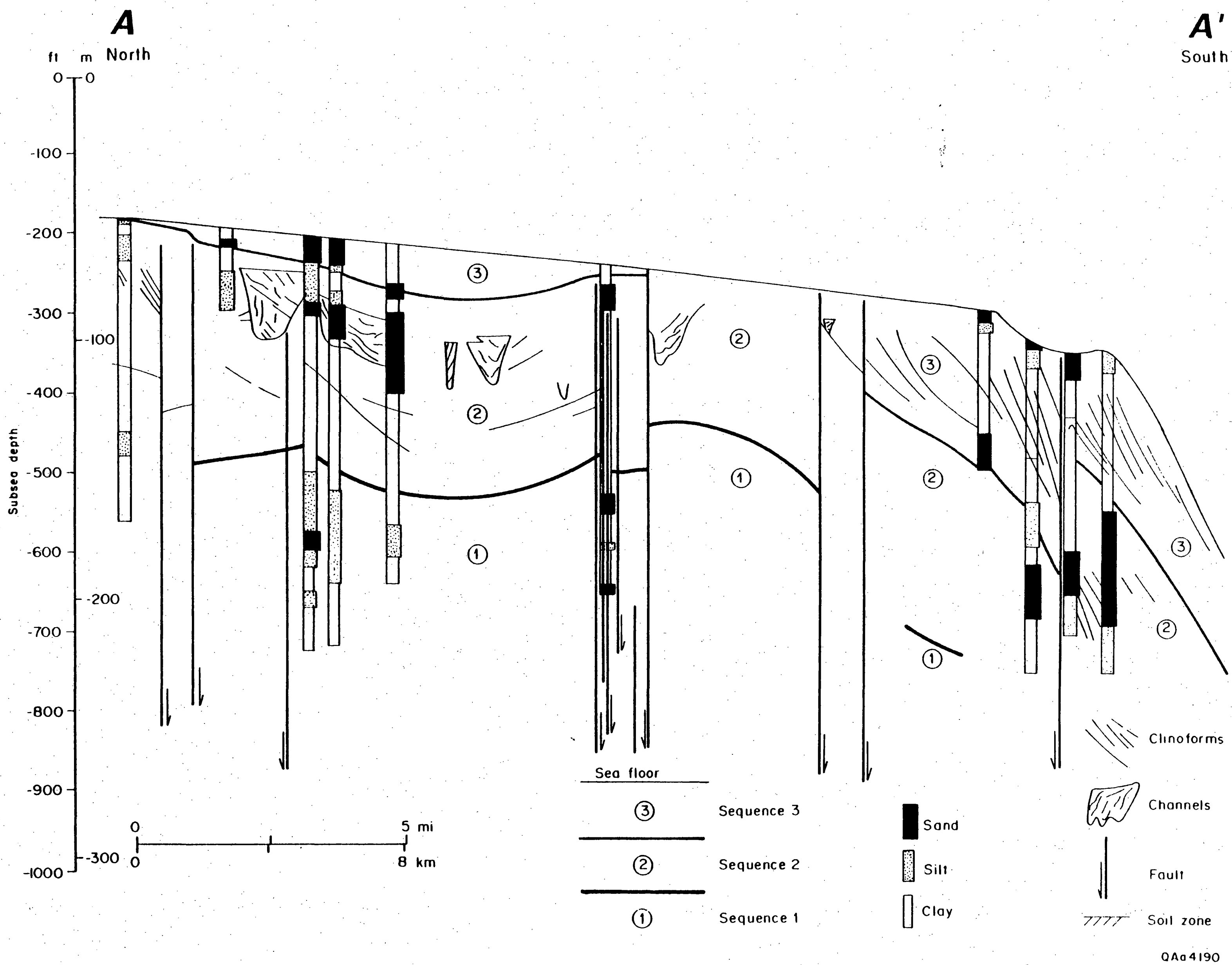

Figure 7. North-south structural dip section A–A′ illustrating thickness, composition, and deformation of the late Quaternary sequences.

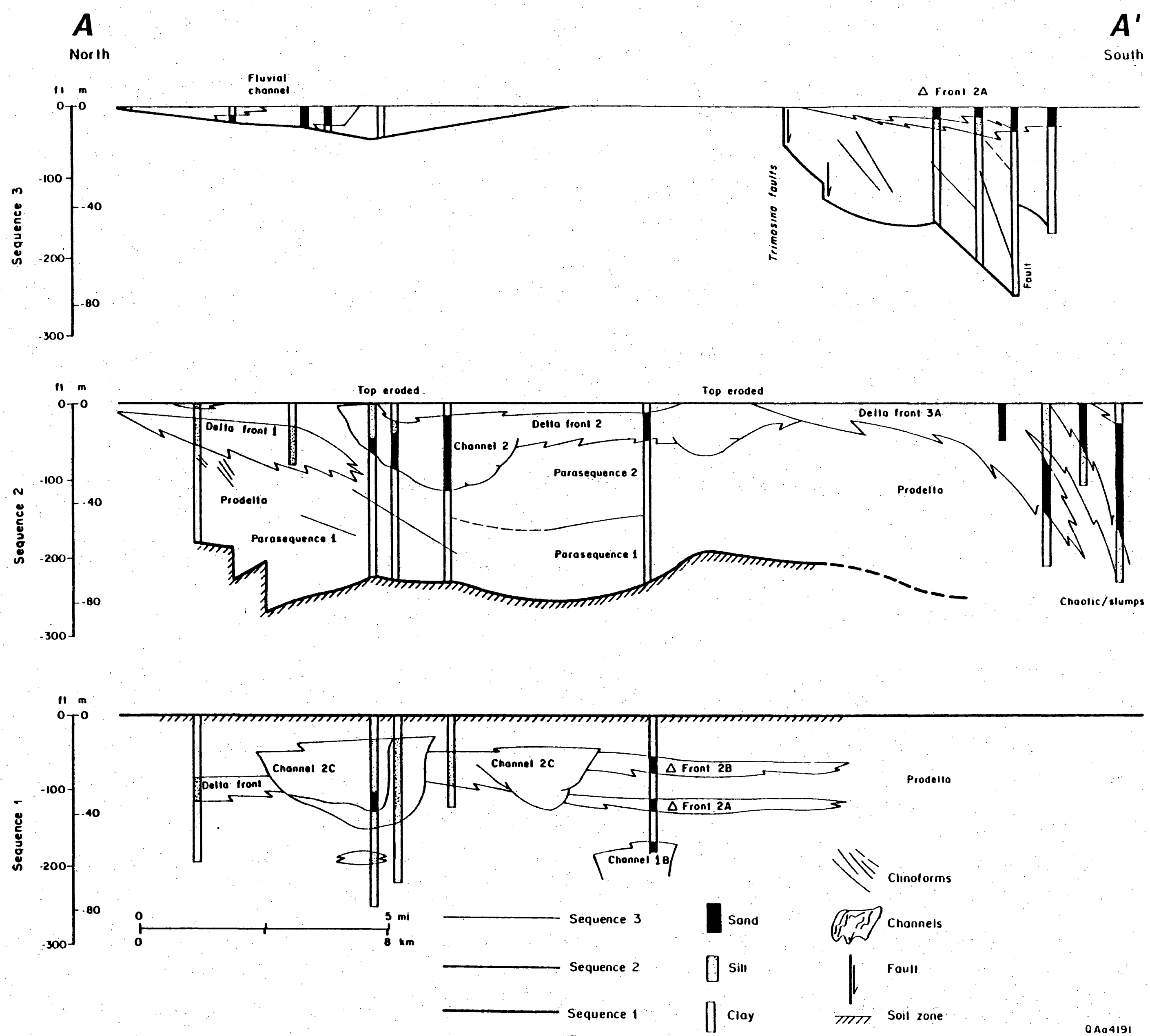

Figure 8. Restored north-south dip section A–A′ illustrating sand-body continuity for each of the three sequences.

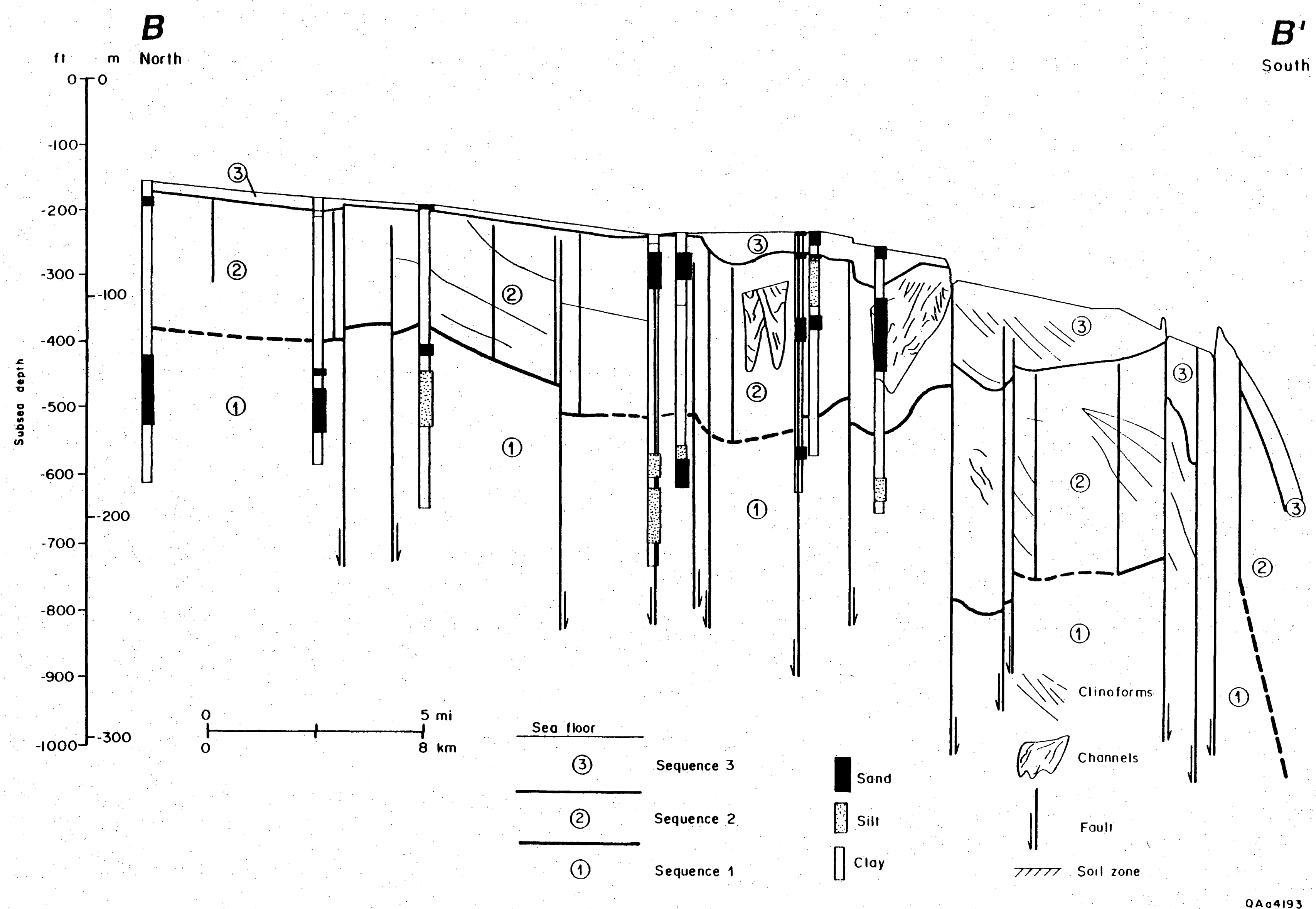

Figure 9. North-south structural dip section B–B′ illustrating thickness, composition, and deformation of the late Quaternary sequences.

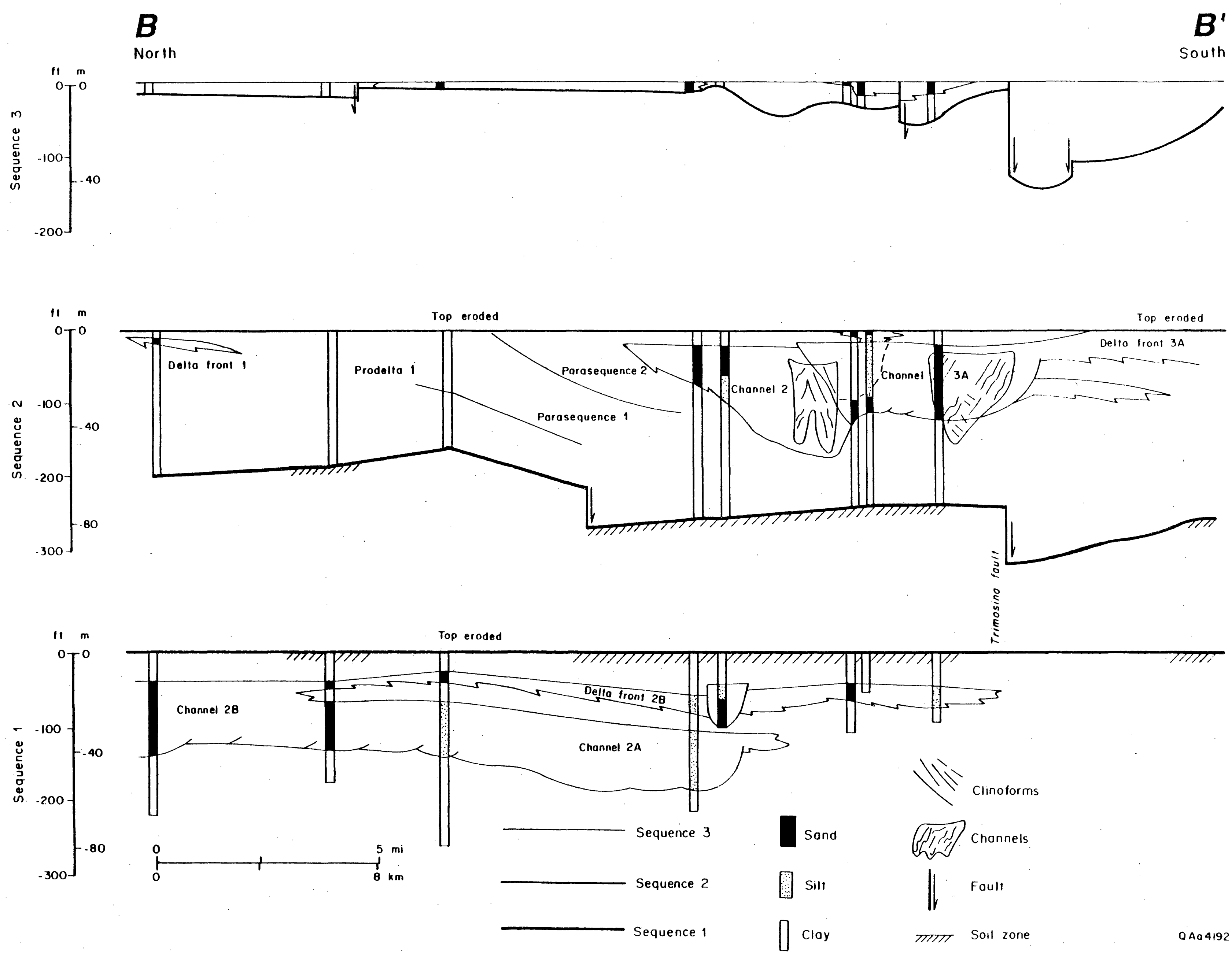

Figure 10. Restored north-south dip section B–B′ illustrating sand-body continuity for each of the three sequences.

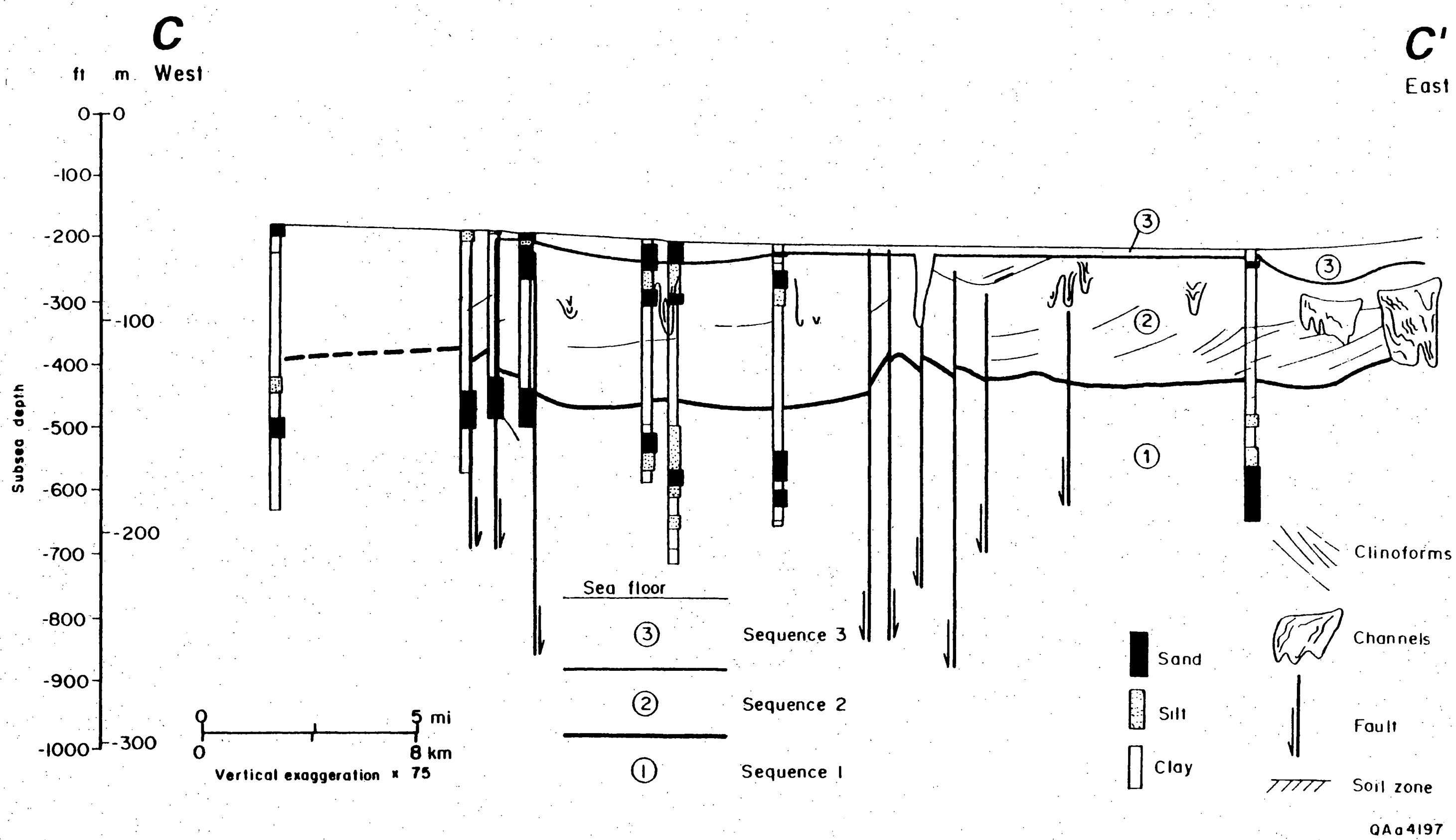

Figure 11. West-east structural strike section C–C′ illustrating thickness, composition, and deformation of the late Quaternary sequences.

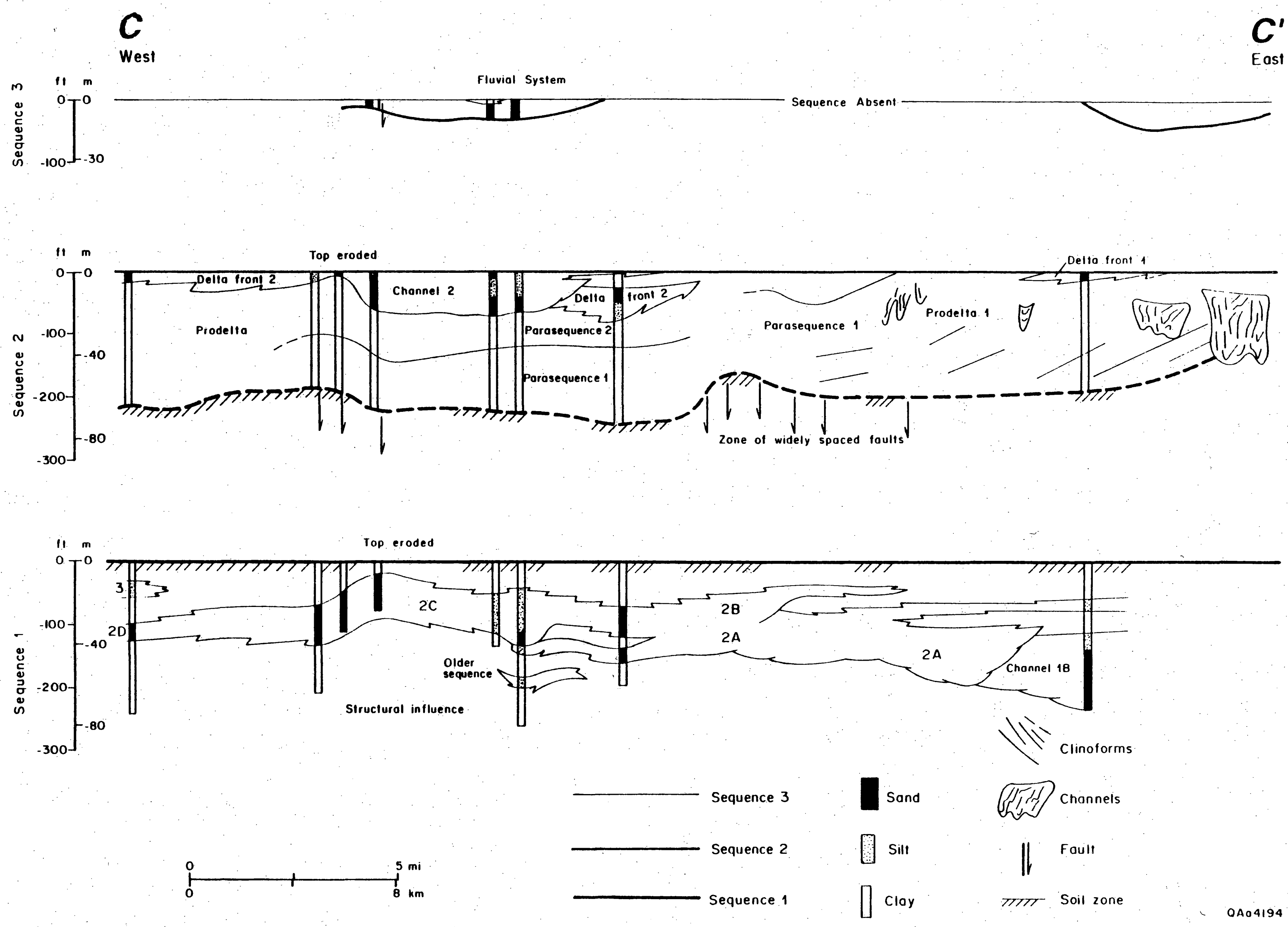

Figure 12. Restored west-east strike section C–C′ illustrating sand-body continuity for each of the three sequences.

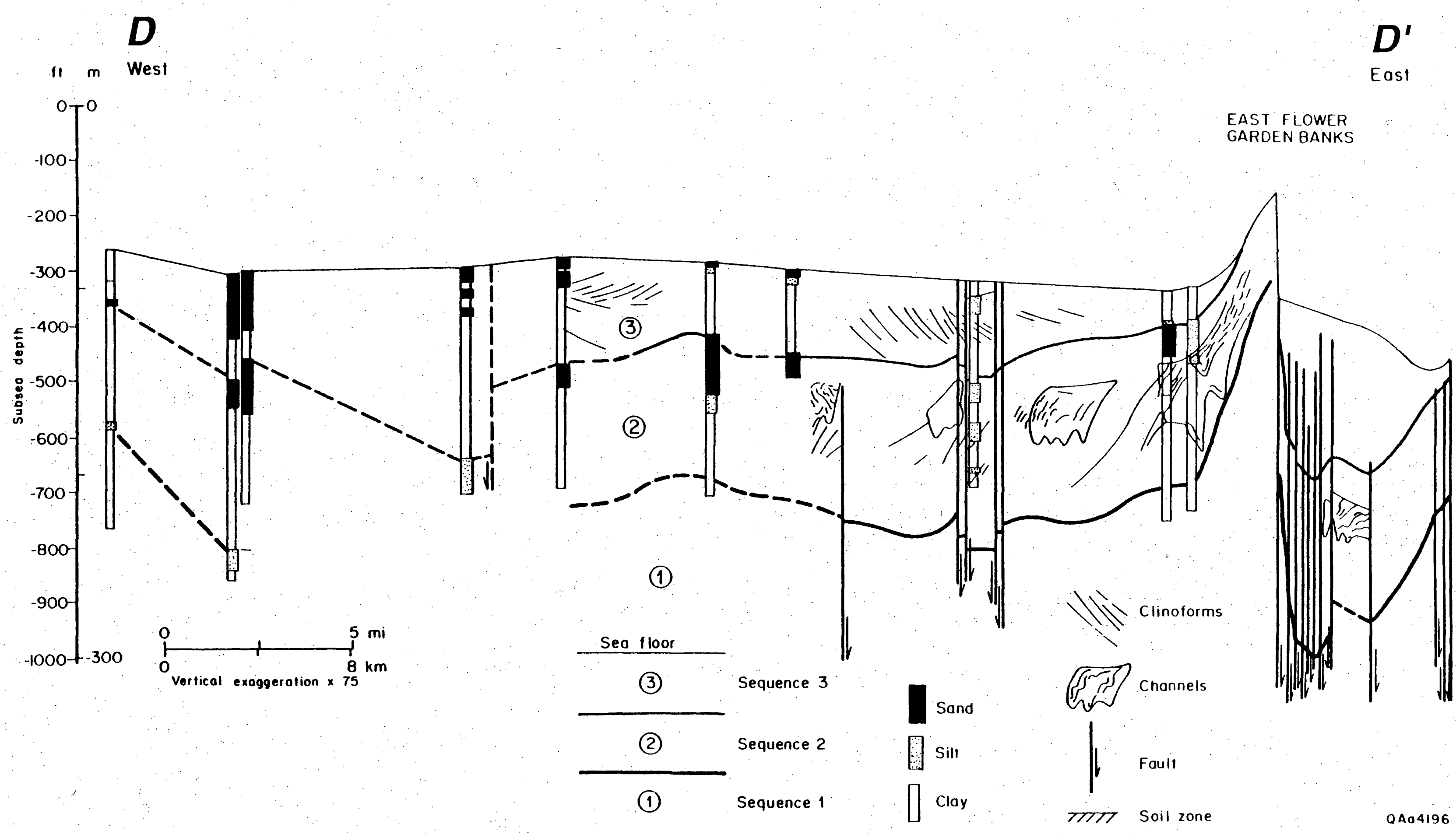

Figure 13. West-east structural strike section D–D′ illustrating thickness, composition, and deformation of the late Quaternary sequences.

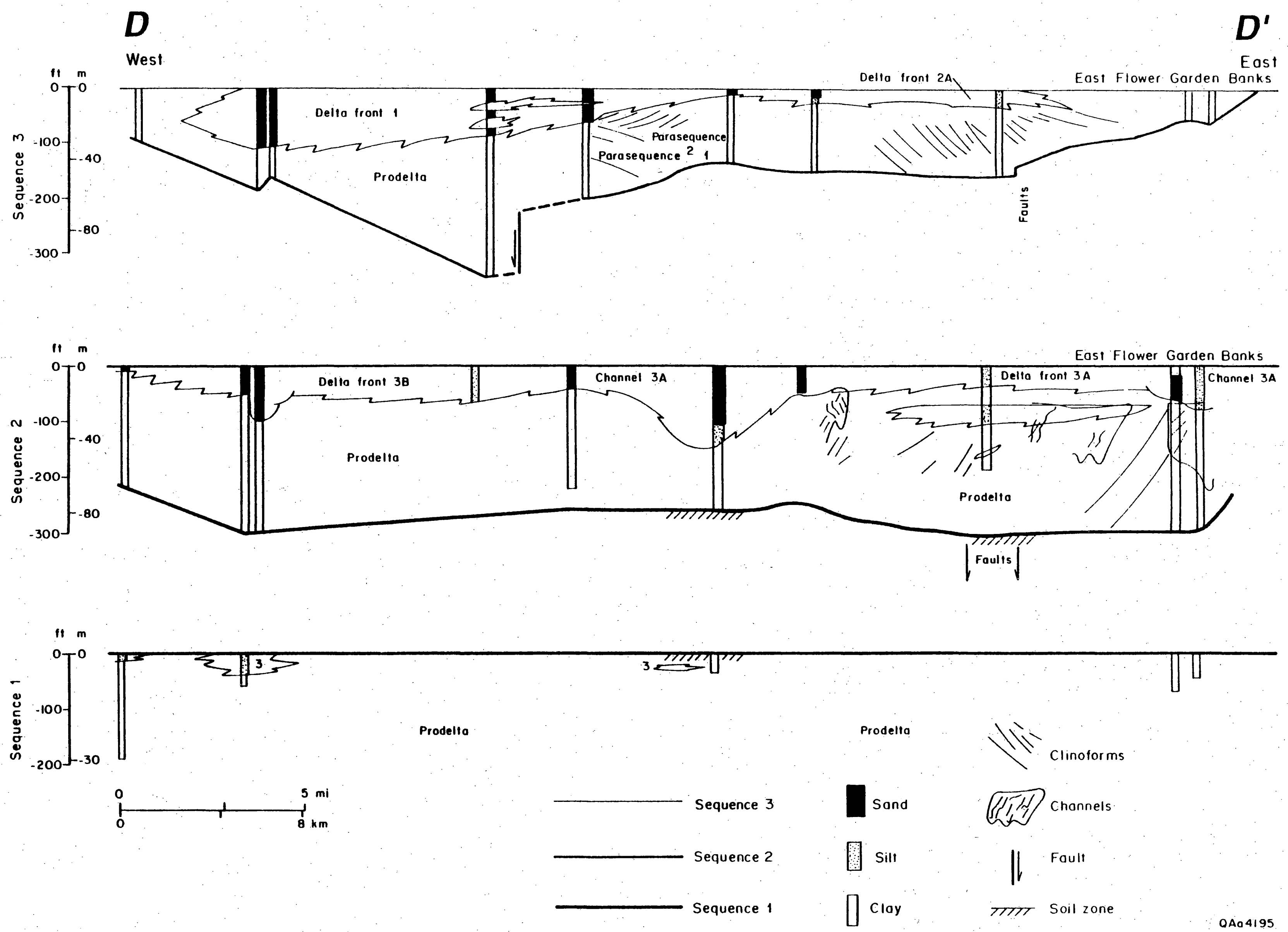

Figure 14. Restored west-east strike section D–D′ illustrating sand-body continuity for each of the three sequences.

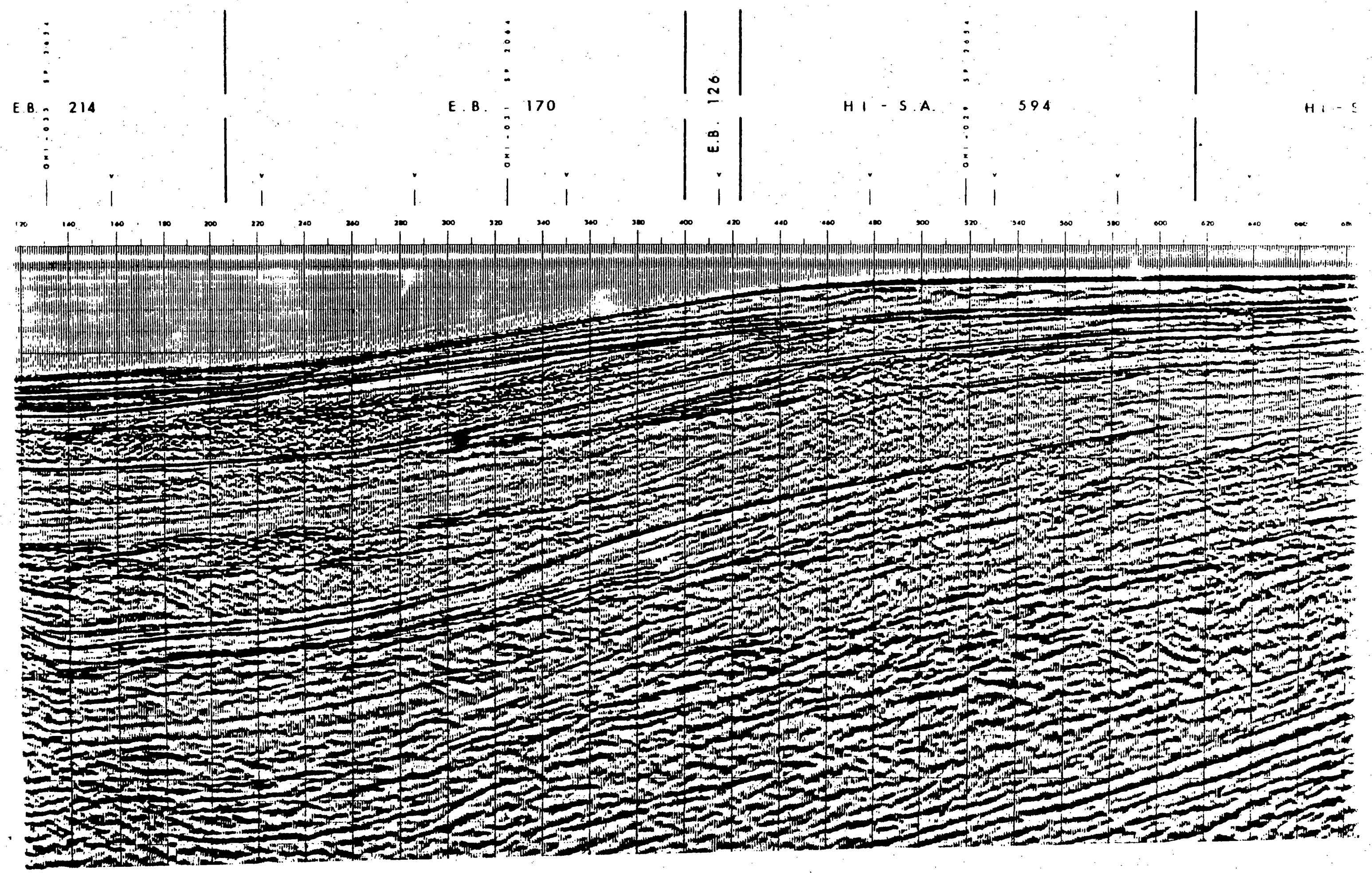

Figure 15. Multichannel seismic profile illustrating stratal characteristics and stacking patterns of late Quaternary shelf-margin deltas.

Submarine Erosion or Hiatal Surfaces

Some high-amplitude, nearly planar surfaces having local erosional features appear to be sequence-boundary candidates on several seismic profiles, but they do not correspond to equivalent strong, continuous reflections on adjacent profiles. These surfaces can occur within the regressive deposits or within the overlying transgressive deposits. In regressive deposits they typically occur near the base of the sequence, and they eventually merge with clinoform reflections. Such areally restricted local reflections cannot be correlated throughout the seismic grid, and they occur progressively higher in the section toward progradation. These surfaces separate delta lobes or progradational parasequences and are interpreted as submarine erosion surfaces produced by a temporary pause in deposition (hiatus) and minor shoreface retreat. Similar onlap and truncation surfaces within sets of clinoforms are interpreted as minor abandonment surfaces associated with allocyclic shifts in depocenters. If these submarine erosion surfaces were misinterpreted as subaerial erosion surfaces of regional significance (sequence boundaries), then the overlying bundles of parallel continuous seismic reflections would be interpreted as thin lowstand and transgressive systems tract deposits, and the thick, overlying progradational wedges would be interpreted as highstand systems tract deposits.

Lithologic Sequence Boundaries

Sequence boundaries mapped on seismic profiles can also be recognized in the descriptions of the foundation borings. The seismic sequence boundaries may coincide with distinct lithologic changes or with abrupt changes in color, water content, and cohesive shear strength, even if the lithology remains unchanged. 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 southern High Island area, a relatively deep and widespread sequence boundary is represented by distinct changes in color of the muddy sediments, which are normally gray after having been 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 necessarily diagnostic of sea-level fluctuations because soils can form as a result of sediment deposition above base level or subaerial exposure of formerly subaqueous deposits. Sediments that were deposited subaqueously can be exposed to the atmosphere and weathered when sea level lowers. The soil horizon that formed during the lowstand would be eroded or buried during the subsequent relative rise in sea level and associated transgression. But soil horizons can also form on coastal-plain sediments that are deposited above sea level in nonmarine environments such as floodbasins, natural levees, delta plains, and alluvial plains during any phase of sea-level change. These soils associated with slow sedimentation and prolonged subaerial exposure 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.

DEPOSITIONAL ENVIRONMENTS

Depositional environments of the shelf-margin deltas were identified on the basis of seismic reflection patterns, sediment textures, accessory materials (organic matter, shell), vertical successions of lithofacies, and paleogeographic position. Maps depicting the paleogeographic distribution of depositional environments for each sequence represent the average position of the environments when the deltas reached their maximum regressive position. The depositional framework within each sequence was based primarily on the three-dimensional distribution of lithofacies and sand-body geometries. The three types of sand bodies recognized in the shelf-margin-delta deposits are associated with fluvial channels (meandering and distributary channels), delta-front environments, and strandplain environments.

Prodelta Deposits

The prodelta deposits are composed primarily of clay and silty clay along with thin layers of sand. Discontinuous patches of marine shells are also present where finer sediments have been removed by winnowing or where low rates of sedimentation caused shell debris to concentrate locally. The prodelta deposits constitute the basal lithofacies in each of the sequences. They typically grade upward into coarser grained delta-front deposits or make erosional contact with overlying fluvial-channel deposits.

Prodelta deposits appear on the seismic profiles as clinoform reflections that typically change shape from low-angle oblique to high-angle sigmoid forms basinward (figs. 8 and 10). This change in depositional style illustrates the effect of progradation into progressively deeper water and the influence of a rise in relative sea level that eventually overcomes sediment influx (Mitchum and others, 1977).

Delta-Front Deposits

The delta-front environment encompasses those sediments that were deposited near the river mouth in distributary-mouth bars and that were later reworked by waves and currents to form broad sand sheets. They are thus influenced by both fluvial and marine processes. The delta-front deposits generally coincide with the tops of the clinoform reflections (figs. 8 through 10).

The delta-front sediments are composed of sandy silt, silty sand, and sand, which constitute an upward-coarsening vertical succession of lithofacies. The presence of abundant silt and gradational contact with underlying prodelta muds help distinguish these sandy sediments from fluvial-channel fills, which exhibit abrupt basal contacts with underlying sediments.

Fluvial Channels

On the seismic profiles, fluvial channels are recognized as cut-and-fill features that have irregular erosional bases and that encompass zones of lateral accretion surfaces (Suter, 1987). They are composed of sand and silty sand and commonly exhibit upward-fining textural patterns. In some borings the presence of coarse sand and gravel is reported near the base of the channels, but these coarser sediments are rare.

Van Wagoner and others (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. In this study, they used channel dimensions, channel shape, and internal reflection characteristics to distinguish between principal meandering trunk streams and distributary channels. Wide, deep, and nested channels exhibiting evidence of lateral migration and repeated occupation (figs. 7 through 14, sequence 2) are interpreted as major alluvial systems that have eroded into the underlying (slightly older) deltaic deposits. In contrast, single, narrow channels that exhibit no evidence of lateral migration are interpreted as distributary channels.

Most of 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. No evidence exists of subaerial exposure at the base of the channels, as might be expected if they were former incised valleys. The bases of the channels truncate clinoform reflections, but they eventually merge with the clinoforms in the direction of progradation. These spatial relationships indicate contemporaneous deposition of distributary-channel and delta-front sediments. The thickness of the channel fill nearly equals the depth of water in which the delta was deposited, but thickness of the valley fill should be substantially less because accommodation space is lost during a fall in sea level. On the basis of these observations, the channels 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 lower sequence boundary beneath the progradational wedges rather than within the wedges at the base of thalweg scour.

Delta-Plain Deposits

The delta plain is a transitional to nonmarine environment that encompasses both subaqueous and subaerial mud flats located between the fluvial channels. Sediments are primarily supplied to the delta plain by overbank flooding along the channels, which causes the mud flats to aggrade as the delta progrades. The delta-plain deposits correspond to high-amplitude parallel seismic reflections that occur at the same elevations and above the fluvial channels. They are composed of clay and silty clay and commonly contain organic material, such as wood fragments, but they do not contain shells except where the delta-plain environment was an interdistributary bay. The fine-grained delta-plain deposits are indistinguishable from other mud-rich sediments except for their stratigraphic position with respect to the fluvial channels. One diagnostic criterion is the soil profile, which will develop on the delta-plain surface if it is subaerially exposed for prolonged periods.

Strandplain Deposits

Relatively thin lenses of sand and silty sand containing some shell beds occur near the top of each sequence. They are 2 to 6 m thick and discontinuous in both strike and dip directions. The sand lenses are interpreted as small barrier islands, spits, and nearshore shoals that formed during the transgressive phase of deposition as the delta plain was inundated by marine waters. These strandplain deposits also coincide with high-amplitude parallel seismic reflections. Because the seismic responses of the strandplain deposits are not diagnostic, they are best recognized in the foundation borings on the basis of their stratigraphic position, patchy distribution, and presence of shell.

SEQUENCE DISTRIBUTION AND COMPOSITION

Sequence 1 (Pre-Wisconsin)

Sequence Thickness and Composition

The oldest stratigraphic sequence is only partly delineated because on most of the seismic profiles the basal sequence boundary is obscured by reflection multiples or degraded quality of seismic data. The pre-Wisconsin sequence is best represented in the north part of the southern High Island area, where foundation borings penetrate most or all of the sequence (figs. 7 through 14). In this updip position, sequence 1 is about 60 m thick.

Sequence 1 typically consists of four lithologic units that exhibit both upward-coarsening and upward-fining vertical facies successions. The basal unit is composed of stiff gray clay and silty clay containing rare layers of shell. Above the clay and silty clay is olive gray to gray silty fine sand. The sand, which ranges in thickness from 4.5 to 45 m (fig. 16), is overlain by sandy silt or silty clay. 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 having thin interlayers of sand and silt and containing some calcareous nodules and shell fragments.

A soil zone forms the upper boundary of sequence 1 (figs. 7 through 14). Deep weathering profiles indicate prolonged periods of subaerial exposure, when sedimentation rates on floodplains are relatively low. The soil zone at the top of sequence 1 is composed of red, brown, or yellow sediments, and here and there these colors mixed with gray. The soil zone ranges in thickness from 1 to 8 m. At the basinward limit of its penetration, the soil zone is less than 1 m thick because the duration of soil development was shorter on the younger sediments and because possibly greater removal by erosion occurred during the subsequent transgression. At several locations two soil horizons are about 12 m apart. The repeated development of soil profiles suggests aggradational processes and frequent subaerial exposure.

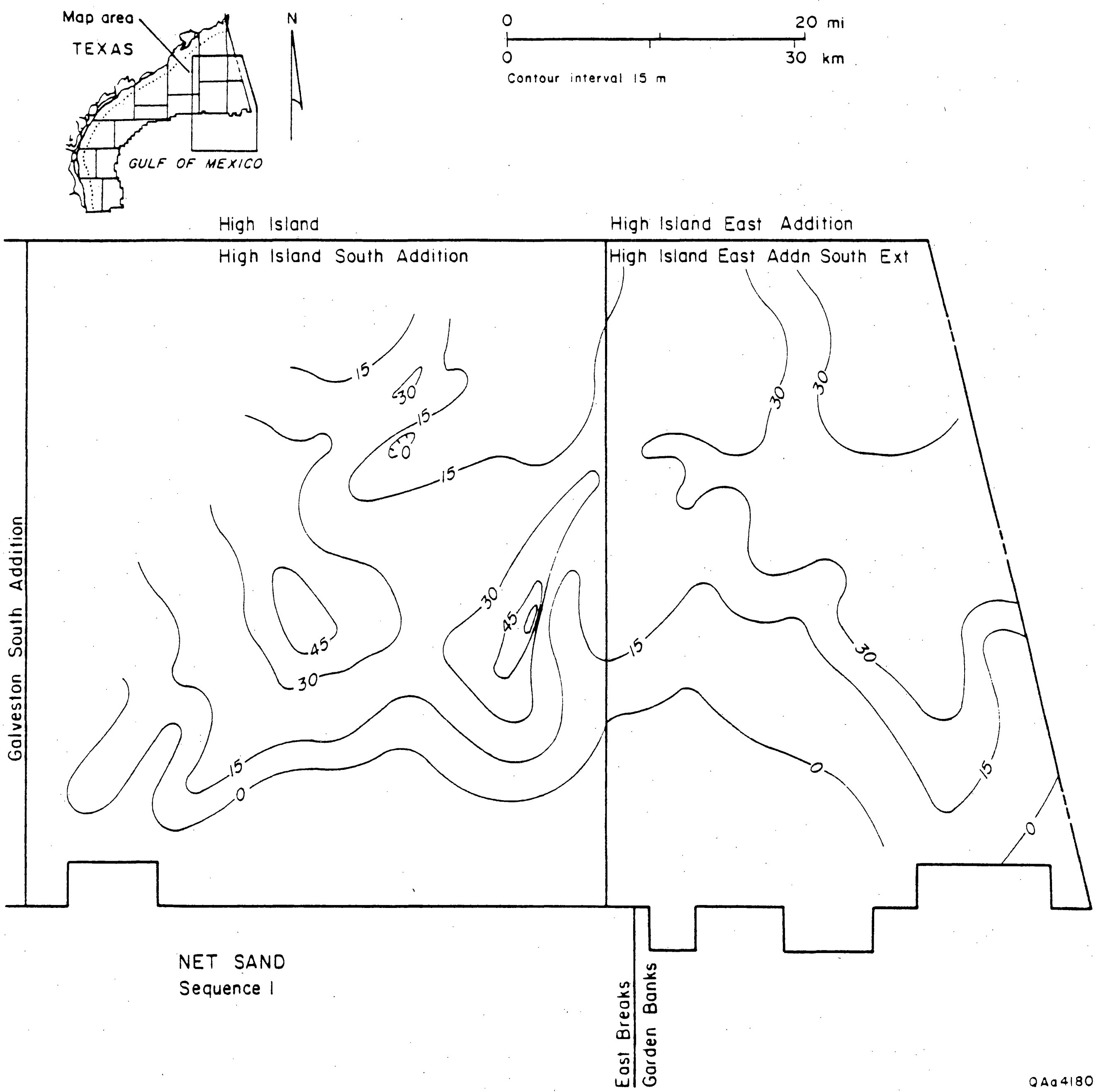

Figure 16. Net-sand isolith of pre-Wisconsin sequence (sequence 1).

Sand Distribution

The sand-body framework of sequence 1 consists of multiple stacked and offset fluvial-channel and delta-front sands (figs. 7 through 14). Most of the sand bodies occur near the middle of the sequence but some occur near the top, depending on position relative to the systems tract. The sand bodies are thick and laterally continuous in updip and middip positions, but they become thinner and more interbedded downdip. The thickest accumulations of sand (>30 m) have elongate geometries that have northeast-southwest orientations and that coincide with the principal fluvial channels (figs. 16 and 17). Mud-rich interfluves separate the channel sand bodies in updip positions, but the sand bodies merge and overlap along the delta front.

Sequence 1 was deposited by at least three delta complexes that prograded to the southwest near the paleoshelf margin (figs. 15 and 17). The directions of progradation are inferred from patterns of deposition (lithofacies) because the seismic reflections are not diagnostic. The oldest delta complex (1), long and narrow, was located mainly in the center of the study area. It is identified by the presence of deep channel and delta-front sands that account for the greatest sand thickness in the sequence (figs. 8 and 12). Because delta complex 1 cannot be traced landward, it may represent the upper part of an older sequence.

The delta complex of intermediate age was encountered in all the borings penetrating sequence 2, and it accounts for the extensive lateral continuity and vertical stacking of most of the sand bodies of sequence 2. This delta complex was supplied by four fluvial systems (fig. 17). Channel systems 2A and 2B in the east are vertically stacked, whereas to the west, channels 2C and 2D are at about the same stratigraphic level, indicating nearly contemporaneous deposition.

The youngest delta complex, located in the south central area (fig. 17), is areally restricted. It occupies the same general position as delta complex 1 except that it prograded farther to the southwest. At the top of sequence 1, thin patchy beds of sand and shell are remnants of

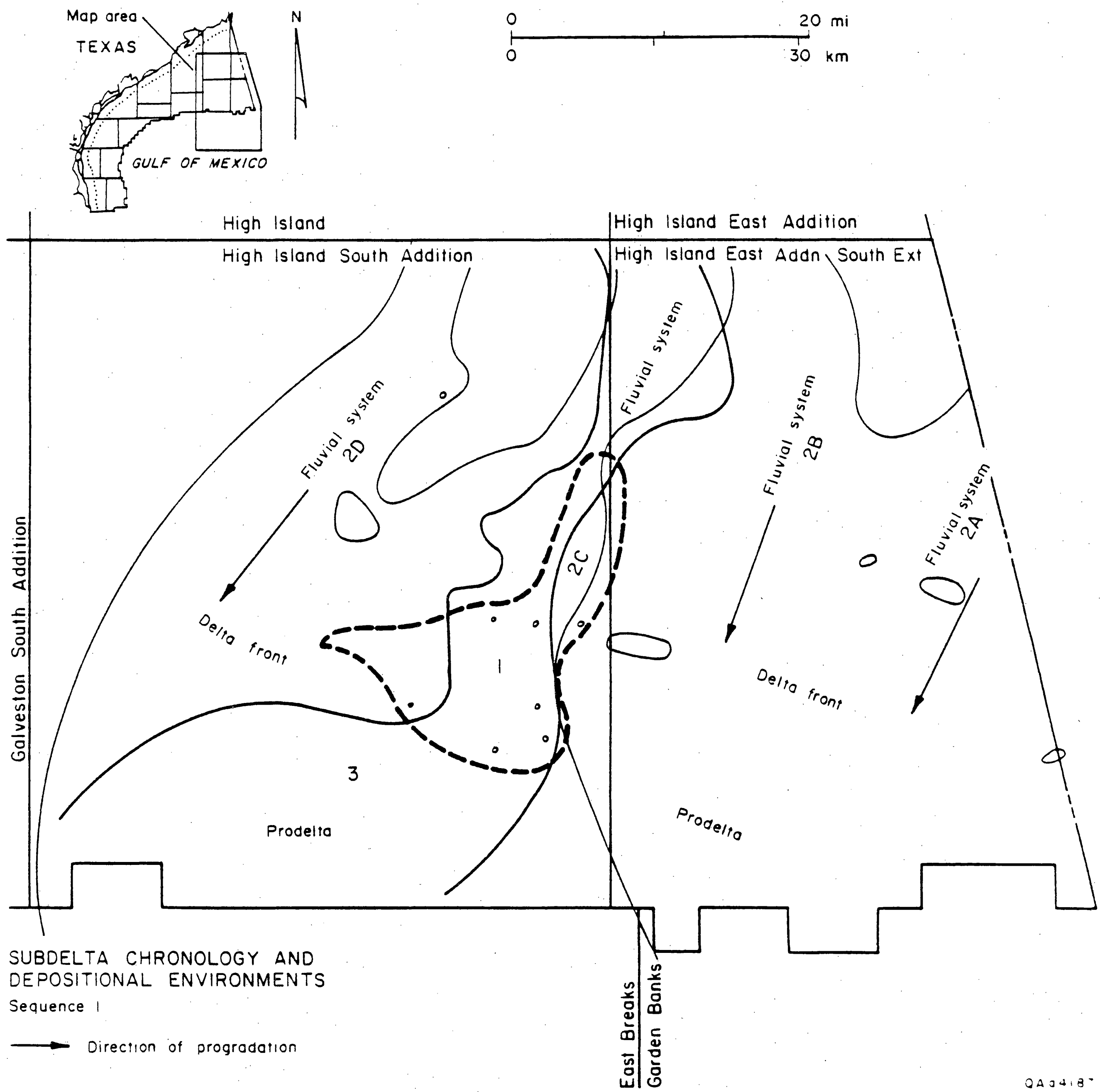

Figure 17. Depositional systems and subdelta chronology of pre-Wisconsin sequence.

shorezone deposits that locally accumulated as the delta became inundated and transgressed (fig. 14 and 17).

The basinward limit of sand in sequence 1 is poorly constrained because few borings penetrate deep enough to encounter sequence 1 near the extant shelf margin. Nevertheless, sand abundance appears to decrease systematically near the *Trimosina* fault zone. Basinward of the *Trimosina* fault zone, the sequence is composed mostly of prodelta mud except to the southeast, where the downdip limit of sand was not penetrated. There sandy sediments seem to extend downslope into the East Breaks and Garden Banks area (fig. 16). The gradual basinward decrease in sand abundance suggests that downslope mass transport of sand was an insignificant process during sequence 1 deposition. However, the zone of low sand abundance on the upper slope of sequence 1 could have been an area of sand bypass where turbidity currents transferred sand from the delta front to the continental slope. If present, the sedimentary evidence of downslope sand transport is basinward of where the sequence is penetrated by available borings.

Depositional History

The vertical facies successions and lithologies of sequence 1 indicate that it was deposited during a prolonged regression that was associated with a falling phase and lowstand of sea level, probably isotopic stage 6 (fig. 2). The delta systems of sequence 1 advanced the shelf margin to the southwest, but the sequence was not greatly affected by contemporaneous deformation. The delta complexes exhibit river-dominated morphologies except near their terminus, where waves and longshore currents modified the distributary-mouth bars and formed broader lenses of sand along the delta front (fig. 16).

The regressive phase of deposition culminated in coastal-plain aggradation above sea level and formation of an extensive soil profile. This subaerial marker is widely preserved across the muddy delta plain, even though it was submerged and partly truncated by marine erosion during the subsequent isotopic stage 5 transgression. The paleosoil is absent where the top of the sequence is composed of sand, and at these sites it probably eroded during the shoreface retreat phase of the regional transgression.

The parallel seismic reflections of sequence 1 observed beneath the present outer shelf indicate aggradation, whereas progradational clinoforms are observed farther basinward beneath the upper slope (fig. 15). The slight basinward shift in depositional patterns of sequence 1 relative to the overlying sequences probably results from a greater fall in sea level during the stage 6 lowstand.

Chronostratigraphic Correlation

A substantial correlation discrepancy exists between "pre-Wisconsin" sediments identified on the high-resolution sparker profiles (Suter and Berryhill, 1985) and those identified using deeper subsurface data (Morton and others, 1991). Suter and Berryhill (1985) 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 countdown method of chronostratigraphic interpretation suggests that the Sangamon highstand deposits, relatively shallow beneath the continental shelf, should be penetrated by the deepest foundation borings.

Petroleum industry paleontologists identified the Sangamon interglacial deposits on the basis of faunal assemblages and the extinction of the foraminifera *Globorotalia flexuosa* (Wornardt and Vail, 1991). Maps and cross sections of this extinction horizon, correlated in well logs and on CDP seismic profiles (fig. 15), indicate that sediments above the *Globorotalia flexuosa* extinction horizon are more than 600 m thick near the extant shelf margin (Morton and Jirik, 1989; Morton and others, 1991) and should not be encountered in even the deepest foundation borings.

There are at least two explanations for the "pre-Wisconsin" correlation discrepancy. First, if the industry "Sangamon" interglacial transgression and highstand mapped in the deeper subsurface is the 130,000-yr-B.P. event, then more Wisconsin sequences have been preserved than have been previously recognized. On the other hand, if the industry "Sangamon" interglacial event is actually older than 130,000 yr B.P., then the three sequences mapped by Suter and Berryhill (1985) could represent all the sequences of Wisconsin age. Without absolute ages or indirect evidence such as oxygen isotope data, this discrepancy is resolved only with difficulty; it cannot be resolved using only foundation borings and sparker profiles. Some correlation charts report that the *Globorotalia flexuosa* extinction horizon is older than 130,000 yr B.P.; consequently, the chronostratigraphic relationships established in the southern High Island area by Suter and Berryhill (1985) and Berryhill and others (1987) were maintained for the purposes of this study.

Sequence 2 (Early Wisconsin)

Sequence Thickness and Composition

Sequence 2, which is entirely penetrated by most of the borings, provides the best stratigraphic control on sequence thickness and sand distribution (figs. 7 through 14). It is at least 60 m thick except where influenced by diapirs (fig. 18). Moderate structural influence is indicated by the relationship between sequence thickness and the principal structural features (compare figs. 4 and 19). The thickest part of the sequence either coincides with a northwest-southeast trending structural sag or is downthrown on the *Trimosina* fault zone (fig. 4). The sequence is more than 90 m thick and abruptly thickens at the paleoshelf margin (fig. 18).

Sequence 2 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 and is composed of sandy silt, sandy clay, clay and sand, and clay and silt.

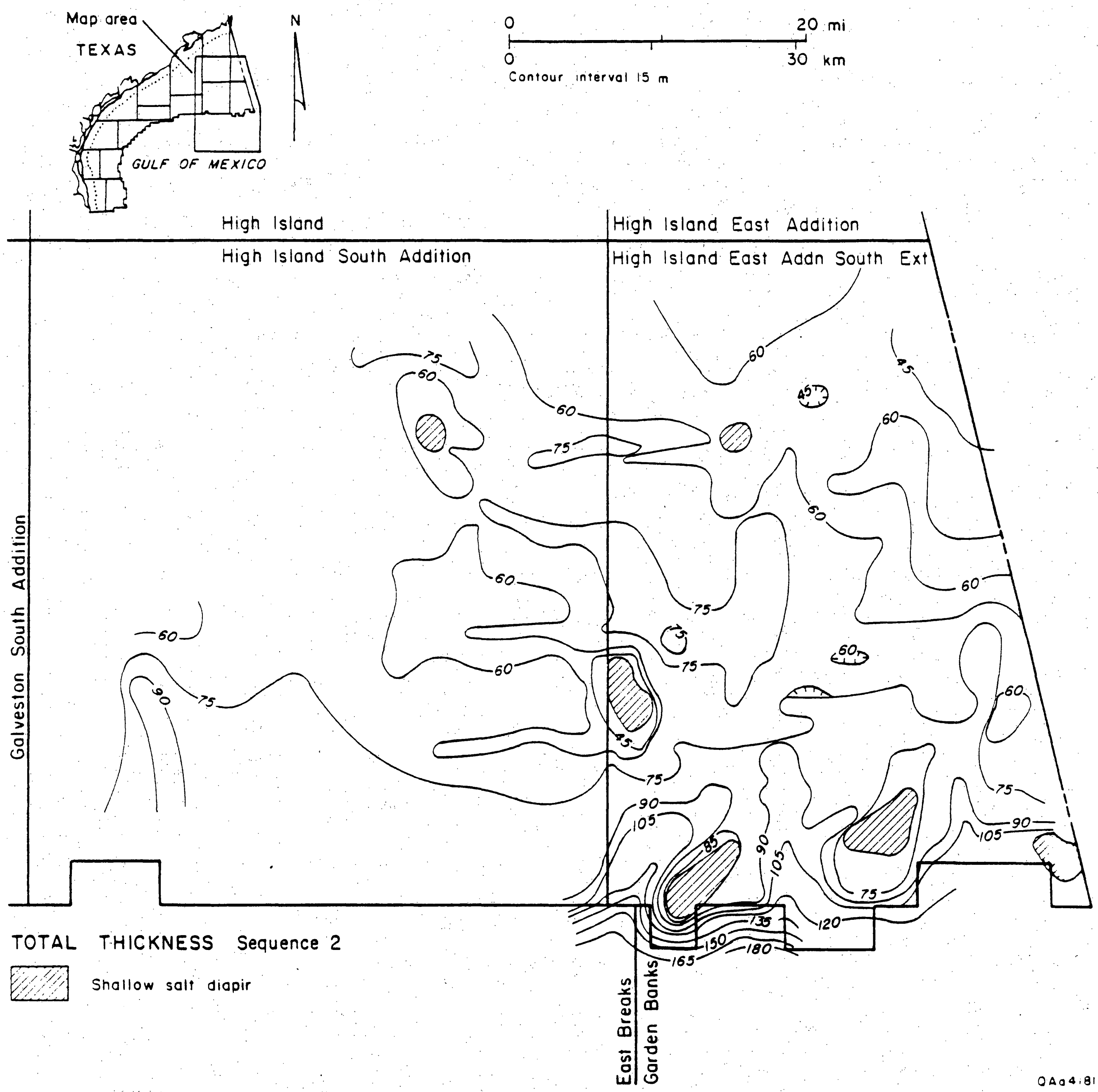

Figure 18. Thickness of early Wisconsin sequence (sequence 2).

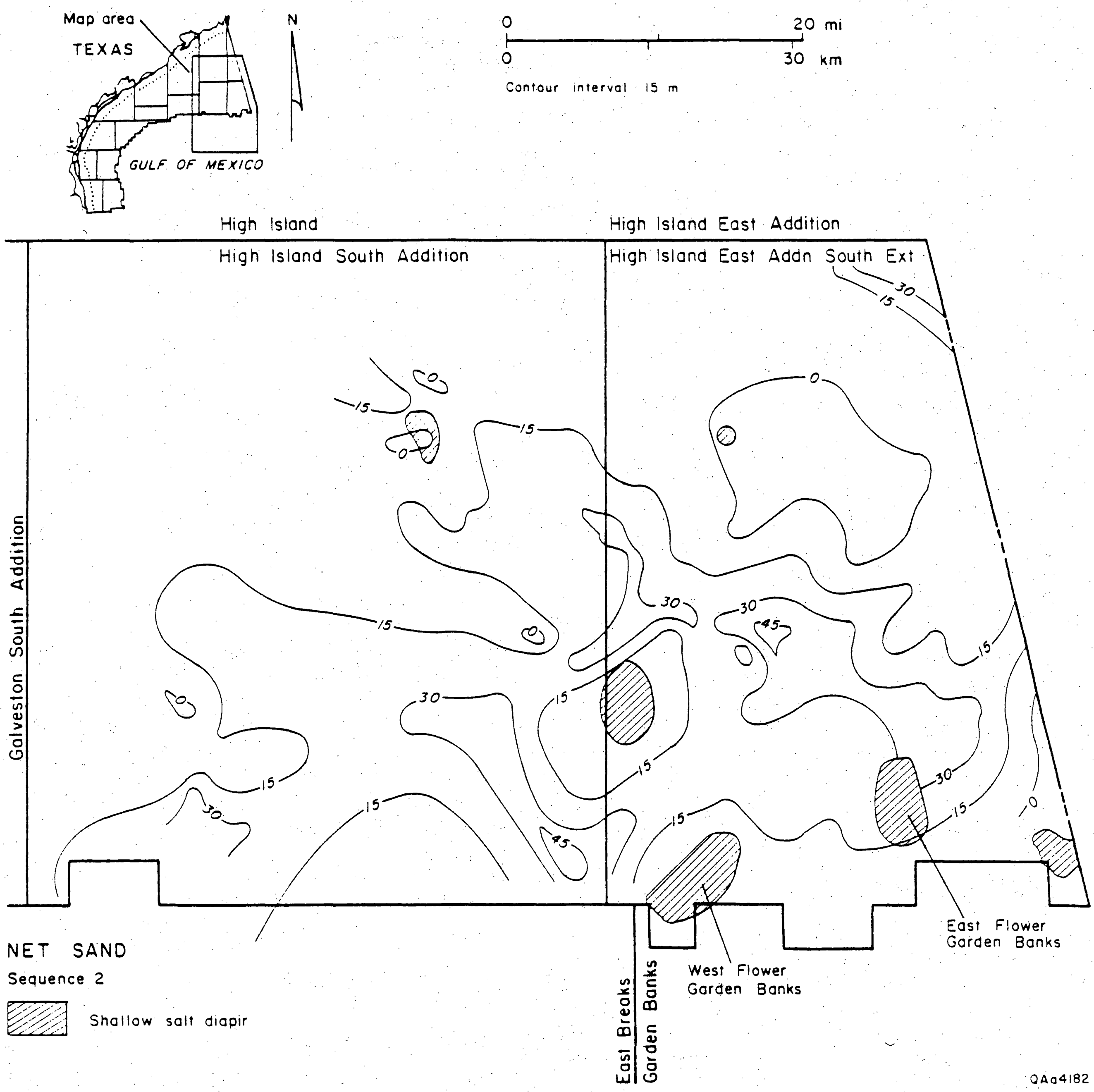

Figure 19. Net-sand isolith of early Wisconsin sequence (sequence 2).

Lithologic unit 3 is composed of sand and silty sand containing shell fragments and organic material concentrated near the top of the unit. The uppermost lithologic unit consists of olive gray clay and silty clay containing 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. This paleosoil is not widespread like the soil at the top of sequence 1. If the paleosoil of sequence 2 was laterally continuous originally, then it was largely removed by submarine erosion during the isotopic stage 3 transgression.

Sand Distribution

Sand, irregularly distributed in sequence 2, ranges in thickness from 0 to more than 45 m (fig. 19). The thickest accumulations of sand (>30 m) have diverse orientations that reflect the strong structural overprint. Sand bodies in sequence 2 concentrate mostly near the top and in the middle of the sequence (figs. 8, 10, 12, and 14). Sand only occurs in the lower third of sequence 2, where deep channels are indicated on the seismic profiles. Sand thickness is also partly related to the underlying distribution of sand in sequence 1. Sand bodies in sequence 2 are commonly thickest were sand bodies are thin in sequence 1 (figs. 8, 10, 12, and 14). In general, sequence 2 sand bodies are laterally continuous, especially the sand bodies associated with delta complex 2. Where sand bodies are discontinuous, they are mostly offset laterally as a result of structural interference or offlap related to parasequence boundaries. However, some of the anomalously thick sand deposits occur where fluvial channels 2 and 3 are vertically stacked.

Sequence 2 consists of three delta complexes that are identified on the basis of both seismic reflection patterns and lithologies (figs. 20 and 21). Delta complex 1 is composed of thick prodelta muds that are inclined to the southwest in the direction of progradation (fig. 21). To the east, these muddy deposits are transected by a large fluvial-channel system that fed deltaic complexes 2 and 3 (fig. 21). The thickest sand bodies are associated with delta complexes 2 and 3 (figs. 8, 10, 12, and 14). Delta complex 2 prograded predominantly to the

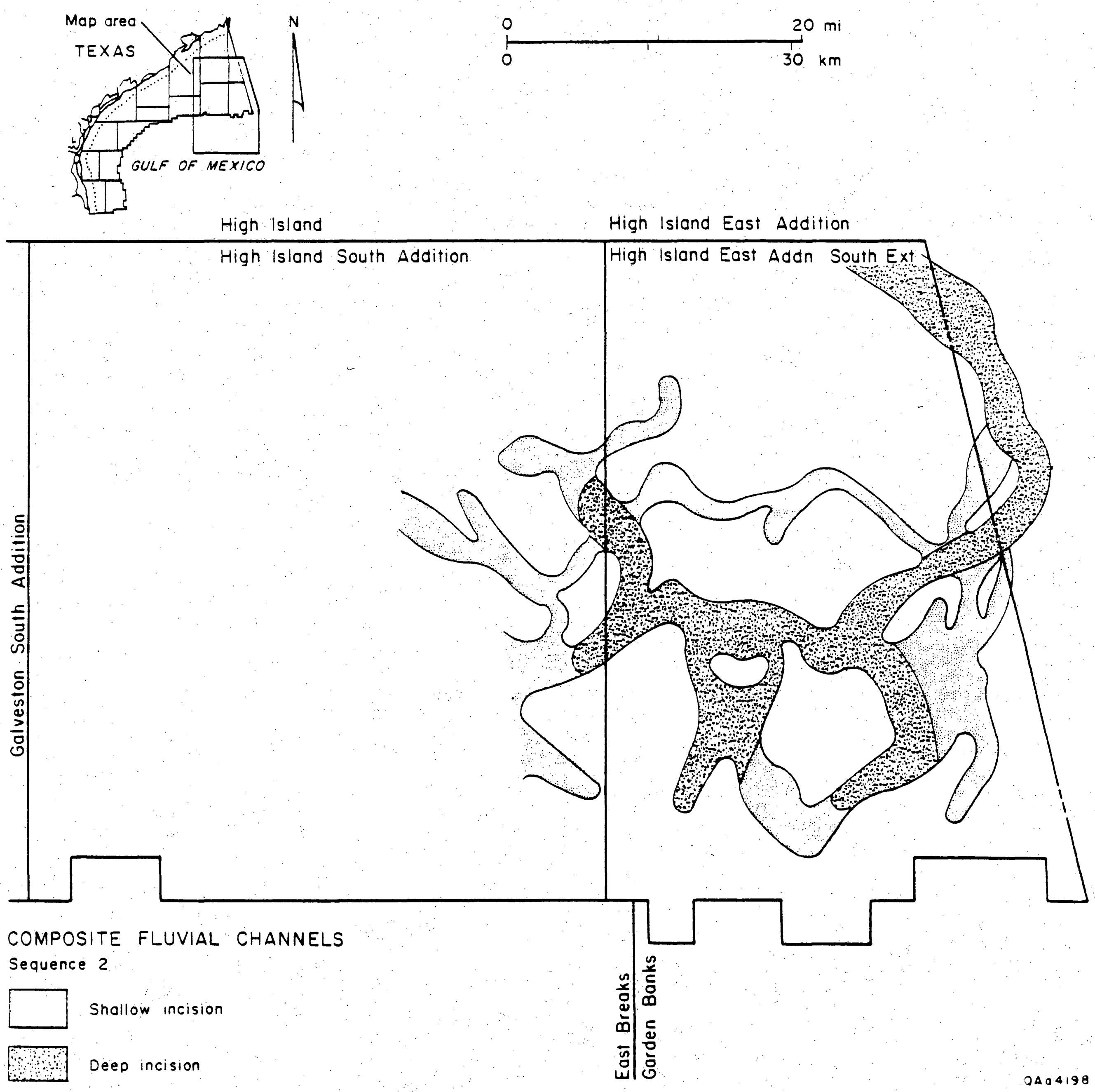

Figure 20. General sizes and locations of early Wisconsin fluvial channels.

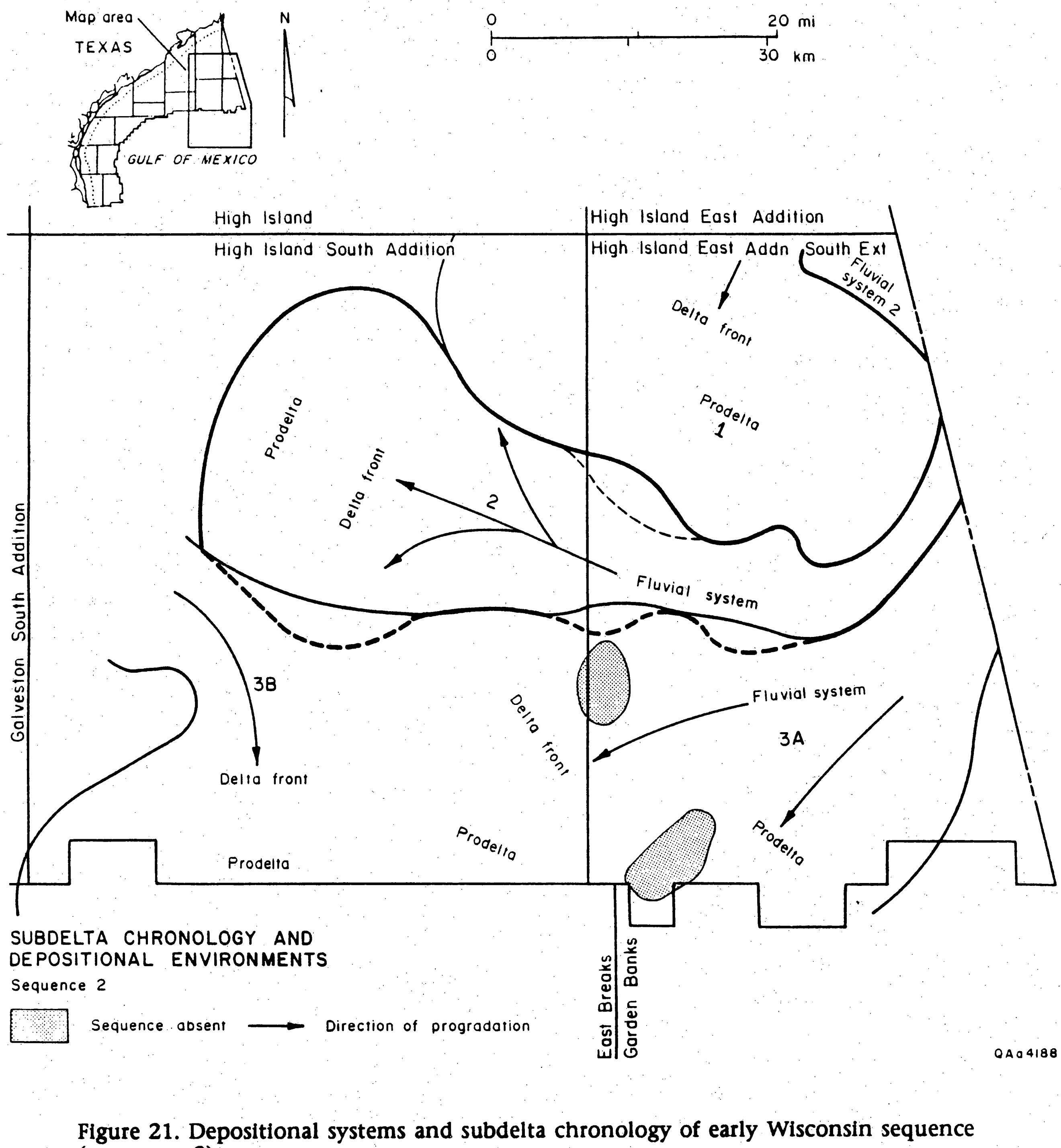

Figure 21. Depositional systems and subdelta chronology of early Wisconsin sequence (sequence 2).

west and onlapped complex 1 to the north (fig. 21). The parasequence boundary between delta complex 1 and delta complex 2 is distinct, but a distinct boundary separating complex 2 from complex 3 is unclear. Fluvial-deltaic complex 3 also prograded to the west, as demonstrated by chaotic seismic reflection patterns and clinoforms dipping to the west.

Overall sand abundance in sequence 2 decreases near the *Trimosina* fault zone (figs. 3 and 19) except locally, where the delta-front deposits of delta complex 3 abruptly increase in thickness and dip (figs. 7 and 8). These anomalies mark the positions of the former unstable shelf margin, where slumping and other mass transport processes transferred sand from the outer shelf to the upper slope. At these sites the downdip limit of sand was not penetrated even by the deepest borings. Sand-body heterogeneities are introduced as a result of delta-front resedimentation at the shelf margin.

Depositional History

The former delta plain and upper soil horizon of sequence 2 were submerged as a result of subsidence and the isotopic stage 3 transgression. Following the transgression and highstand, sea level began to fall as continental glaciers expanded. Sequence 2 is interpreted as a regressive fluvial-deltaic sequence deposited during the isotopic stage 4 falling phase and lowstand. At least 25 to 40 m of water remained over the shelf-margin platform even after sea level fell during the stage 4 period of glaciation. These water depths are indicated by the heights of clinoform reflections, where the sequence 2 deltas prograded onto the platform constructed by the pre-Wisconsin deltas. The predominant westerly direction of progradation of sequence 2 cuts across the depositional grain of the pre-Wisconsin sequence.

Filled fluvial channels in sequence 2 are about 38 m deep and 5 km wide (figs. 10, 12, and 20), representing a large fluvial system comparable in size 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 older extrabasinal rivers constructed and later filled large channels. The fluvial sands are gray, indicating a reducing environment of deposition, rather than tan or brown, which would indicate an oxidizing environment. The channel deposits are composite fills representing several episodes of stream reoccupation as the system aggraded in response to a relative rise in sea level.

Individual delta complexes of sequence 2 retain much of their fluvial dominance except near the paleoshelf margin, where waves and longshore currents partly reworked the sands and deposited them across the delta front. Despite the wave modifications, thick individual river-mouth deposits can still be observed at the shelf margin (figs. 8 and 19).

Sequence 3 (Late Wisconsin–Holocene)

Sequence Thickness and Composition

The deposition of sequence 3 was controlled by sea-level position, paleogeography, and contemporaneous structural deformation. Sequence 3 is either absent or less than 20 m thick on the upthrown side of the *Trimosina* fault zone (fig. 22). Where it is present landward of the fault zone, the sequence onlaps broad, structurally controlled depressions and troughs (figs. 7 through 14) that were created by late salt withdrawal and coastal-plain subsidence. On the continental platform, where sequence 3 is thin and fills broad sags, it is composed mostly of gray clay (figs. 10 and 12).

Basinward of the *Trimosina* fault zone, the sequence thickens to more than 120 m (figs. 7 through 14 and 22) as a result of progradation into relatively deep water at the shelf edge, where high subsidence rates along the faults added new accommodation space. This suggests that the fault zone had some relief on the seafloor and was being displaced while sequence 3 was deposited. The thickest part of sequence 3 (figs. 4 and 22) is associated with (1) the brow of the delta constructed during maximum progradation, (2) counterregional faults and adjacent

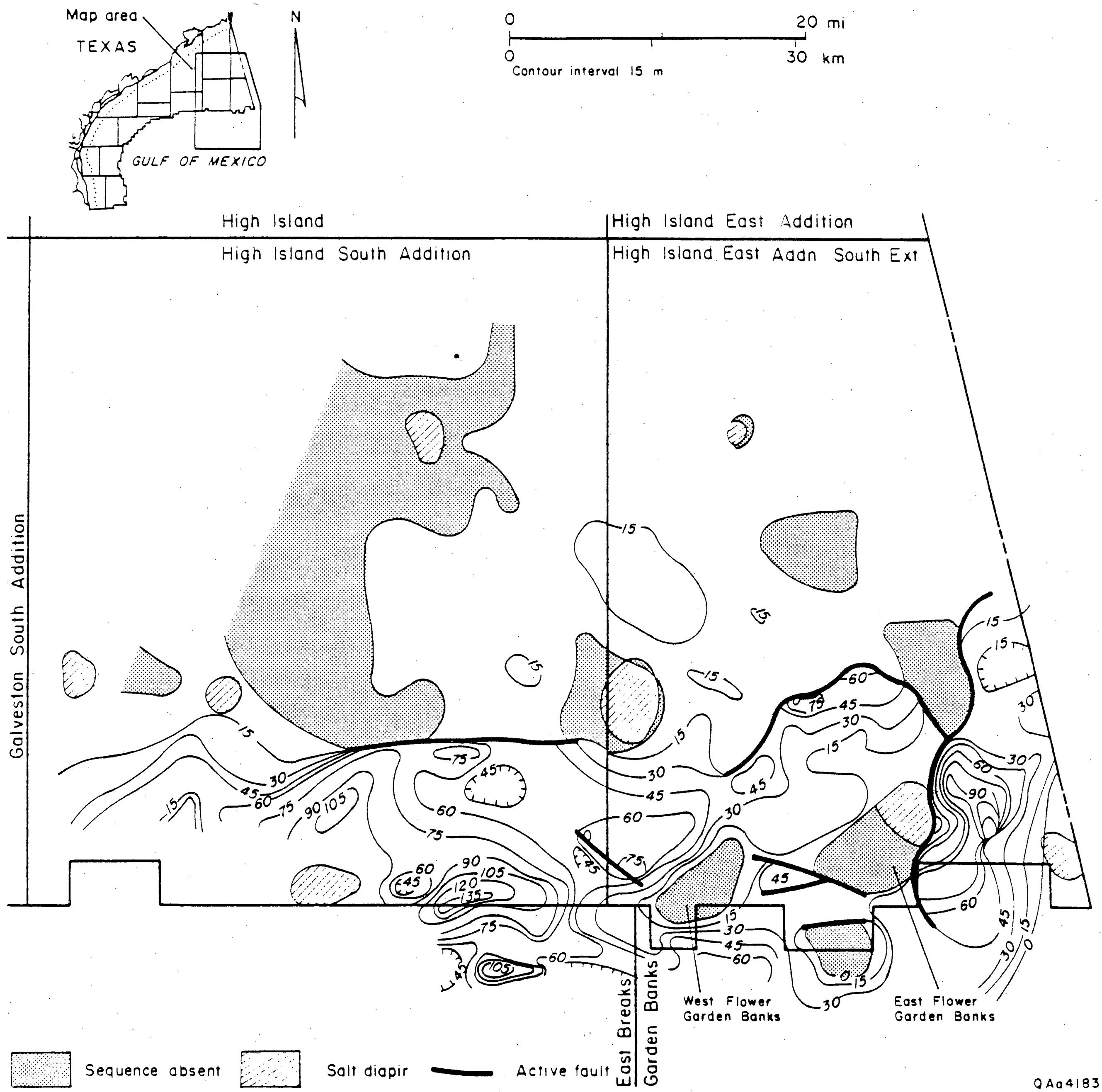

Figure 22. Thickness of late Wisconsin sequence (sequence 3).

bathymetric highs that acted as sediment dams, or (3) bathymetric lows created by local subsidence in salt-withdrawal minibasins (fig. 5).

Where sequence 3 is greatly expanded near the contemporaneous shelf margin, it is composed mostly of two lithologic units. The thick lower unit consists of olive gray clay containing thin interlayered beds of sand and silt (figs. 7, 8, 13, and 14). 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.

Transgressive deposits of sequence 3 are from 0 to 10 m thick; however, most of the borings penetrate from 3 to 6 m of very soft olive gray clay and fine sand and silty sand containing variable concentrations of shell fragments. Thickness of these young reworked transgressive deposits is also partly controlled by recent deformation. Transgressive deposits are thin over recent structural uplifts (topographic highs) and thicker in lows created by recent subsidence along reactivated faults. In some areas, transgressive deposits are completely absent over topographic highs (figs. 7 through 14).

Only the sequence 3 transgressive sediments were correlated strictly on the basis of lithology and physical properties. This deviation from established correlation procedures was necessary because the transgressive sediments are so thin that their seismic record is obscured by the broad width of the seismic bubble pulse. The late transgressive deposits of sequence 3 are recognized using sediment composition, induration, and water saturation of the muds. Generally very soft to soft, they contain as much as 80 percent water and 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. Descriptions of a few borings suggest that the transgressive deposits at some sites are composed of both sand and mud. These variable 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.

Sand Distribution

Sand deposits in sequence 3 are either relatively thin dip-oriented fluvial channels or moderately thick strike-oriented delta-front deposits (fig. 23). The thickest sand deposits are restricted to the shelf margin and generally have an east-west orientation that reflects progradation directions as well as wave reworking and alongshore redistribution (fig. 23). Landward of the *Trimosina* fault zone, the sequence contains less than 10 m of sand, which is associated with an elongate trend that coincides with a structural low (fig. 4). Where the sequence is expanded downdip of the fault zone, sand bodies near the top of the sequence have variable continuities and thicknesses that are directly proportional to the available accommodation space (rates of subsidence). In general the sand in sequence 3 is thickest where sand bodies in sequence 2 are also thick (fig. 8). Most of the sand is associated with a single sand body that is massive updip and becomes interbedded with mud near the shelf margin (fig. 14). These thin interbedded and discontinuous sand beds in the prodelta facies may be examples of shingled turbidites described by Vail and Wornardt (1991) and Lindsay and others (1984).

Sequence 3 was deposited by two delta complexes that were entirely controlled by structural lows and the *Trimosina* fault zone (fig. 24). The oldest delta complex prograded to the south and to the east along the fault-controlled shelf margin. The delta was later overlapped by the second delta complex that also prograded to the southeast. Delta complex 2 was supplied by two fluvial systems that eroded into the top of sequence 2 as sea level fell during the isotopic stage 2 glaciation.

The transgressive sand bodies of sequence 3 are thin, highly discontinuous, and patchy (figs. 9 through 14 and 23). They represent reworked beach sands and possibly inner-shelf shoals that were constructed as the beach eroded during the isotopic stage 1 transgression.

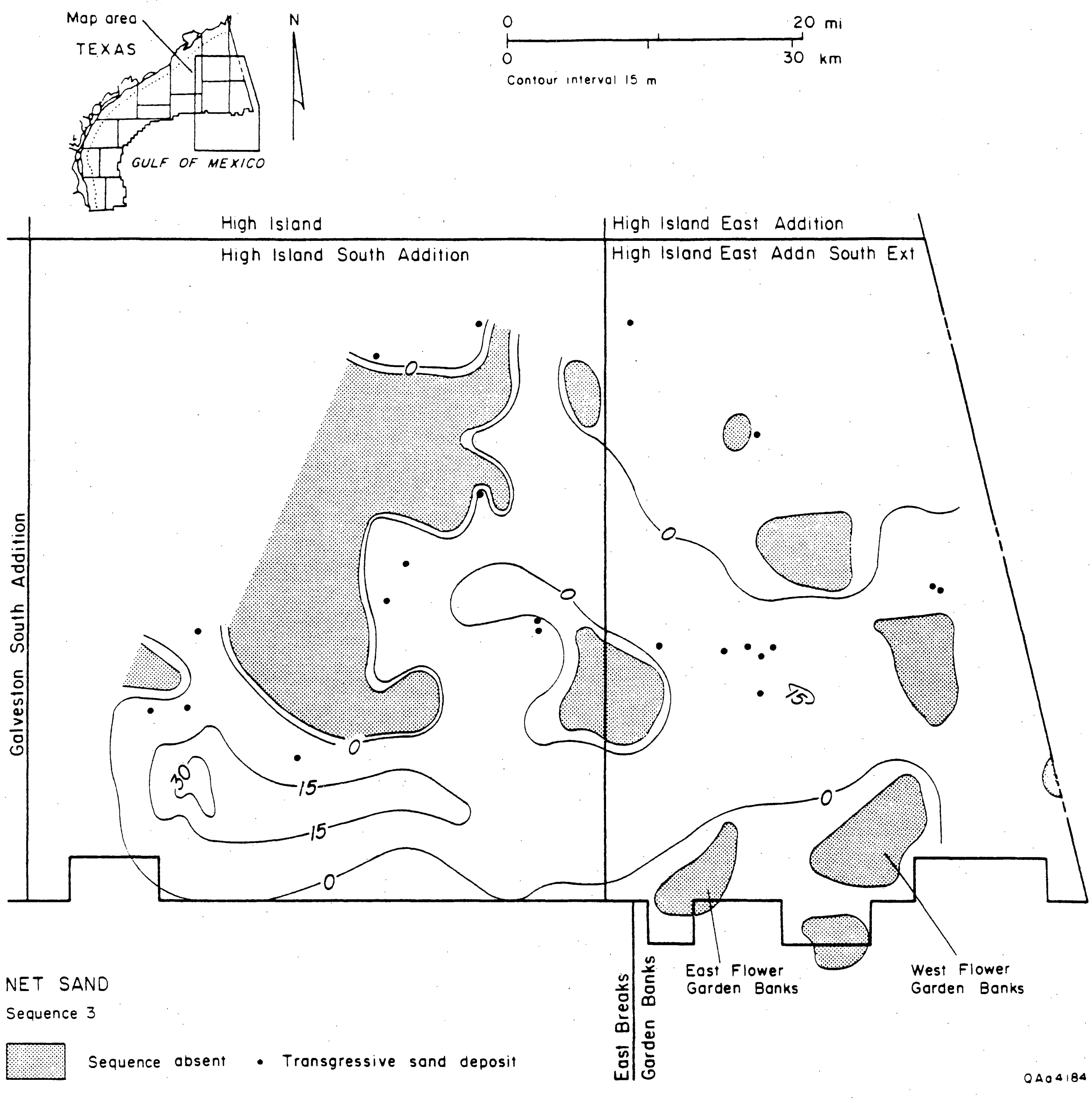

Figure 23. Net-sand isolith of late Wisconsin sequence (sequence 3).

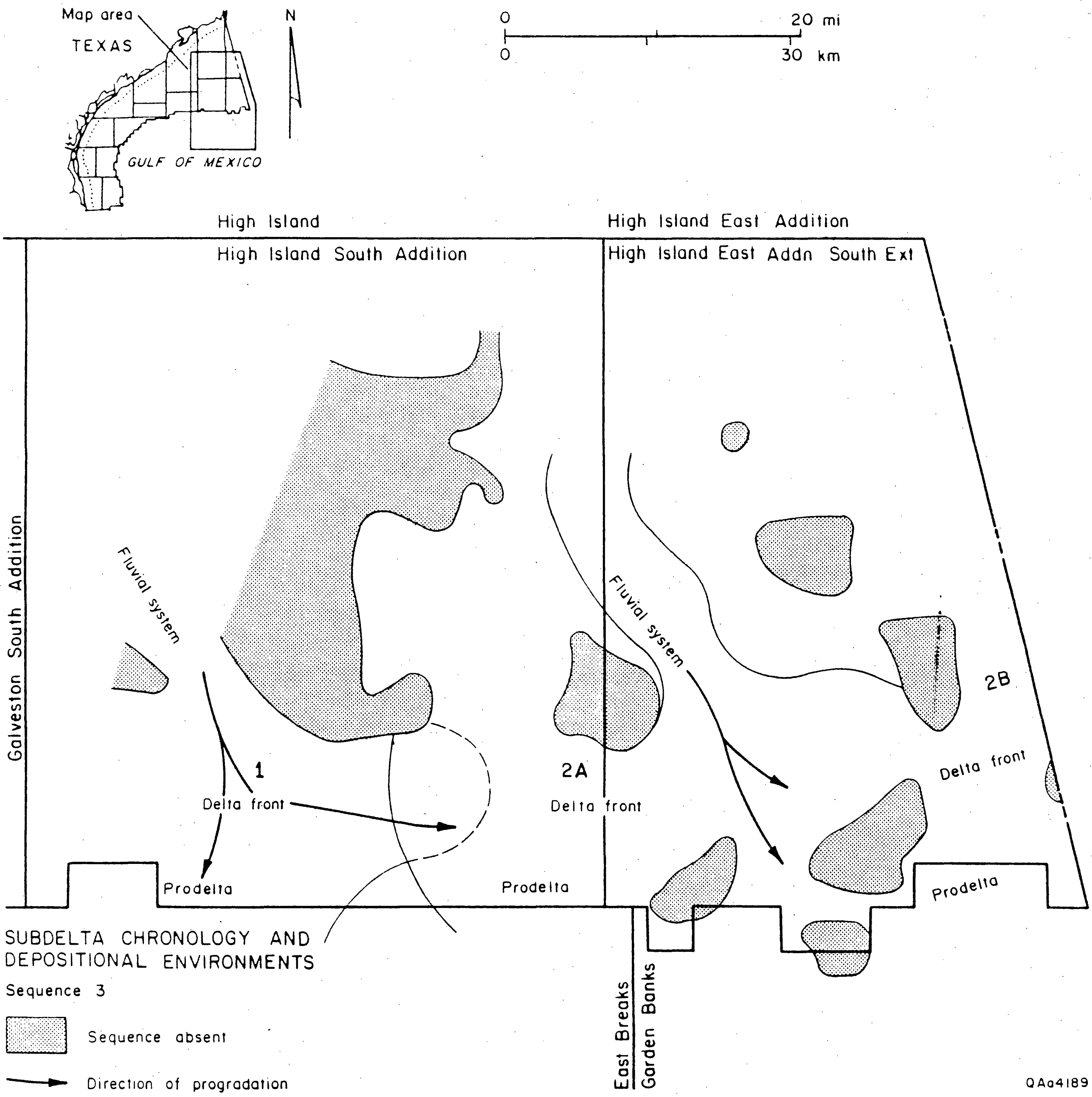

Figure 24. Depositional systems and subdelta chronology of late Wisconsin sequence (sequence 3).

Depositional History

The clinoform reflections, upward-coarsening facies successions, and paleogeographic setting indicate that sequence 3 is composed of thick regressive deltaic deposits overlain by much thinner transgressive deposits. Deltaic progradation accompanied a falling sea level and lowstand, whereas the transgression occurred during a eustatic rise in sea level that was accelerated by continued subsidence especially basinward of the *Trimosina* fault zone. Fault scarps on the present seafloor more than 10 m high (fig. 9) indicate that fault displacement continued after deposition of sequence 3 ended.

The southerly or easterly component of progradation in the late Wisconsin sequence cuts across the depositional grain of the early Wisconsin sequence. As shown by heights of clinoforms, the late Wisconsin delta system prograded into water that was 60 to 90 m deep. Despite steep depositional slopes, the muddy prodelta deposits are only locally contorted and exhibit only minor horizontal displacement.

The thin transgressive deposits of sequence 3 overlie the regressive deposits of the same sequence, and they onlap sequence 2 landward of the *Timosina* fault zone. If preserved by the next regressive phase of deposition, they will be the only depositional record of sequence 2 over much of the continental shelf and they could easily be misidentified because of their similarity to the transgressive deposits at the top of sequence 2.

ORIGINS OF SHELF-MARGIN DELTAS

The late Quaternary shelf-margin deltas of the western Gulf Coast Basin illustrate how durations of sea-level phases and syndepositional structures influence the development and distribution of individual sequences and the sedimentary facies within the sequences. Each of the shelf-margin delta systems displays unique depositional characteristics such as sand abundance, progradation direction, lobe geometry, and degree of syndepositional deformation. The shelf-margin deltas are products of relatively rapid falls in sea level (fig. 2) that produced

Type 1 unconformities (Posamentier and others, 1988) and well-defined incised valleys. Some of these incised valleys were mapped by Nelson and Bray (1970), Suter (1987), and Thomas and Anderson (1988). Despite repeated relatively rapid falls in base level, no evidence exists of submarine canyons (or other zones of sediment bypass along the Texas shelf margin) that would have supplied submarine fans on the slope or basin floor during each lowstand event. As the continental margin prograded, the sandy facies of each successive sequence was deposited slightly farther basinward, and each shelf-margin delta represents the most basinward position of thickest sandy sediments. However, each shelf-margin delta system possessed a different potential for downslope transport of sand into deep water. Only sequence 2 seems to have caused slumping and mass transport of large volumes of sand onto the adjacent upper slope.

The pre-Wisconsin sequence was deposited by a sand-rich delta system that prograded southwesterly, being only slightly influenced by contemporaneous structures or antecedent topography. The delta complex was fed by multiple distributaries that deposited abundant sand in both fluvial and delta-front environments. This delta system contains more sand and a greater concentration of sand than the other two sequences (figs. 16, 19, and 23).

The early Wisconsin delta system is composed of multiple lobes that generally prograded to the west and southwest. An exception was an intermediate lobe where sediment transport was deflected by salt diapirs, which caused some progradation to the north (fig. 21). Sand bodies of the early Wisconsin delta system are mostly elongate parallel to the channel axes. The locations of the channel axes, and thus the locations of sand bodies, were partly influenced by syndepositional structures. Fault escarpments and diapirs protruding on the upper slope directly controlled location of the wave-modified river-dominated delta system that deposited sequence 2.

The late Wisconsin sequence was deposited by a mud-rich, river-dominated delta system that was also greatly influenced by contemporaneous structures. Accommodation space was extremely limited as a result of lowered sea level. The late Wisconsin delta system consequently was trapped between the *Trimosina* growth-fault escarpment and large salt diapirs protruding on the upper slope, which limited the basinward deposition of the subdeltas. The delta system also prograded southeasterly, in contrast to all of the other systems that had a strong westerly component of progradation. The progressive decrease in abundance and concentration of sand between sequences 2 and 3 was probably a result of changing stream load or changing hydrodynamic efficiency of nearshore processes.

DISCUSSION

Stacking patterns of the stratigraphic sequences can be inferred by the spatial arrangement of lithofacies and seismic reflection patterns of each sequence. The spatial evidence indicates that the late Wisconsin sequence did not prograde as far basinward as did the pre-Wisconsin and early Wisconsin sequences. These inferences are confirmed by the CDP seismic profiles (fig. 15), which show that the package of clinoform reflections of the late Wisconsin sequence is landward of the same seismic facies of the older sequences. This backstepping stacking pattern represents a retrogradational phase of deposition that was most likely caused by progressively shorter periods of sea-level lowstand and a reduction in sediment supply. Judging from the reconstruction of sea-level positions (fig. 3), the late Wisconsin fall in sea level was fast enough and low enough that delta progradation beyond the early Wisconsin shelf margin would have been expected if the duration of lowstand was longer and the sediment supply remained the same as during deposition of the older sequences.

Delta morphology, lateral boundaries of delta lobes, channel positions, and locations of interfluves were all partly controlled by active structural features. On passive unstable shelf margins the influence of structures on depositional patterns generally decreases with time as a stable platform is constructed and underlying mobile sediments are displaced. However, the late Quaternary depositional sequences of the High Island area record a progressive increase in structural influence so that the youngest sequences are much more confined than older sequences. The thickness and lithologies of the pre-Wisconsin sequence are essentially unaffected by local structure, whereas the thickness and lithologies of the late Wisconsin sequence are highly variable and entirely controlled by local faulting and subsidence.

Rapid subsidence and concomitant deposition of the late Wisconsin sequence on the downthrown side of the *Trimosina* fault zone prevented reworking by marine processes and wider distribution of sand along depositional strike. Preservation of the sequence by rapid subsidence along the fault escarpment and prevention of its deposition across the platform illustrate how sequences are expanded at faults and the interval is thin or absent updip.

Average thickness of each sequence is about 60 to 75 m, depending on location relative to the shelf margin and the extent to which structural activity locally controlled accommodation space. Sequence thickness also depends on postdepositional preservation, which is related to the interplay between sea-level fluctuations and structural dynamics. The tops of each sequence are truncated and missing where postdepositional erosion occurred over a structural high. Erosional truncation or entire absence of a sequence is progressively greater for each younger sequence. The sequences are thickest where shelf-margin deposition was not greatly influenced by contemporaneous structures and the physiographic break in slope coincides with the brow of the delta at the position of maximum progradation.

The general uniformity of sequence thickness and facies architecture between sequences 1 and 2 indicates that the processes controlling progradation of those deltas were essentially the same for each sequence. This also means that the rates of subsidence at the shelf margin and the rates of eustatic sea-level fluctuations were approximately the same for each sequence. Otherwise a difference could be observed in sequence distribution, as between sequences 2 and 3.

The fluvial, deltaic, and strandplain sand bodies of the shelf-margin deltas concentrate either within the middle or near the top of each seismic sequence. These stratigraphic positions of sandy facies relative to the sequence boundary appear to be different from those reported by Van Wagoner and others (1990) and Vail and Wornardt (1991). Those workers indicated that the 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 basinward. Sand bodies are located immediately above the sequence boundary (1) at extreme updip sites, where fluvial and estuarine sand bodies are deposited and (2) at extreme downdip sites, where submarine fans and other sand-rich turbidites are deposited. However, at the depositional shelf edge, deltaic progradation onto the sequence boundary causes deposition of the sand bodies above the sequence boundary at a distance at least equivalent to the water depth.

Syndepositional structures influenced the locations of fluvial and deltaic sand bodies, but most of the active faults only indirectly increased the thickness of sand within the same sand body. Instead, an increase in net sand on the downthrown side of a fault was caused by preferential location of fluvial channels or deposition of additional sand bodies in response to more rapid subsidence and increased accommodation space (figs. 8, 10, 12, and 14). Apparent thinning of sand bodies on the upthrown side of a fault or over a structural high is typically caused by truncation of the top of the sequence and erosion of the sand body. This removal of sand related to postdepositional erosion is common in sequence 2 (figs. 7 through 14). It is less common in sequence 3 because regressive sediments of this sequence were not deposited over structural highs.

Interpretation of the seismic profiles using sequence stratigraphic criteria that emphasize an erosional unconformity as a sequence boundary (Posamentier and others, 1988) would result in an interpreted geologic history for each shelf-margin sequence that is substantially different from the one presented. If the base of deepest channel incision within each sequence is interpreted as the sequence boundary, then the erosional surface would be the sediment bypass surface. The lowstand systems tracts would thus not have been encountered in our study, but would have been deposited farther basinward of the shelf margin. Some lowstand systems tract deposits would be represented by the basal fill within the fluvial system. The transgressive systems tract would be represented by most of the channel fill and the overlying marine mudstones of each regional transgression. Each highstand systems tract would include the progradational succession from the inferred downlap surface to the base of the overlying erosional surface (sequence boundary). Instead of using the erosional surface as the sequence boundary, we interpret the downlap surface as the correlative conformity of the sequence boundary and the overlying thick progradational deltaic wedge as the lowstand systems tract. The thin, patchy, marine reworked sediments at the top of each sequence are the transgressive systems tract and, if present, the highstand systems tract deposits are extremely thin marine muds that are indistinguishable from prodelta deposits of the early lowstand systems tract.

CONCLUSIONS

1. Compositional, structural, and depositional differences of each late Quaternary shelf-margin sequence reflect a unique combination of sediment supply, eustatic fluctuations, and subsidence near the shelf margin. Each sequence is composed of two components, a thick regressive succession that is overlain by a thin transgressive component.

2. Delta construction of the shelf margin was accomplished by progradation partly oblique to rather than entirely perpendicular to the shelf-slope break.

3. Contemporaneous structural deformation controlled the thickness of each sequence, the directions of delta progradation, and the locations of major fluvial channels. Structural features also partly controlled the lapout positions of parasequences. Structural influence on delta geometries and facies patterns progressively increased with time. The oldest sequence was only slightly influenced by syndepositional structures, whereas the youngest sequence was dominated by active faults and salt diapirs.

4. On common-depth-point seismic profiles, a downlap surface is inferred by the termination of reflections at the toes of the clinoform reflections. However, high-resolution seismic profiles reveal that clinoform reflections actually become asymptotic at their toes, the asymptotes forming a series of parallel, high-amplitude reflections. The downlap surface is thus

not a true surface, but a zone that becomes progressively higher and younger in the direction of progradation.

5. The downlap zones are composed of marine muds without high concentrations of shell debris, as would be expected in condensed sections. Instead, the shell beds in the deltaic deposits represent minor depositional hiatuses that coincide with local abandonment surfaces within the prodelta lithofacies or corresponding clinoform seismic facies. These shell zones are discontinuous and uncorrelatable with regional condensed sections.

6. Evidence of submarine erosion and reworking of the delta surface during transgression (ravinement surface) is not widely observed probably because rapid subsidence coupled with rapid eustatic sea-level rise quickly submerged the delta plain below wave base and the depth of effective wave reworking. If the most recent ravinement surface is preserved, it is obscured by the bubble pulse in the seismic records.

7. The early Wisconsin deltaic sequence exhibits two different types of fluvial channels. The largest channels are deep, nested channel complexes that record multiple phases of alluvial incision and fill. These large meandering channels commonly cut through much of the sequence but rarely are incised below the basal sequence boundary. The smaller channels are shallow, coinciding with the tops of the clinoform reflections. The shallow channels appear to be associated with normal delta progradation and the superposition of distributary channels over the delta-front facies.

8. No evidence exists that incised valleys or submarine canyons formed along the paleoshelf margin, even though moderately large rivers were present and sea-level curves indicate several periods during the past 100,000 yr when sea level fell rapidly.

9. Parasequence boundaries separating the subdeltas of each delta complex are ill defined because they are essentially conformable reflections. Away from structural highs, onlap is rarely observed that would indicate distinctly different times of deposition onto a preexisting surface.

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