A Thesis

entitled

A Flood-Tidal Delta Complex, The Holocene/Pleistocene Boundary, and Seismic Stratigraphy in the Quaternary Section off the Southern Assateague Island Coast, Virginia, USA

by

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Submitted to the Graduate Faculty as partial fulfillment of the requirements for the Master of Science Degree in Geology

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An Abstract of

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The Atlantic inner-shelf off the coast of Assateague Island preserves a complex stratigraphy in the Quaternary Section. Holocene highstands have been transgressing the island, yielding the current state. Seismic Reflection data off the Assateague Island coast reveal flood-tidal delta facies, tidal inlet channels, tidal creeks, and Holocene sand sheets. These facies indicate a former back-barrier environment off the Assateague Island coast, revealing a former offshore position of Assateague Island. The defining characteristic of the back-barrier environment is a flood-tidal delta complex approximately 4.5 km by 3.5 km. An isopach map of the flood-tidal delta reveals a ramp and facies thinning away from the ramp. Two main reflections are interpreted to be shoreface ravinements in the study area, which are time-transgressive subaqueous erosional surfaces produced during the landward movement of the transgressive systems tract. A third reflection is interpreted to be the pre-transgressive surface, marking the Holocene/Pleistocene Boundary. The depth of the pre-transgressive surface was interpreted throughout the entire study area. It

resulted in a clean surface with a channel-like depression near seismic line 18. This depression is interpreted as being part of a lowstand drainage system.

Dedication

I dedicate this thesis to my family and friends, the people that have given me so much and have always had faith in me.

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I am very thankful to the many people that assisted me on this project. First, I am very grateful for all the thoughts and time from my advisor, Dr. David Krantz. His intellect and teaching were instrumental in the success of my project. Also, I would like to thank committee members Dr. Ricky Becker for his GIS expertise, and Dr. Alison Spongberg for her feedback.

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List of Abbreviations

DEM………………..Digital Elevation Model

GPR…………………Ground Penetrating Radar

HST…………………Highstand Systems Tract

ka…………………....Thousand Years Ago

LGM………………...Last Glacial Maximum
LIDAR……………...Light Detection and Ranging
LST………………..…Lowstand Systems Tract

Ma…………………..Millions of Years Ago

NOAA………………National Oceanic and Atmospheric Administration
NOS…………………National Ocean Survey

TST…………………Transgressive Systems Tract

USGS…………….…United States Geological Survey

List of Symbols

® ....Registered Trademark

Chapter 1

Introduction

The Atlantic coast of the Northeastern United States, specifically the Delmarva Peninsula, contains a complex offshore stratigraphy, giving insight into the sea-level history of the Quaternary Period. This stratigraphy preserves paludal, coastal, and inner-shelf environments from the late Pleistocene into the Holocene, created as a response to a dramatic decrease and then increase in sea level. The inner-shelf stratigraphy contains specific depositional facies environments such as: marine sand sheets, offshore shoals, backbarriers, overwash, tidal flats, inlet deltas, spits, inlet channel, paleochannels, and tidal creeks (Toscano et al., 1989; Foyle and Oertel, 1994; Wikel, 2008). This study focuses on Holocene tidal-inlet complexes and a Late Pleistocene incised valley in the Atlantic inner-shelf off southern Assateague Island.

Tidal inlets are a dynamic feature on modern coastlines. Inlets connect the ocean to backbarrier lagoons, areas where sediment is transported then stored. The sediment dynamics of the inlets themselves are greatly influenced by waves, currents, and tides (Moslow and Heron, 1978; Finkelstein, 1986; FitzGerald, 1988; Buynevich et al., 2003; FitzGerald et al., 2008; Seminack, 2011; Seminack and Buynevich, 2013; Hayes and FitzGerald, 2013).

Assateague Island has had a number of inlets documented through its history. The inlets can remain open from decades to centuries. Figure 1-1 displays the historical inlets of Assateague Island from circa 1690 until present (Amrhein 1986; Krantz et al., 2009). The maps were compiled using early maps from European settlers on the East Coast. It is important to understand the genesis and closure processes of inlets to understand past, present, and future inlet systems as well as event chronology (i.e., storm processes and the rise and fall of sea level) (Banaszak, 2010; Seminack, 2011; Maike et al., 2013). Storms play an important role in creating and reworking inlet complexes. The stratigraphic evolution of inlets can be documented using geophysical techniques such as ground penetrating radar (GPR) or marine seismology (Buynevich et al., 2003; Buynevich and Donnelly, 2006; Seminack, 2011).

As sea level fell to a minimum at 22-20 ka, the modern Atlantic shelf was left exposed as a coastal plain. At the same time, river base level dropped dramatically to sea level at 125-130 m below present, and the regional rivers incised. These drainage systems appear in the stratigraphic record as identifiable incised paleovalleys that subsequently filled with sediment due to rising sea level during the Holocene. The major Quaternary paleovalleys of the Susquehanna River and tributaries have been mapped south of Assateague where they cross beneath the Virginia Eastern Shore (Colman et al., 1990; Oertel and Foyle, 1995; Foyle and Oertel, 1997). Paleovalley systems also have been explored crossing the inner shelf of Maryland (Toscano et al., 1989), and Delaware (Childers, 2013). This study will complement the Upper Pleistocene and Holocene stratigraphic context interpreted from other studies. The continued understanding of the ocean stratigraphy is a benefit to engineers and geologists as they search for nearshore sources of beach-nourishment sands and for sites to construct offshore wind turbines (Childers, 2013).

The primary source of data for this study are the 160 km of high-resolution seismic and side-scan sonar data collected in 2004 and 2005 by Drs. David Krantz and Douglas Levin off the coast of Assateague Island. The seismic profiles clearly represent the Holocene transgressive stratigraphy sitting on a foundation of Upper Pleistocene deposits. The data will be used to interpret the Assateague offshore stratigraphy. The study focuses on evaluating the Holocene-Pleistocene boundary (pre-transgressive surface) and mapping the Holocene sedimentary facies.

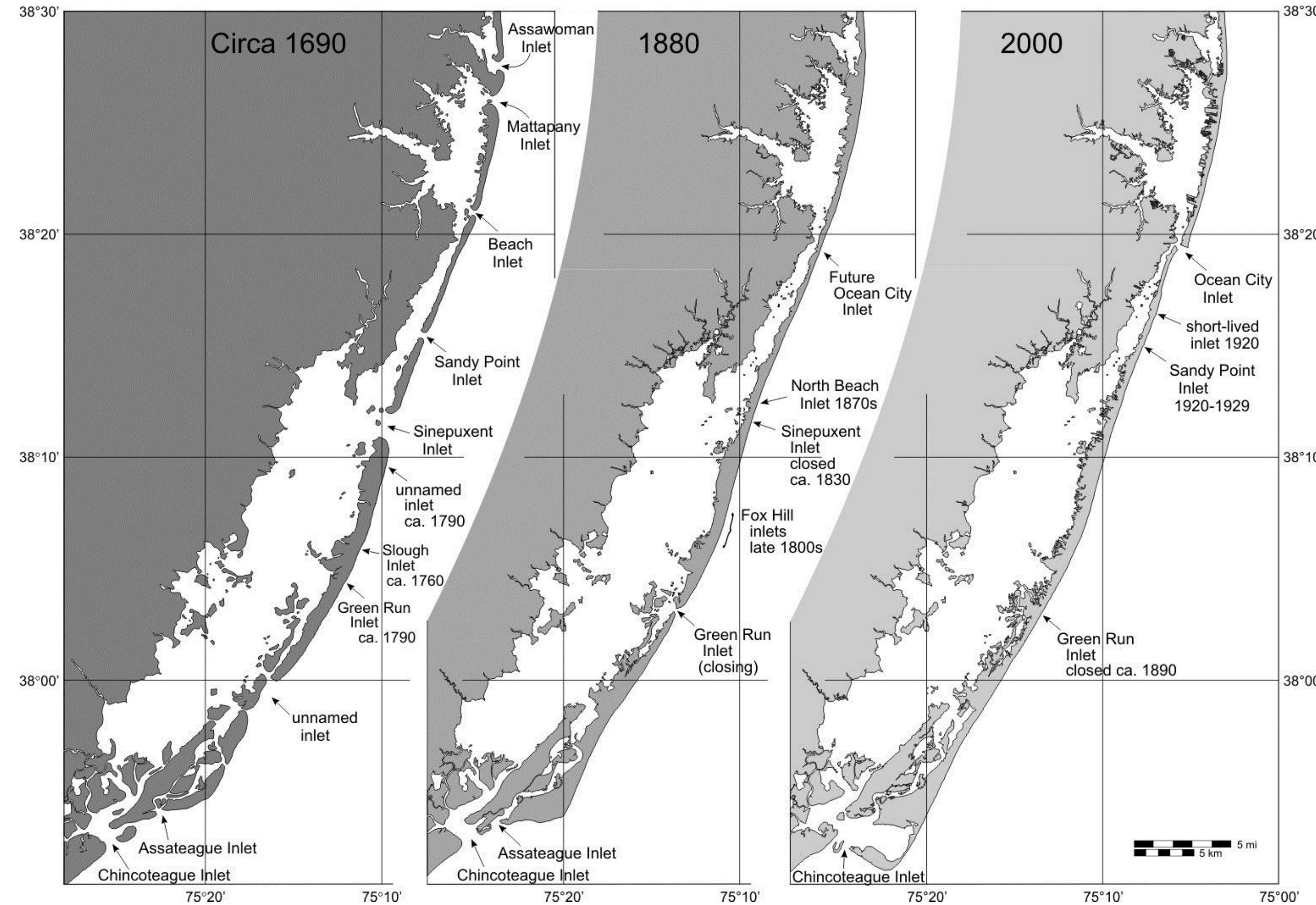


Figure 1-1
A map displaying historical inlets on Assateague Island from circa 1690 to present. Modified from Krantz et al. (2009).

Chapter 2

Background

Assateague Island is located on the Mid-Atlantic Coast of the United States within the states of Maryland and Virginia (Figure 2-1). Assateague is a long, linear, wave-dominated barrier island that is retrogradational in nature (Posamentier and Allen, 1999; Hobbs, 2004). The dominant direction of regional longshore transport is from north to south, due to the dominant wave direction. Nor'easters (winter storms) are the primary factor in the yearly sediment transport (Hobbs, 2004). These storms historically have played a major role controlling the morphodynamics and transgressive nature of the island.

The dynamics of the island can be seen through the changing morphology as well as the evolution of tidal-inlet channels. Assateague Island consists of historic and relict inlets such as Green Run, Sinepuxent, and Ocean City Inlets (Figure 1-1). Ocean City Inlet has been the ongoing project of the Army Corps of Engineers since it was breached by a hurricane in 1933. The inlet separates the northern end of Assateague Island from Ocean City, Maryland. The inlet is stabilized by jetties and maintained by dredging. This project has been an ongoing case study of sediment budgets and effects of jetties on coastal systems (Pendleton et al., 2004).

The other major inlet on Assateague Island is Chincoteague Inlet, which is a natural inlet on the far southern end of the island. This inlet is crucial to ship traffic in and out of Chincoteague Bay and for access to docks on Chincoteague Island. The navigation channel through the inlet is maintained by active dredging. From 1995-2006 roughly 473,000 m^3 of sediment was removed from Chincoteague Inlet (Hobbs, 2004).

Physical Setting

The Delmarva Peninsula formed during the Pliocene and Pleistocene Epochs (5-1.5 million years ago, Ma). Glaciers were located as close as 160 km north of the Delmarva Peninsula in eastern Pennsylvania (Schupp, 2013). As the glaciers retreated, the sediments that had been scraped off the landscape were deposited into the Delaware and Susquehanna river systems that transported them southeast to the Delmarva Peninsula (Krantz et al., 2009). These sediments were derived from the Precambrian bedrock of the Canadian Shield and the Paleozoic bedrock of the Appalachian Mountains. These sediments built the core of the Delmarva Peninsula, which consists of Pliocene fluvial, deltaic, and marginal marine deposits, and Pleistocene braided-river outwash. Pleistocene ocean shorelines are preserved as ridges near the coast on the Delmarva Peninsula, signifying high sea levels in the past.

The core of the central Delmarva Peninsula was built up mostly during the Pliocene as a series of lower delta-plain, deltaic, and marginal marine deposits associated with the upper Yorktown Formation and the Beaverdam Formation (Owens and Denny, 1979; Mixon, 1985). The land surface was further mantled by braided-river deposits as outwash from the distal end of the continental glaciers, producing the Columbia Formation in Delaware during the latest Pliocene or earliest Pleistocene (Groot and Jordan, 1999). Then sea-level highstands through the middle and late Pleistocene successively cut into and reworked these Pliocene and lower Pleistocene deposits. Highstand deposits are preserved as shorelines along the Atlantic coast of the Delmarva (Demarest and Leatherman, 1985; Toscano and York, 1992).

The Wisconsinan glaciation (22-20 ka) (the last glacial maximum, LGM) resulted in sea level falling 125 m below present (Fairbanks, 1989; Lambeck and Chappell, 2001; Lambeck et al., 2002). This caused a long exposure of the modern continental shelf as a broad coastal plain. The ocean shore at that time was approximately 100 km seaward of the present position (Figure 2-2) (Schupp, 2013). During the lowstands, intermediate river systems that drained central Delmarva flowed across the present continental shelf and downcut as much as 40-50 m (Foyle and Oertel, 1994). The lowstand incised valleys were subsequently, and progressively, filled as sea level rose after the LGM.

The melting of glaciers occurred once again ~18 ka, increasing sea level (Schupp, 2013). From 15 ka through 10 ka, sea level rose ~10 times faster than the present rate (Krantz et al., 2009; Schupp, 2013). The underlying geology serves as a record for these historic events. The Pliocene Yorktown Formation (5 to 3 Ma) underlies the Maryland and Virginia coast (Owens and Denny, 1979; Mixon, 1985). The upper Pliocene Beaverdam Formation overlies the Yorktown Formation north of approximately Wallops Island. This unit is important as the unconfined aquifer in the Assateague Island region. Moving upward stratigraphically, the Sinepuxent Formation, Late Pleistocene in age, lies beneath present day Assateague Island and consists of fine-grained shallow marine sediments (Owens and Denny, 1979). The general stratigraphy of the Assateague Island region is presented as a cross section in Figure 2-3, a time scale in Figure 2-4, and a stratigraphic column in Figure 2-5.

The Sinepuxent Formation occurs within the study area as the uppermost unit of the Pleistocene. The overall lithology is a dark, poorly sorted, very fine to medium sand, but, the lower part of the formation is fine grained with thinly bedded clay (Owens and Denny, 1979).

During the Holocene Epoch (11 ka – present), the barrier islands began a landward migration across the shelf to the present position (Krantz et al., 2009; Schupp, 2013). During this transgression, back-barrier lagoons such as Chincoteague Bay flooded as sea level rose then eventually slowed (Krantz et al., 2009; Schupp, 2013). The modern system appears to have been in place by approximately 5,000 years ago, although the timing is poorly constrained.

Assateague Island is composed primarily of sand. There are two sources likely for the active sediment input into the Assateague system. Storms cause erosion of the Atlantic shoreface down to depths of 11 to 13 m (Kraft, 1971). Another source is erosion of the northern Pleistocene headlands near Rehoboth Beach and Bethany Beach, Delaware (Honeycutt and Krantz, 2003; Hobbs, 2004; Schupp, 2013).

The inner-shelf shoals of Assateague Island, both shoreface-attached and shoreface-detached, are important as a regional sand resource and as a preserved record of the most recent phase of the Holocene transgression. The formation of shoals is somewhat debated, however a generally accepted mechanism was synthesized by McBride and Moslow (1991). This model involves the opening of an inlet on a barrier-island coast, inlet migration, waning of the inlet, inlet closure with wave reworking of the seaward sand, and finally ridge detachment (Figure 2-6) (McBride and Moslow, 1991). The offshore bathymetry of Assateague Island with the shoals can be seen in Figure 2-7.

Tidal-Inlet Geomorphology

Tidal inlets are a dynamic component of barrier island systems. They serve as a conduit between open-ocean and back-barrier environments (Hayes and Kana, 1976; Hayes and FitzGerald, 2013). They consist of 30-50% of all the sedimentation with the barrier island system (Hayes and Kana, 1976). The typical tidal inlet geomorphology is presented in Figures 2-8. The three main depositional features of tidal inlets are flood tidal deltas, ebb-tidal deltas, and recurved spits (Hayes and Boothroyd, 1969; Hayes and Kana, 1976).

Flood-tidal deltas are shoals of sediment deposited in the back barrier from tidal currents flowing landward through the inlet (the flooding tide). This sediment transport produces a deposit with a landward-migrating geometry. The deposition of flood-tidal deltas occurs as incoming ocean water encounters the calm lagoonal waters. The morphology of flood-tidal deltas is presented in Figure 2-8. The numbers below correlate to the labelled features in Figure 2-8 and the descriptions (Hayes and Kana, 1976; Hayes, 1980).

1) *Flood Ramp* — sandy seaward-facing slope consisting of sand waves showing orientation of the flood deposits;

2) *Flood Channel* — tidal channel that bifurcates around the flood ramp;
3) *Ebb Shield* — an elevated portion of the flood-tidal delta system, on the bay side of the ramp, that prevents re-working of the delta from ebb currents;
4) *Ebb Spit* — spits formed on the flanks of the flood-tidal delta from ebb currents;
5) *Spillover Lobe* — as defined by Hayes and Kana (1976), "lobate bodies of sediment formed by unidirectional currents."

As the tide flows back out of the back-barrier lagoon, sediments are carried seaward through the throat of the tidal inlet. This transport forms swash bars on the seaward side of the island, known as the ebb-tidal delta. Along the coast of Assateague Island, ebb-tidal deltas are not well developed seaward of open inlets, and not preserved once the inlet closes. This is due to the high wave energy that continually erodes any deposition on the ocean side of the island. A schematic of a wave-dominated tidal-inlet system can be seen in Figure 2-9.

Figure 2-1

Assateague Island Location Map. Assateague Island is located on Atlantic Coast of the Delmarva Peninsula within Virginia & Maryland, USA.

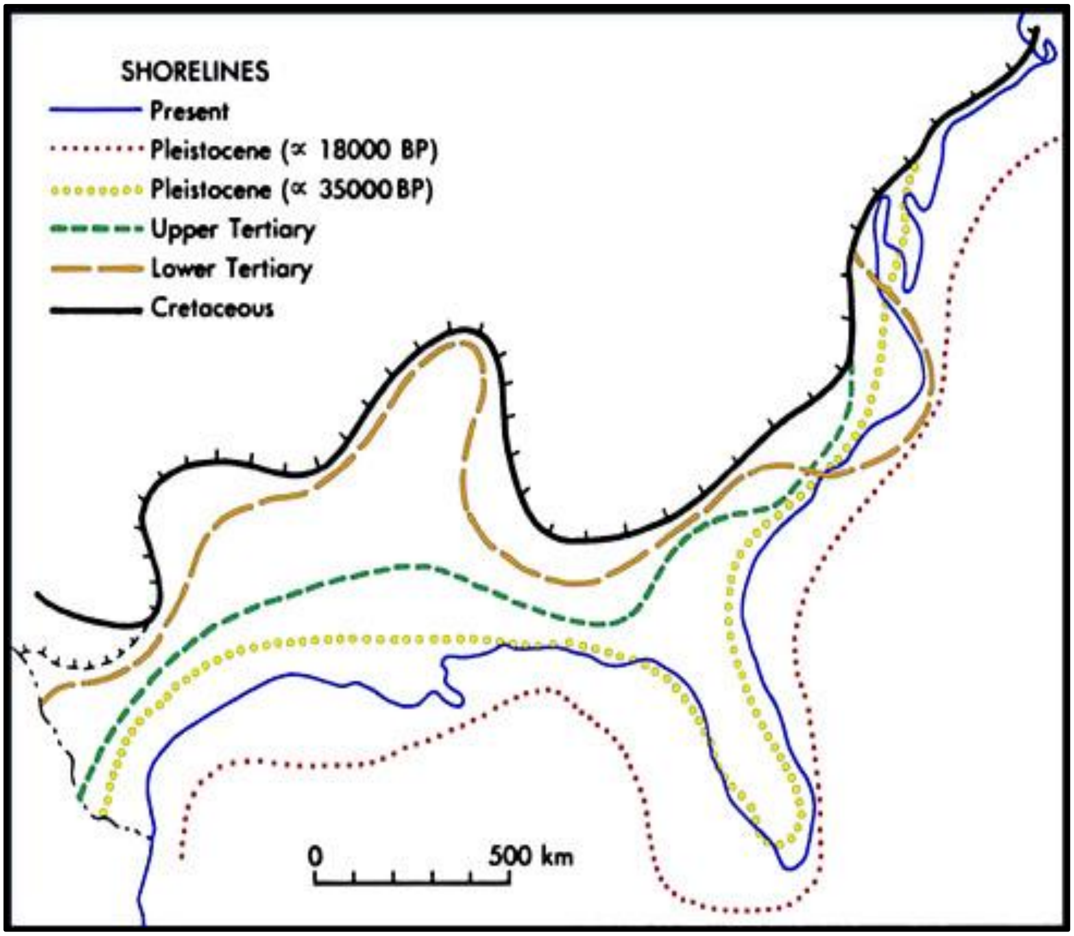


Figure 2-2

Ancestral Pleistocene shorelines giving insight into former glacial-interglacial cycles (NOAA, online data).

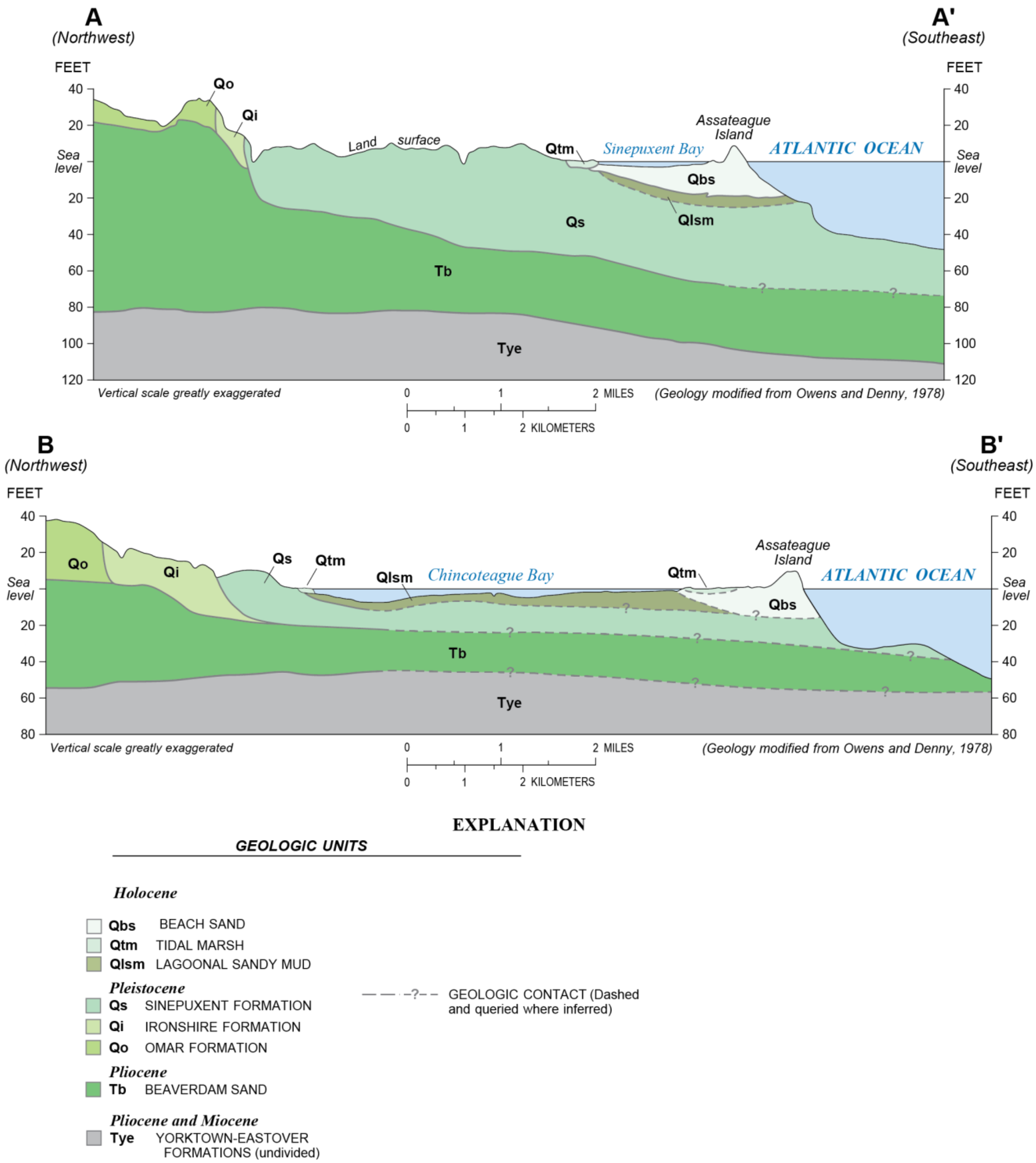


Figure 2-3

Cross section portrays Assateague Island in reference to the subsurface stratigraphy. Note: The Sinepuxent Formation appears in the seismic profiles of this study as the unit immediately below the base of the Holocene sequence. Modified from Owens and Denny (1979) and Krantz et al. (2009).

Era	Period	Epoch	Age	mya	Atlantic Coast Events
Cenozoic	Quaternary	Holocene			Barrier islands moved to their present position when the rate of sea level rise slowed and the low-lying basins flooded to create the Coastal Bays
				0.01	Continental glaciers began melting again, causing sea level to rise and push sediments westward from the (present-day) continental shelf (about 18,000 years ago)
		Pleistocene	Tarantian (Upper)	0.13	
			Ionian (Middle)	0.78	Last Glacial Maximum of the Wisconsinan glaciation (about 22,000-20,000 years ago); sea level dropped to 125 m below present; rivers incised the basins that would later become the Coastal Bays estuary
			Calabrian	1.8	
			Gelasian	2.6	
	Tertiary / Neogene	Pliocene	Piacenzian		
			Zanclean	5.3	Delmarva Peninsula began forming from glacial sediments being carried down rivers
		Miocene	Messinian		
			Tortonian		
			Serravallian		
			Langhian		
			Burdigalian		
			Aquitanian	23.0	

Figure 2-4

Geologic time scale of Atlantic Coastal events highlighting the Neogene until present (Schupp, 2013). This graphic was created by Trista Thornberry-Ehrlich (Colorado State University).

Era	Period		Epoch	Age*	Unit	Description	Approx. depth**
Cenozoic	Quaternary		Holocene	0.01–present	Barrier sand (Qbs); Tidal marsh (Qtm)	Barrier sand: light-colored, well-sorted, fine- to very coarse-grained feldspathic quartz sand with gravel and shell fragments; extensive cross bedding due to wave action; up to 6 m (20 ft) thick. Tidal marsh deposit: clay/silt with high organic matter; unconsolidated and soupy; less than 5 m thick; exposed along western (bay) side of island.	Surface to –4.5 m
						Lagoonal sandy mud/silt	–4.5 to –8 m
						Base of basal peat layer marks the base of Holocene deposits; sometimes exposed along bay side or (after large storms) interior, and chunks sometimes erode from shoreface and are carried onto beach.	–8 to –10 m
			Pleistocene	1.2–0.8	Sinepuxent Formation (Qs)	Marginal marine deposit. Coastal sequence of dark-colored, poorly sorted, silty, fine to medium sand with thin beds of peaty sand and black clay. Abundant heavy minerals (amphibole and pyroxene materials). All major clay mineral groups present (kaolinite, montmorillonite, illite, chlorite). Sand consists of quartz, feldspar, and abundance of mica (muscovite, biotite, and chlorite) that makes Qs lithologically distinct from older underlying units.	–10 to –23 m
	Tertiary	Neogene	Pliocene	24–1.8	Beaverdam Formation (Tb)	Medium sand with scattered beds of coarse sand, gravelly sand, and silty clay, interbedded with clay-silt laminae. Unweathered Beaverdam Sand sediments may be pale blue-green or white; weathered sediments are orange or reddish brown.	–23 to –30 m
			Pliocene–Miocene		Yorktown-Eastover formations (undivided) (Tye)	Lenticular silts, clays, and fine sand	Below –30 m
						Grey, medium- to fine- grained shelly sand	
			Paleocene				

Figure 2-5

Stratigraphic column of the Assateague Island region. The graphic was created from historical data within the region.

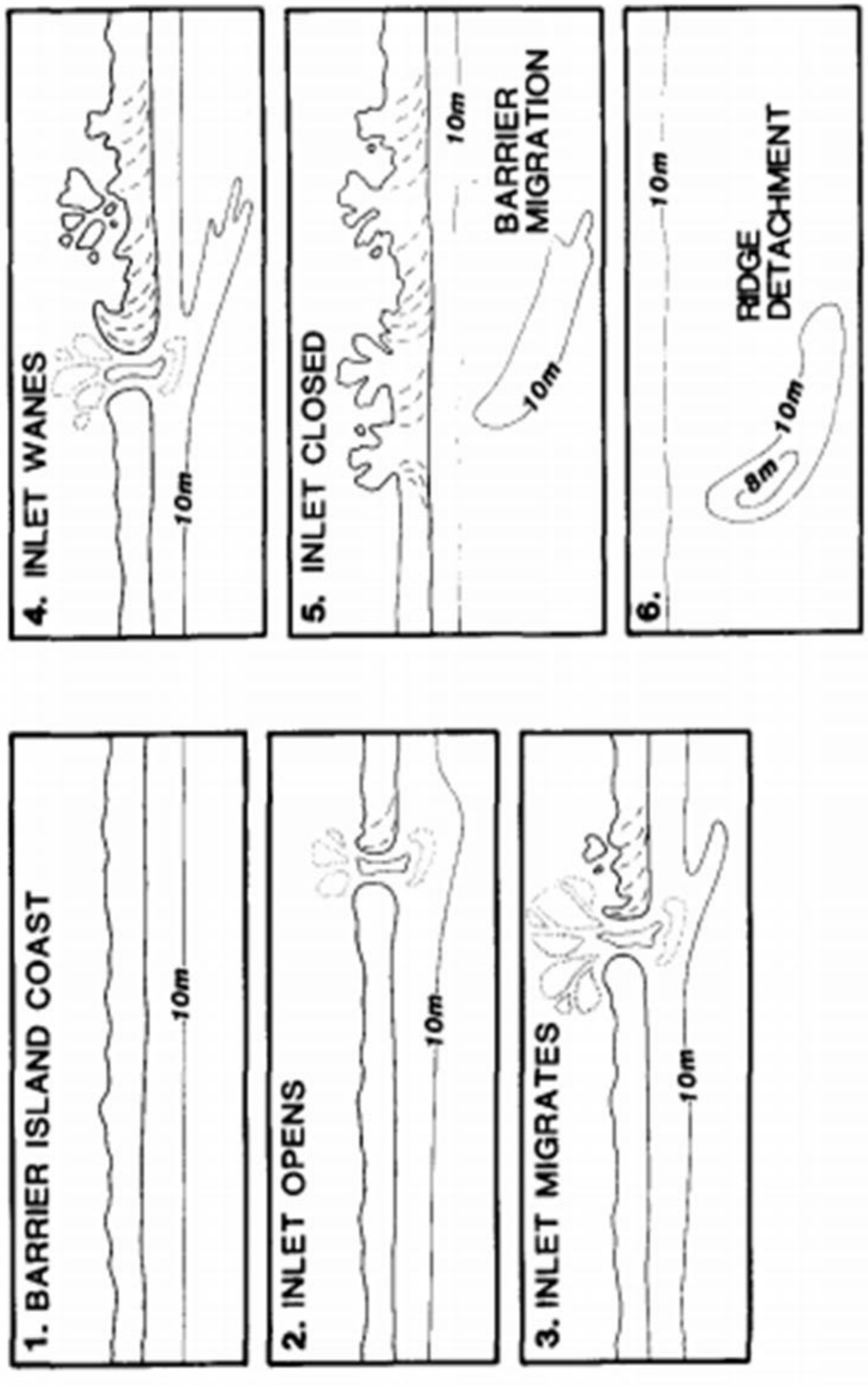


Figure 2-6

A schematic representing ridge detachment and shoal formation (McBride and Moslow, 1991).

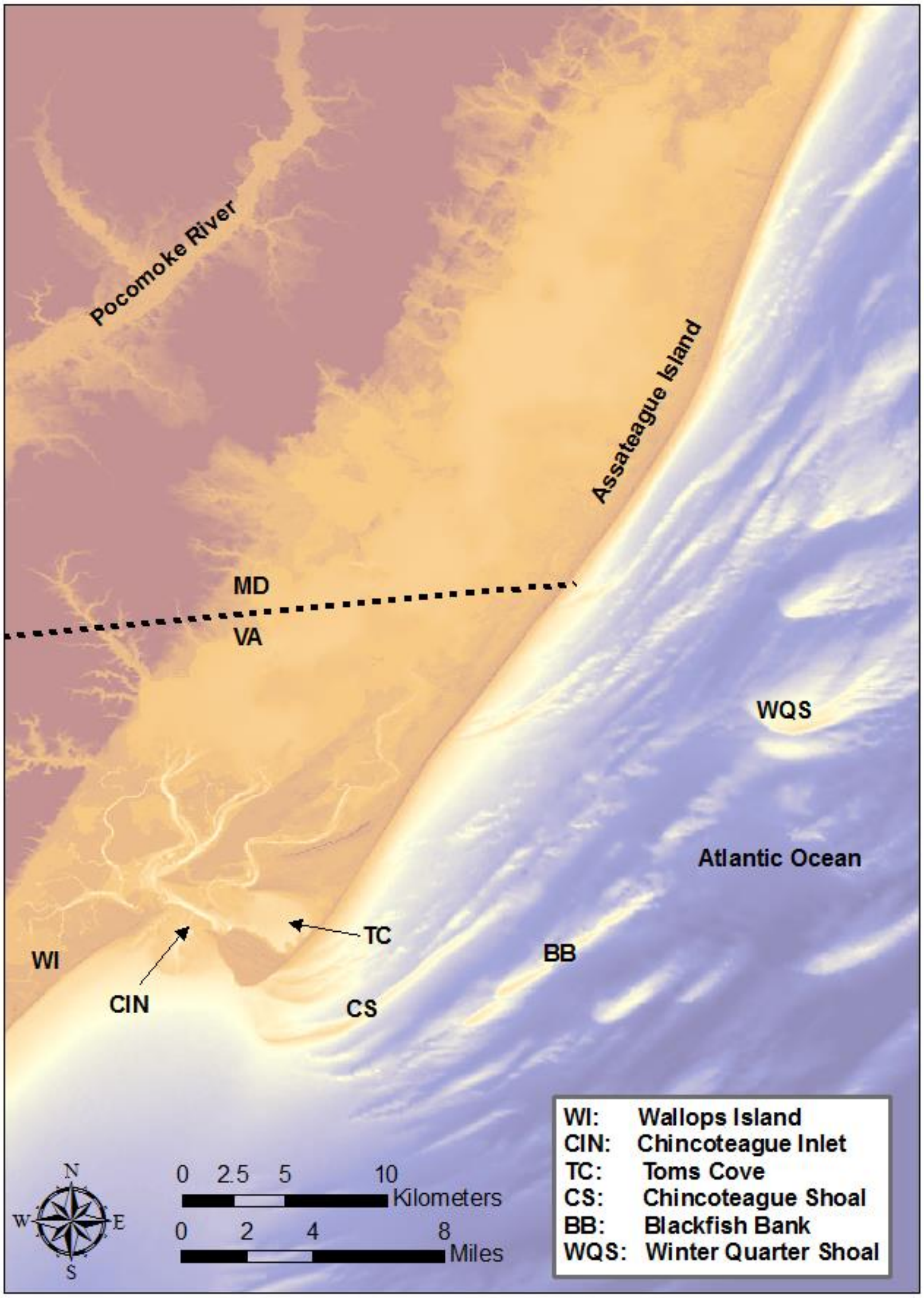


Figure 2-7

DEM of inner shelf bathymetry adjacent to Assateague Island, with major shoals and coastal features labeled.

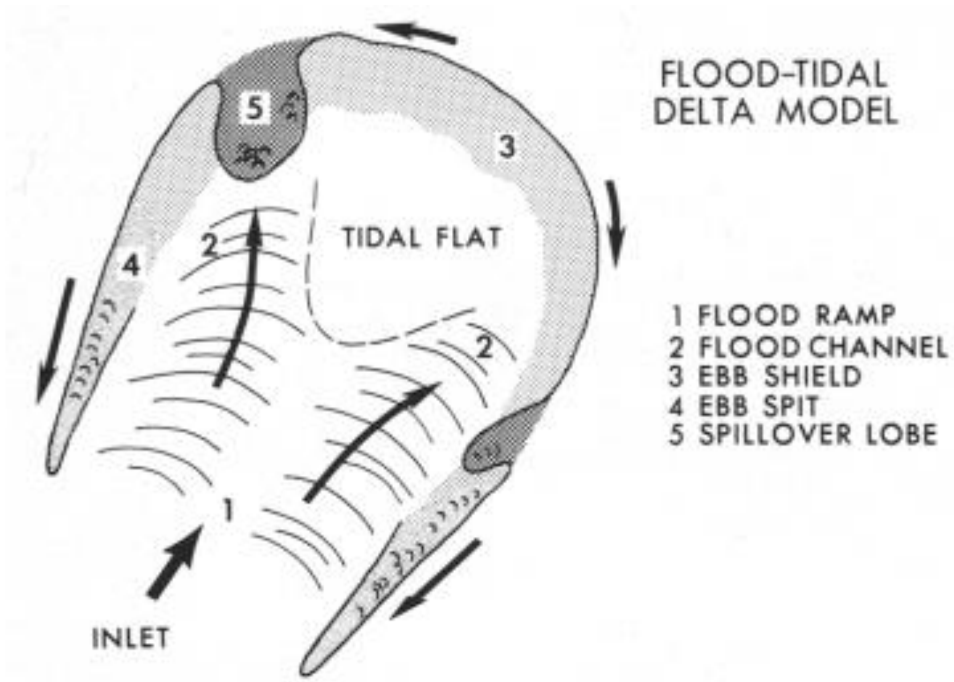


Figure 2-8

A schematic representation of the fundamental components of a flood-tidal delta complex (Hayes and Kana, 1976; Hayes, 1980).

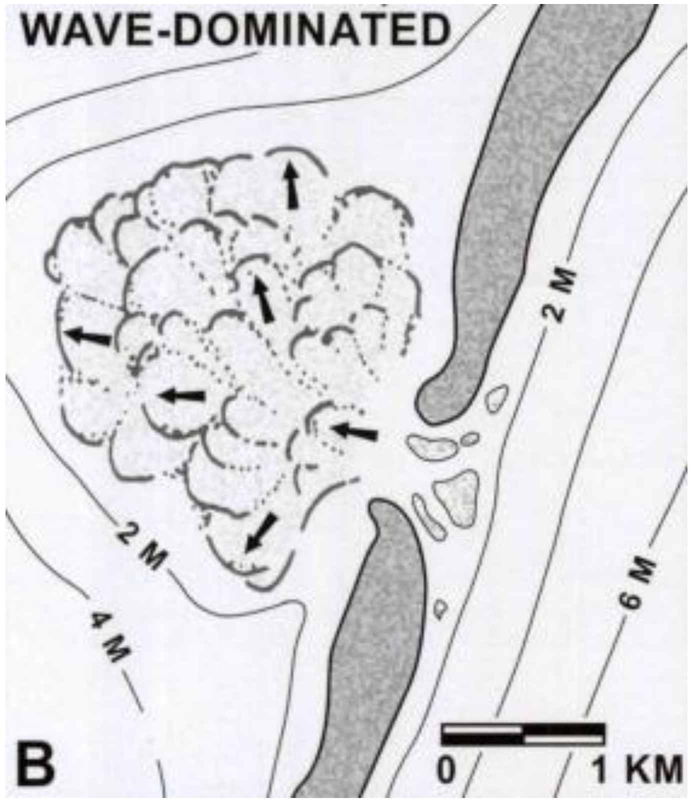


Figure 2-9

Schematic of a wave-dominated tidal inlet system with a flood-tidal delta complex in back-barrier (Hayes and FitzGerald, 2013).

Chapter 3

Methodology

Geophysical Mapping

Seismic data were collected in the Atlantic Ocean off the southern coast of Assateague Island (Figure 3-1). The Atlantic Ocean data were collected in June of 2004 and the Chincoteague Bay data were collected in May of 2005 (Krantz and Levin, 2005). The data were acquired using an EdgeTech® chirp seismic system (Figure 3-2), to survey over 150 km (Figure 3-1). The system has a swept-frequency of 500 Hz – 12 kHz and had a vertical resolution of roughly 20 cm (Wikel, 2008). Also, an Edgetech® side-scan sonar system (100 kHz, 300 m swath) was used to interpret seafloor features and enhance seismic interpretation (Wikel, 2008).

The seismic data were processed using EdgeTech® Discover processing software. The p-wave velocity for the data were 1500 m/s for penetration through saturated sediment, allowing for the interpreted depths of reflectors. Seismic reflections were interpreted by their internal configurations, which included variable, transparent, parallel, sub-parallel, wavy, gas-attenuated, channel-fill geometry, and chaotic characteristics (Wikel, 2008). The seismic data were used to interpret the sequence stratigraphic context to generate an isopach map of sediments of flood-tidal delta deposits.

The GPS and way-point data from the seismic survey were imported into Excel®. Also, the interpreted elevations for the top of stratigraphic units were placed into Excel®, then mapped using Surfer® 8 and ESRI® ArcGIS®.

Kriging was the interpolation technique used to compose maps from the interpreted elevation data. Kriging is a geostatistical tool that assumes the distance between sample points and then reflects that was a spatial correlation inferring points where no data are present (ArcGis, Online data). This method is similar to the method used by Childers (2013).

Maps

The maps for this study were created using ESRI® ArcGIS 10. The maps include:

1) A location map with the seismic track lines projected onto the inner-shelf bathymetry (Figure 3-1).
2) A bathymetric map created from data collected and compiled by the National Ocean Survey (NOS). The map has an approximate 10-m cell size (1/3 arc second). The vertical datum is for mean high water. The map was constructed in October 2009 to support NOAA’s Tsunami Inundation

Program (NOAA). Offshore shoals that have been identified previously by other researchers are clearly visible on the map (Figure 2-7).

3) A contour map of the Holocene/Pleistocene Boundary (4-12). This map was made by interpreting the depth at the base of the Holocene from the seismic data. The elevations depths were then imported into Surfer® 8 to construct the map.

4) A surface map of the Holocene/Pleistocene boundary (Figure 4-9). This map was made by interpreting the depth at the base of the Holocene from the seismic data. The elevations depths were then imported into Surfer® 8 to construct the map.

Geophysical Interpretation

In this thesis, the concepts and applications of sequence stratigraphy are used to address the late Quaternary depositional history of the Maryland and Virginia inner shelf similar to studies by Foyle and Oertel (1992) and Foyle (1994). Sequence stratigraphic methods were developed to interpret major eustatic sea-level events, with important applications in petroleum exploration. A series of landmark papers defined and developed the sequence stratigraphic methods that will be applied to this study (Vail et al., 1977; Mitchum, 1977; Van Wagoner, 1988; Galloway, 1989; Posamentier and Allen, 1999).

The methods of sequence stratigraphy rely on three related assumptions (Posamentier and Allen, 1999):

1) *Accomodation Space* — The open space available for sediment deposition. For an inner shelf setting, this is the vertical space between the sea floor and wave base.
2) *Sediment Supply* — The rate at which sediment is placed into a system. The amount of sediment is controlled by climate, the regional fluvial drainage, and substrate lithology.
3) *Physiography* — The presence or absence of a shelf/slope break, gradients of the shelf or slope, and subsidence or uplift.

Below are some important sequence stratigraphic terms for this study:

1) ***Sequence*** — A relatively conformable succession of genetically relatable strata that are bounded by unconformities (Vail et al., 1977).
2) ***Systems Tracts*** — Genetically linked depositional systems (Posamentier, 1988). More detailed descriptions are presented of separate systems tracts below.
3) ***Ravinement Surface*** — A time-transgressive subaqueous erosional surface produced during the landward movement of the transgressive systems tract (Posamentier and Allen, 1999) (Figure 3-3).

Three primary systems tracts are used to explain coastal evolution: Lowstand Systems Tract (LST), Transgressive Systems Tract (TST), and Highstand Systems Tract (HST).

The LST occurs during falling sea level, stillstand, and slow initial rise of sea level (Posamentier and Allen, 1999). The sediments present within this tract include deposition during the final phase of regression and the initial rise of sea-level. The early LST is formed during the fall of sea level that marks the onset of fluvial incision. The late LST is a time in which sea level will begin to rise slowly (Posamentier and Allen, 1999).

The TST, the tract seen in this study, consists of deposition that occurs from the onset of coastal transgression until maximum transgression (Posamentier and Allen, 1999). TST sedimentation occurs when relative sea-level rise increases. “The rate at which new accommodation space is added eventually exceeds the rate at which sediment is supplied, so that shoreline transgression is initiated” (Posamentier and Allen, 1999).

The HST consists of progradational deposits that form when the rates of accumulation exceed the accommodation space. At this time, the rate of relative sea-level rise slows dramatically (Posamentier and Allen, 1999). Toward the end of the HST, sea level is no longer rising, and may be falling, and only seaward progradation of the sediment occurs (Posamentier and Allen, 1999). A diagram of the seaward systems tracts is presented in Figure 3-4.

The concept of the LST, HST, and TST is applicable for both coastal dynamics and oil and gas exploration. Shallow stratigraphic systems can be used as modern analogs to develop better models for oil and gas exploration. For example, The LST could be a target source rock, and the TST can act as a source, seal, or reservoir. An example of the facies would be as follows: the source could be a marine shale or bay/lagoonal sediments, a condensed section of shale could act as a seal, and the transgressive lag and incised-valley fill could act as the reservoir (Posamentier and Allen, 1999). The HST has the potential for being either a source or seal. However, a reservoir in the HST would most likely be related to shelf deltas or bayhead deltas (Posamentier and Allen, 1999).

All of the systems tracts were outlined to give a perspective on the sequence stratigraphic interpretation of coastal zones and the tracts applications. However, for this study the TST encompasses the entirety of the geophysical data.

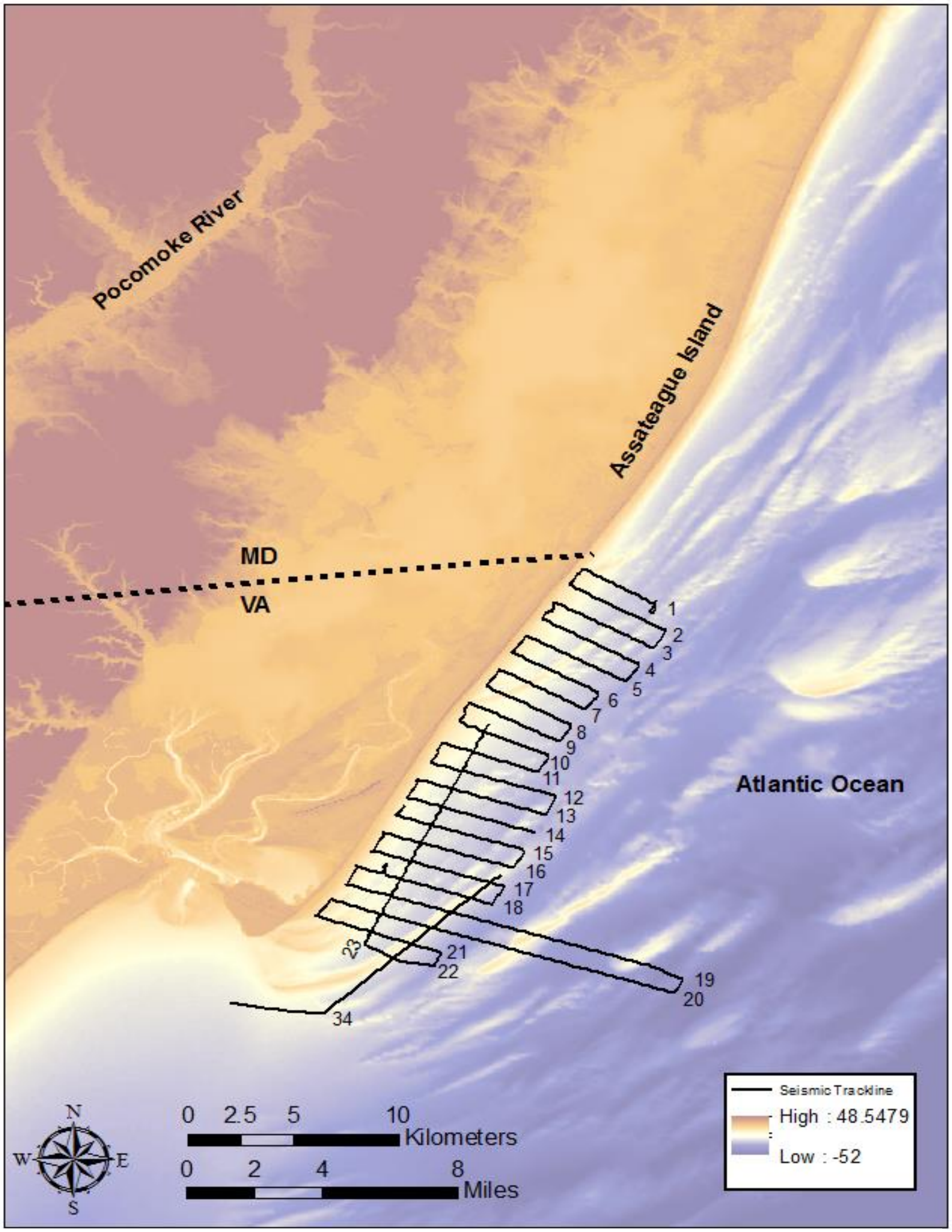


Figure 3-1

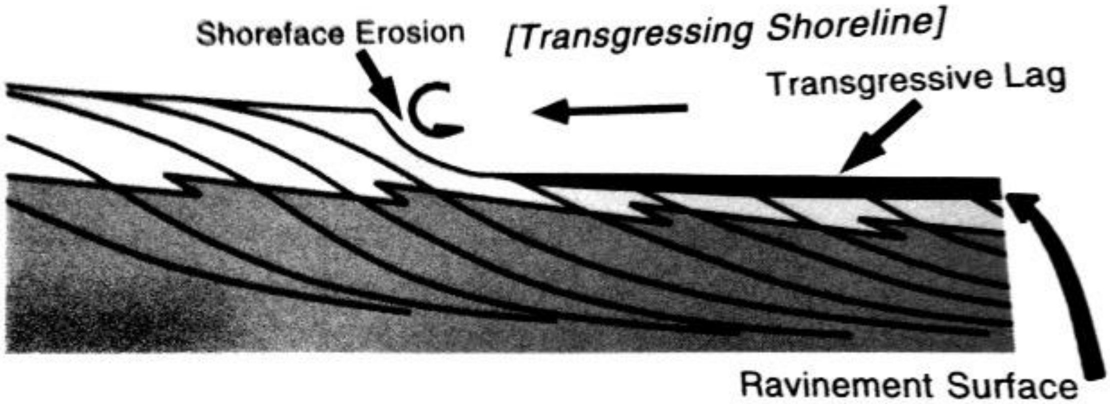
Seismic tracklines for this study and bathymetry of the shoreface off central and southern Assateague Island. Illumination of the bathymetric DEM from the northwest.

Figure 3-2 (Left)

The Edgetech® 512i towfish chirp system. Image from Norton (2008)

Figure 3-2 (Right)

The Edgetech® Discover acquisition and processing software system. Image from Norton (2008).



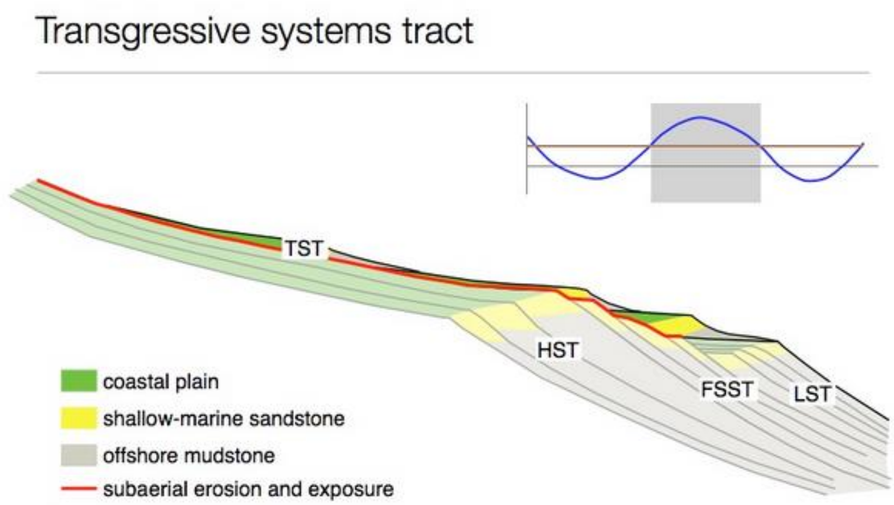


Figure 3-3 (Top)

Transgressive erosion of the shoreface and formation of a ravinement surface. Image from Posamentier and Allen (1999).

Figure 3-4 (Bottom)

Schematic representation of sequence stratigraphy systems tracts (University of Georgia, Stratigraphy Lab)

Chapter 4

Results, Discussion, and Conclusion

Facies and Interpretation of Primary Surfaces and Seismic Facies

For this study, approximately 160 km of seismic data were interpreted and analyzed, revealing a complex stratigraphy of the inner shelf. Figure 4-1 shows the three primary surfaces of the study area. Figures 4-2, 4-3, and 4-4 display representative examples of the stratigraphy as uninterpreted and interpreted seismic profiles. The seismic-stratigraphic units are characterized as units A, B, and C. Unit A contains transparent and parallel reflections bounded by the seafloor above and reflection R1 below. Reflection R2 is seen locally atop Unit B in the stratigraphy. Unit B consists of a vast array of reflections, complex internal reflections, and a variety of subunits bounded below by reflection R3. The reflections within Unit B are identified as parallel to sub-parallel reflections that can be traced laterally within the strata. The identifiers for the primary units and major surfaces are adapted from Toscano (1989), Foyle and Oertel (1994), and Wikel (2008) (Table 4-1).

In support of the seismic data, side-scan sonar data also was acquired during the same 2004 survey. The side scan sonar gives backscatter from the ocean floor that may be used to link the subsurface stratigraphy with the inner-shelf morphology. The backscatter for the top of Unit A appears as relatively uniform, low-amplitude return that is typical of areas with sand ridges and sand sheets. The backscatter from exposures of Unit B consisted of a low-hard return. The hard return would be represented by a lagoonal environment.

Wave-ravinement surfaces are represented as reflections R1 and R2. R1 is an active ravinement on the seafloor and is exposed at the seafloor in the troughs at the base of offshore shoals. The reflection is slightly concave and similar to the seafloor. This type of ravinement was observed by Wikel (2008) in the Assateague Island region and by Schwartz and Birkemeier (2004) on the North Carolina coast.

Ravinement surface R1 is a high-amplitude, continuous reflection in the uppermost parts of the stratigraphy. It is observed at the shoreface as an active transgressive ravinement and at the base of offshore shoals as a proto-ravinement at depths from 0 to 3 m below the seafloor (Wikel, 2008).

Ravinement surface R2 is a localized ravinement seen inconsistently through the seismic data. It is seen in the southern region of the study area (Figure 4-1). R1 and R2 often merge moving landward, armoring the shoreface against wave attack. R2 formed during an earlier transgressive phase of the shoreface prior to spit extension of southern Assateague Island (Wikel, 2008).

Reflection R3 is seen at greater depth within the seismic data. It represents the pre-transgressive surface, a sequence boundary that marks the Holocene/Pleistocene boundary. This surface is seen at the base of the lagoonal sediments of Unit B2. Due to being at the base of lagoonal sediments, methane gas can mask the R3 reflection. Investigators in the region have noted varying depths of R3, ranging from a few meters to over 10 meters depth (Morton and Donaldson, 1973; Finkelstein and Kearney, 1988; Oertel et al., 2008). The Holocene/Pleistocene boundary in this study is seen at an average depth of 15-20 m. However, at the southern portion of the study area depths greater than 30 m are observed. A shallow gradient occurs as the reflector moves seaward, the eastern sections of lines 19 and 20 display this characteristic.

Unit B is interpreted as Holocene, and consists of varying depositional environments associated with the back barrier such as: flood-tidal deltas, tidal creeks, filled tidal channels, and lagoonal sediments.

The lower section of Unit B consists of variable, continuous to discontinuous , parallel to subparallel, and gas-masking reflections, interpreted as lagoonal sediments (Unit B1). These facies have been well preserved below the inner shelf. However, the sediments of the upper section of Unit B1 have been re-worked due to wave attack and inlet processes. The organic materials within the lagoonal sediments produce a large amount of methane gas. The gas affects the resolution of the lagoonal sediments and surrounding sediments in the seismic profiles. On some profiles, gas can also be seen as bubbles within the water column.

Unit B2 consists of transparent to chaotic and variable-fill reflections, which is interpreted as inlet channel and tidal-creek fill. Channels overlie and cut into the lagoonal sediments (Unit B2). These types of channels are seen sparsely beneath the Maryland inner shelf but are seen frequently through this study area. Tidal channels are observed as part of a large flood-tidal delta complex (Figure 4-2) and are seen further south in the study area (Figure 4-3).

Landward-dipping sigmoidal, oblique, and parallel clinoforms signifying flood-tidal delta deposits lie above the tidal-channel sediments of Unit B2 (Figures 4-4). A massive flood-tidal delta complex marks a past back-barrier environment when Assateague Island was positioned approximately 5-8 km seaward of its current position. The prominent main ramp of the flood-tidal delta complex (Unit B3) can be seen in seismic line 8 (Figure 4-4). The flood-tidal delta deposits are distinctly cut by tidal creek facies representing the presence of the ebb flow on the complex. The creeks are seen as narrow V-shaped reflections cross-cutting the strata below (Figures 4-5). An isopach of the flood-tidal delta complex demonstrates the structure and topographic relief of the complex (Figure 4-6).

The tidal channels of Unit B2 are seen through the central and southern seismic lines (Figures 4-2 and 4-3). Minor tidal channels are seen within a flood-tidal delta complex in seismic lines 3-5, 7, 9, 10, 12, and 13. (Figures 4-2, 4-4, and 4-7). A major channel oriented from northwest to southeast appears in lines 13 to 20 (Figures 4-3 and 4-8). The channel from lines 13-18 maintains a constant NW-SE orientation, then turns sharply east in lines 19 and 20 (Figure 4-8). Oertel et al. (2008) argued that a channel in approximately this position turned south in orientation as part of a Holocene drainage system (Figure 4-10). However, this is not seen in the seismic data.

The sand sheets and shoals, Unit A, interpreted as Holocene, are the uppermost facies seen within the geophysical data. The sand sheets and shoals are discontinuous throughout the study area, consistent with their time-transgressive character. Unit A was extensively investigated by Wikel (2008) to evaluate and correlate the alongshore vulnerability of Assateague Island to erosion.

Results

The Holocene-Pleistocene boundary is observed as a shallow seaward-dipping surface (Figure 4-9). This can be seen as a contour map in Figure 4-10. It has a relatively consistent elevation seaward, making a fairly distinct boundary. However, the southern portion of the surface, near seismic track lines 17-19, indicates channel-type geometry. The channel likely served as part of the Chincoteague Watershed, an early Holocene drainage system, consistent with the interpretation of Oertel et al. (2008) (Figure 4-11). The drainage likely was directed to an abandoned portion of the early to middle Pleistocene Susquehanna River valley (Oertel et al., 2008). It likely formed during a glacial lowstand in a fluvial environment strong enough to scour and erode the previously deposited estuary muds. At the onset of the Holocene transgression, the ancestral drainage network began flooding, and the barrier-island system began to move landward. Assateague Island was estimated to be as far as 60 km seaward of its present position and anchored by the next interfluve to the east. This distance would allow for the intermediate-sized Chincoteague Watershed to be located behind the ancestral position of Assateague Island. The Holocene-Pleistocene boundary displays a steepened elevation closer to the modern shore. This could represent a shoreline that was created during a previous sea-level cycle. Regional evidence now suggests that one or more highstands during MIS3, the Wisconsinan Interstadial, were at or near present sea level (Ramsey, 2010)l

A massive flood-tidal delta complex appears in the subsurface seismic data off the southern coast of Assateague Island (Figure 4-8). A large flood-tidal delta ramp marks the core of the complex, the area where most sediment deposition would occur (Figure 4-4). The ramp is seen primarily in seismic line 8. As the flood-tidal delta extends into the back-barrier environment, the deltaic deposits thin out dramatically. An example of thinning facies is presented in line 9 (Figure 4-7). Smaller tidal-creek channels, about 20 to 40 m wide and 2 m to 6 m deep, cut into the flood-tidal delta deposits, an indicator of ebb-flow. Also, the flood-tidal delta deposits are cut by larger tidal-inlet channels. The size of the complex, including the flood-tidal delta and tidal-inlet facies is approximately 4.5 km by 3.5 km. This is similar in size to modern tidal inlet complexes in wave-dominated settings along the modern Atlantic coast, such as at Indian River Inlet in Delaware.

The tops of the flood-tidal delta deposits and the banks of the tidal creeks are stratigraphic indicators of past sea-level heights. These environments represent past sea-levels within approximately 1 m, accounting for tidal ranges in a wave-dominated setting. The Delaware coastal region has been intensely studied using radiocarbon dates from peat and basal peat samples to understand past sea-levels through the Holocene transgression (Ramsey and Baxter, 1996; Ramsey, 2010). The data from Ramsey and Baxter (1996) consists of the dates closest to the study area. Figure 4-12 displays a sea-level curve from Ramsey and Baxter (1996). Throughout the study area, the top of the flood-tidal delta deposits are consistently 9-11 m below sea-level. Figure 4-5 displays seismic lines 5, 6, and 9. These lines consist of flood-tidal delta and tidal creek deposits. The relative depth below sea-level can be observed. The sea-level curve from Ramsey and Baxter (1996) shows a leveling off of sea level around 10 m below present sea-level from 5-7 ka. There are no actual dates from Assateague Island of when it was presently in situ. However, barrier island lagoons further south have been dated to about 5 ka. Therefore, the flood-tidal delta complex could have formed at about 5 ka or earlier.

Conclusions

Geophysical data were collected and used to interpret Holocene coastal depositional environments off the coast of Assateague Island in the Atlantic inner shelf. Seismic data were used to identify different facies and reflectors. These facies were used to chronologically understand the Holocene transgression. Facies such as marine shoals, inlet deltas, inlet channels, lagoonal environments were present within the data. The major reflectors in the seismic data were interpreted as wave ravinements (R1 and R2) and the Holocene/Pleistocene Boundary (R3).

A flood-tidal delta complex existed 5-7 Ka, off the southern portion of Assateague Island during a former seaward position. This conclusion is supported by geophysical evidence linked to a regional sea-level curve based on radiocarbon dates. The only previous author to acknowledge this complex is Wikel (2008).

The location of the Holocene/Pleistocene boundary demonstrates the location of the pre-transgressive surface and insight into previous subsurface bathymetry. This has a channel-like geometry scoured into it. This channel may have been part of a Holocene drainage system consistent with Oertel et al. (2008).

The knowledge and understanding of these complex coastal systems is pivotal for understanding past, present, and future climate change. This specific system can be important to the regional Holocene geology, as similar systems have been seen and interpreted in Virginia and Delaware. The continued research offshore of Assateague will help piece together the interpretation of the Holocene transgression.

Table 4-1

Primary seismic-stratigraphic units and surfaces. Modified from Wikel (2008).

Unit	Subunit or Facies	Reflector	Internal Reflection Configuration	Sidescan Sonar Signature	Depositional Environment	Toscano et al. (1989) Units	Foyle (1994) Units	Wikel (2008) Units
A	A2		transparent, discontinuous parallel and subparallel	uniform, high backscatter	marine sand sheet and shoals	Q5	A	A2
		----R1----	Transgressive wave ravinement					
	A1		transparent, discontinuous parallel and subparallel	mottled, medium backscatter	marine shoals and spit platform			A1
		----R2----	Local ravinement (R1/R2 merge as they approach shore)					
B	B4		variable, mostly transparent, short discontinous parallel and subparallel		backbarrier (e.g., overwash, tidal flat)			B5
	B3		sigmoid, oblique, parallel clinoforms	mottled, hummocky low to medium backscatter	inlet delta and spit	Q4		B4
	B2		transparent to chaotic, variable channel fill geometry		inlet channel and tidal creek fill		B	B3
	B1		variable, continuous to discontinous, parallel and subparallel, wave, gas masking		lagoonal/estuarine			B2
		----R3----	Holocene/Pleistocene Unconformity					
C	*undifferentiated*	**----R4----**	long, continuous and discontinuous, parallel to wavy	*no surface expression*		Q2	C-G	C
				no surface expression		Q1		
		----R5----	Pleistocene/Tertiary Unconformity					
	undifferentiated		N/A	*no surface expression*		T1	T	D

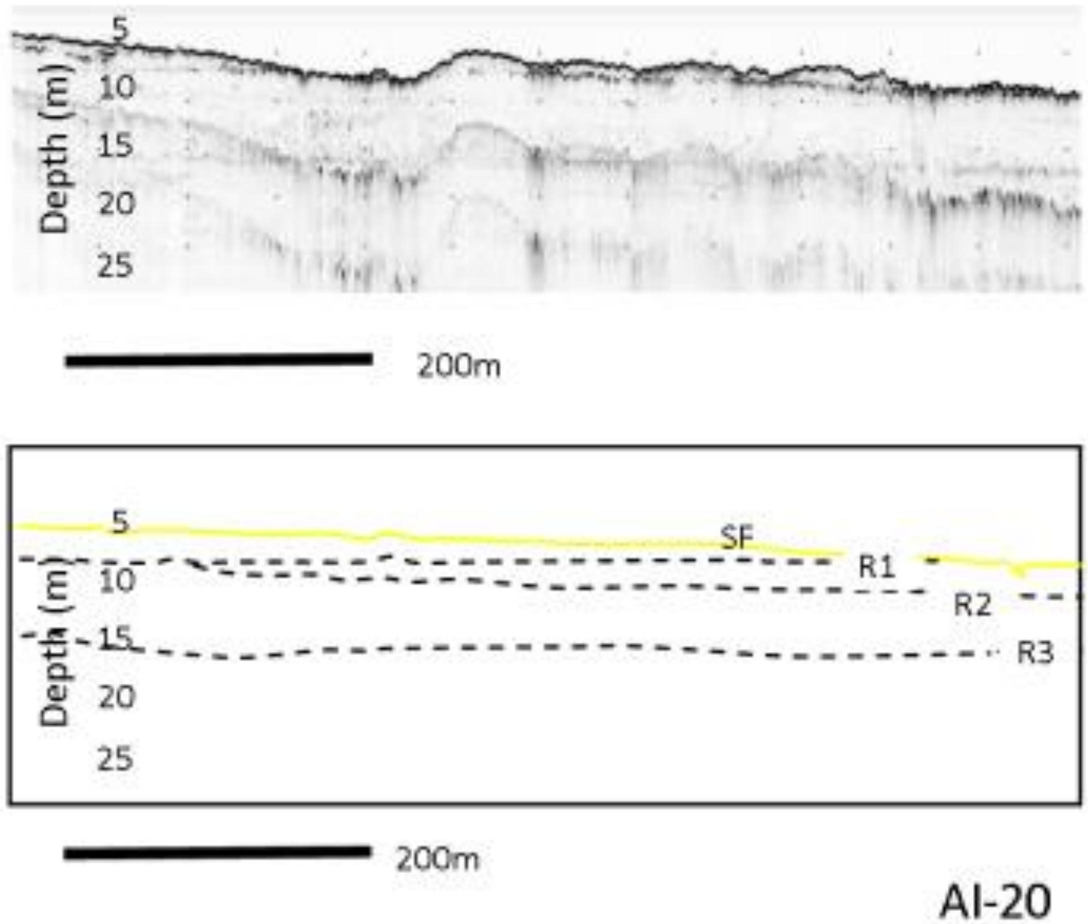


Figure 4-1

The seismic section (short portion of line 20) displays the primary reflection surfaces. R1 and R2 are seen armoring the shoreface.

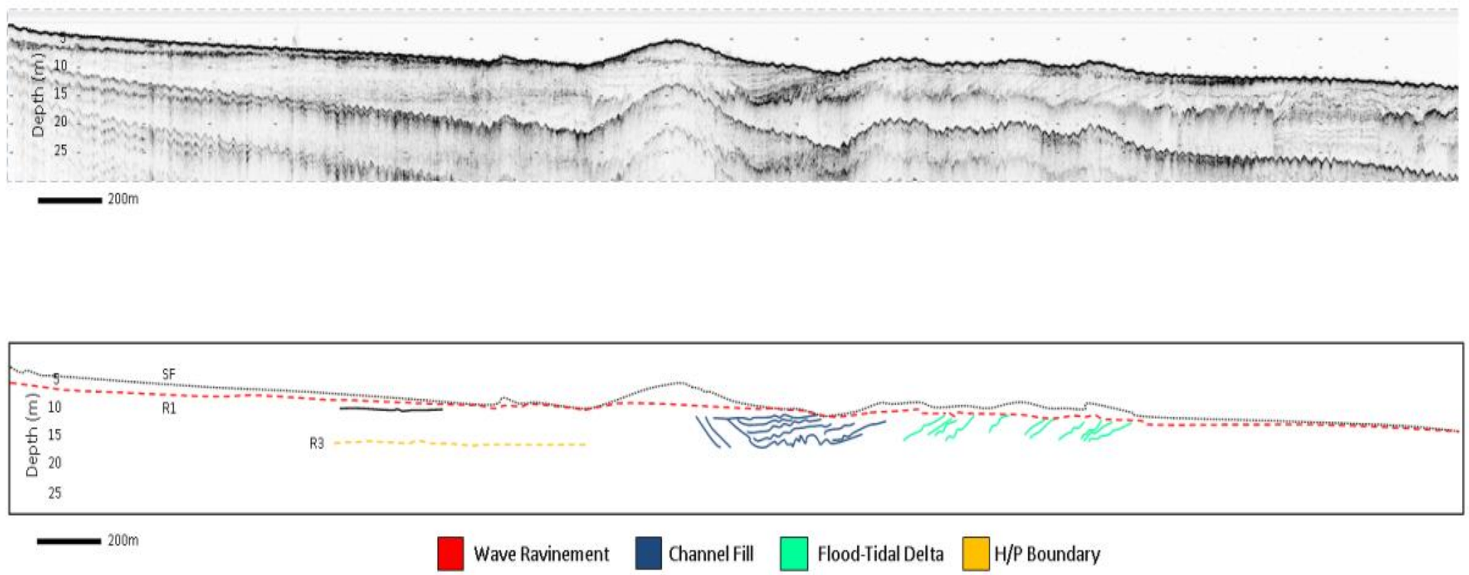


Figure 4-2

Seismic line AI-05 demonstrating tidal-inlet and flood-tidal delta facies.

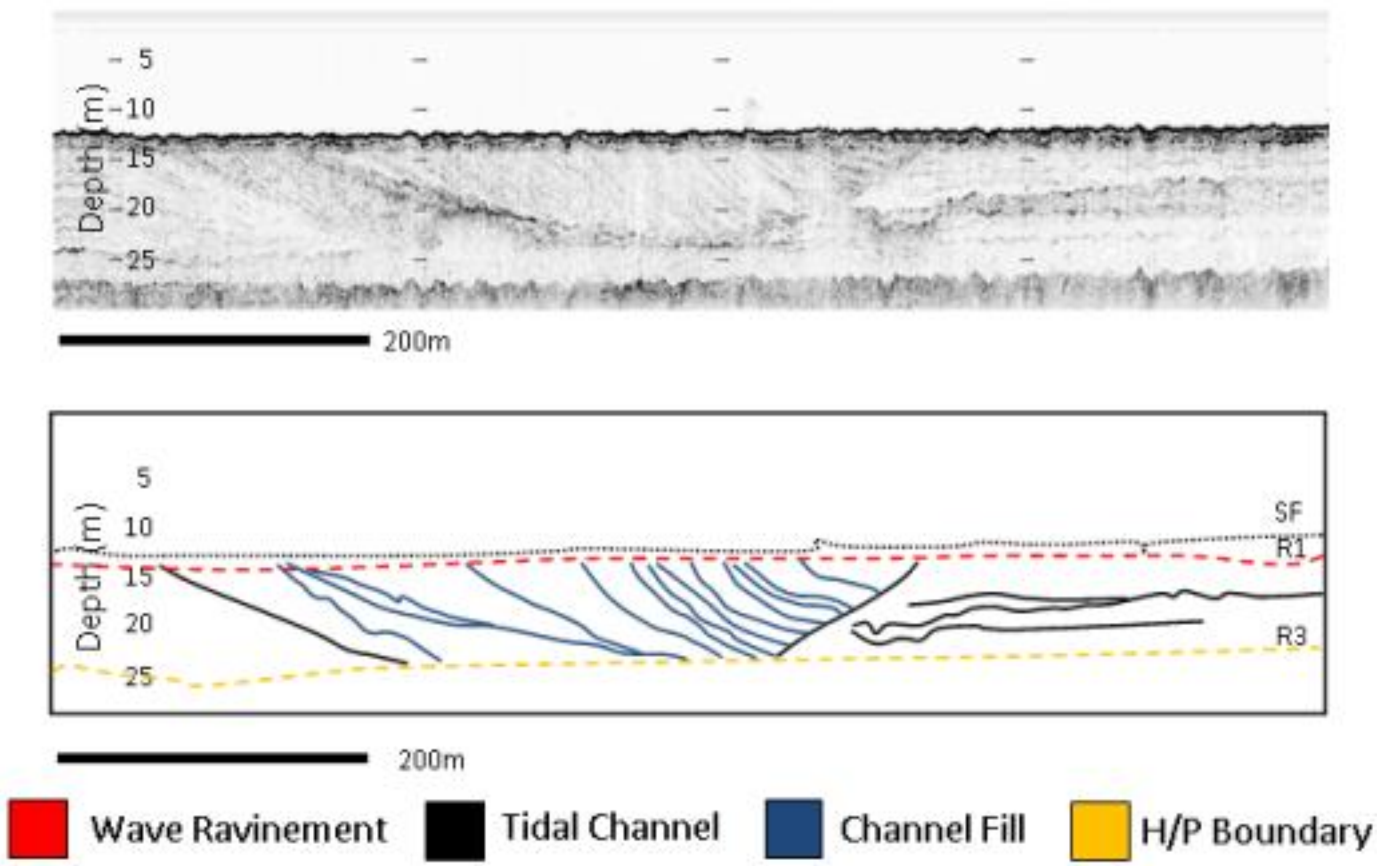


Figure 4-3

Seismic Line AI-17 representing the paleochannel geometry observed in the central to southern regions of the study area.

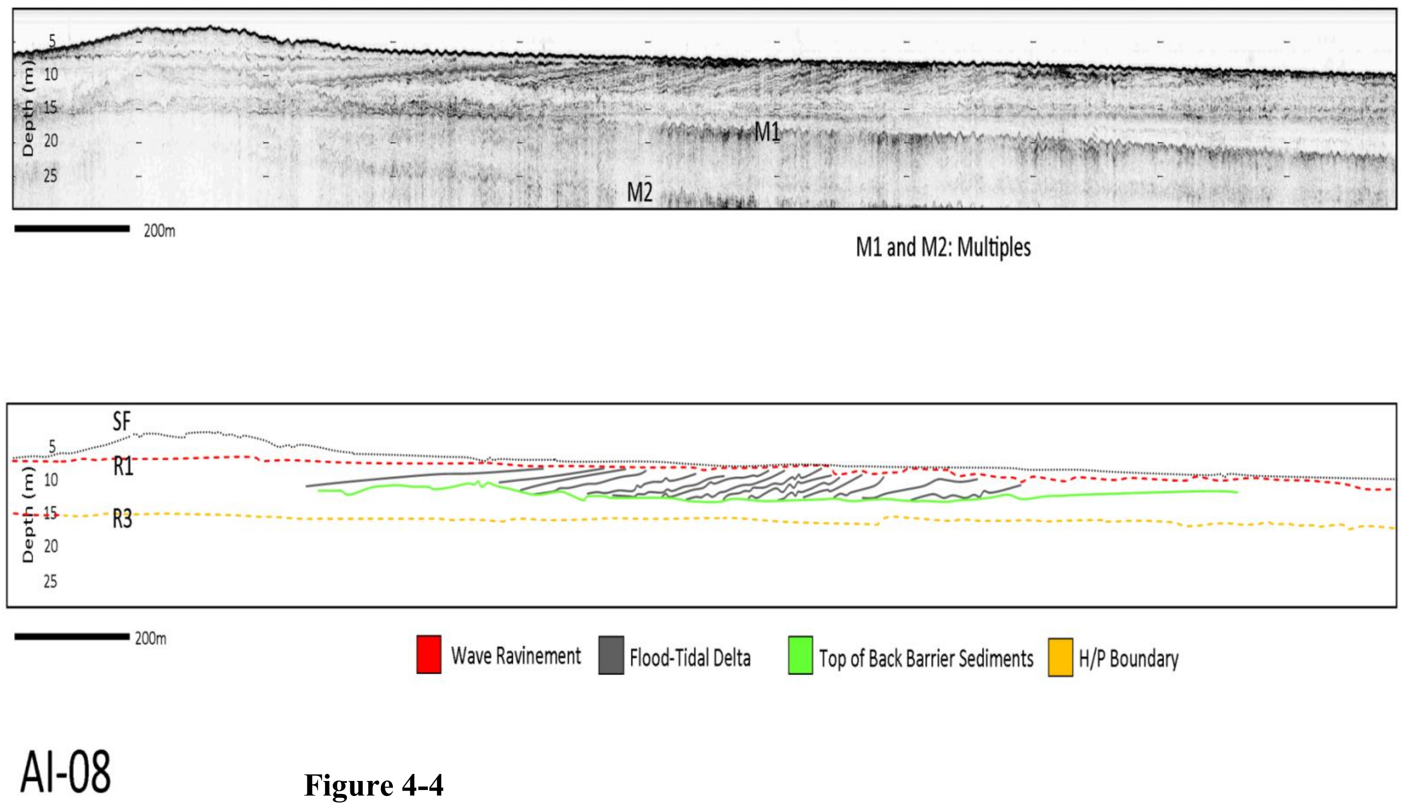


Figure 4-4

Seismic line AI-08 indicating the flood-tidal delta ramp facies.

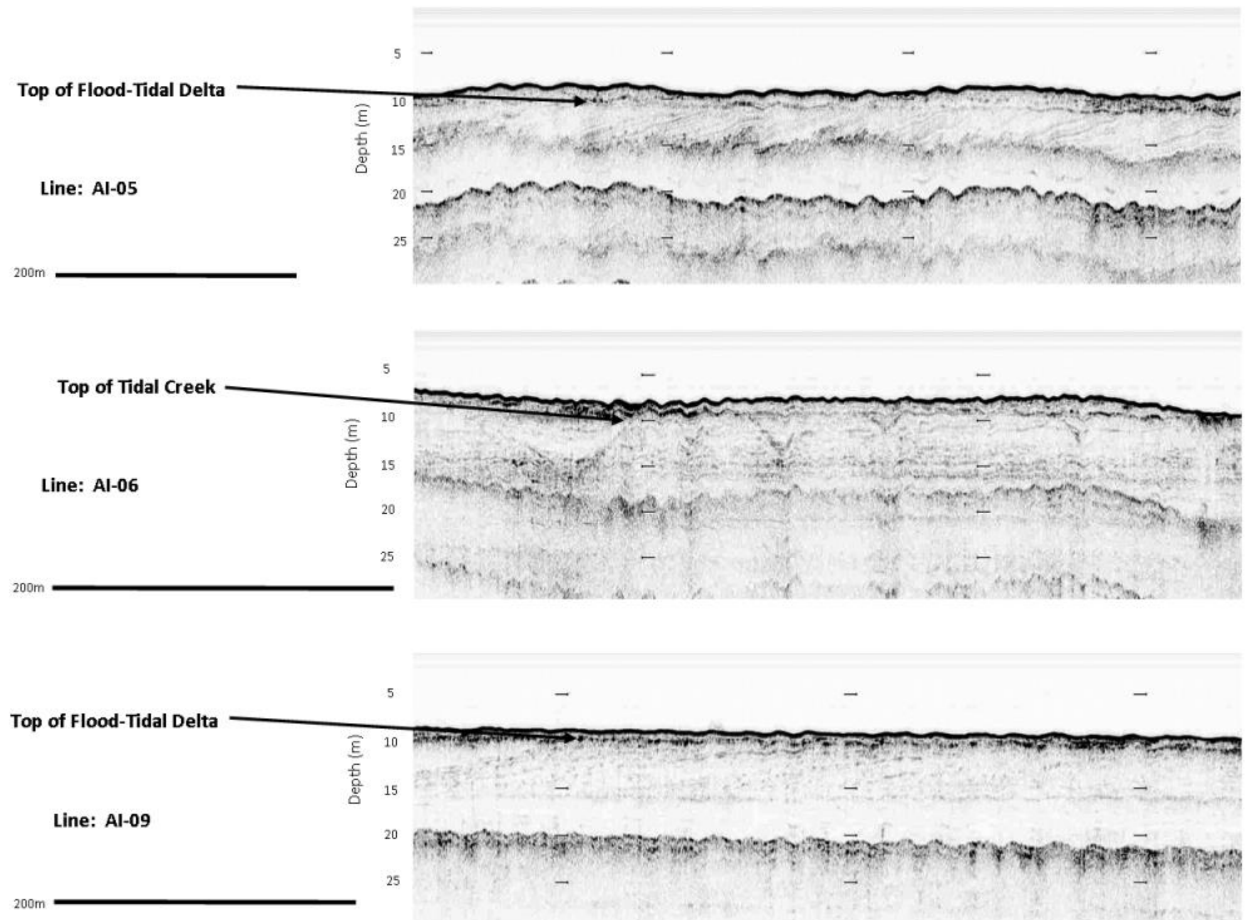


Figure 4-5

Seismic Lines 5, 6 and 9 showing flood-tidal delta and tidal creek depths below present sea-level.

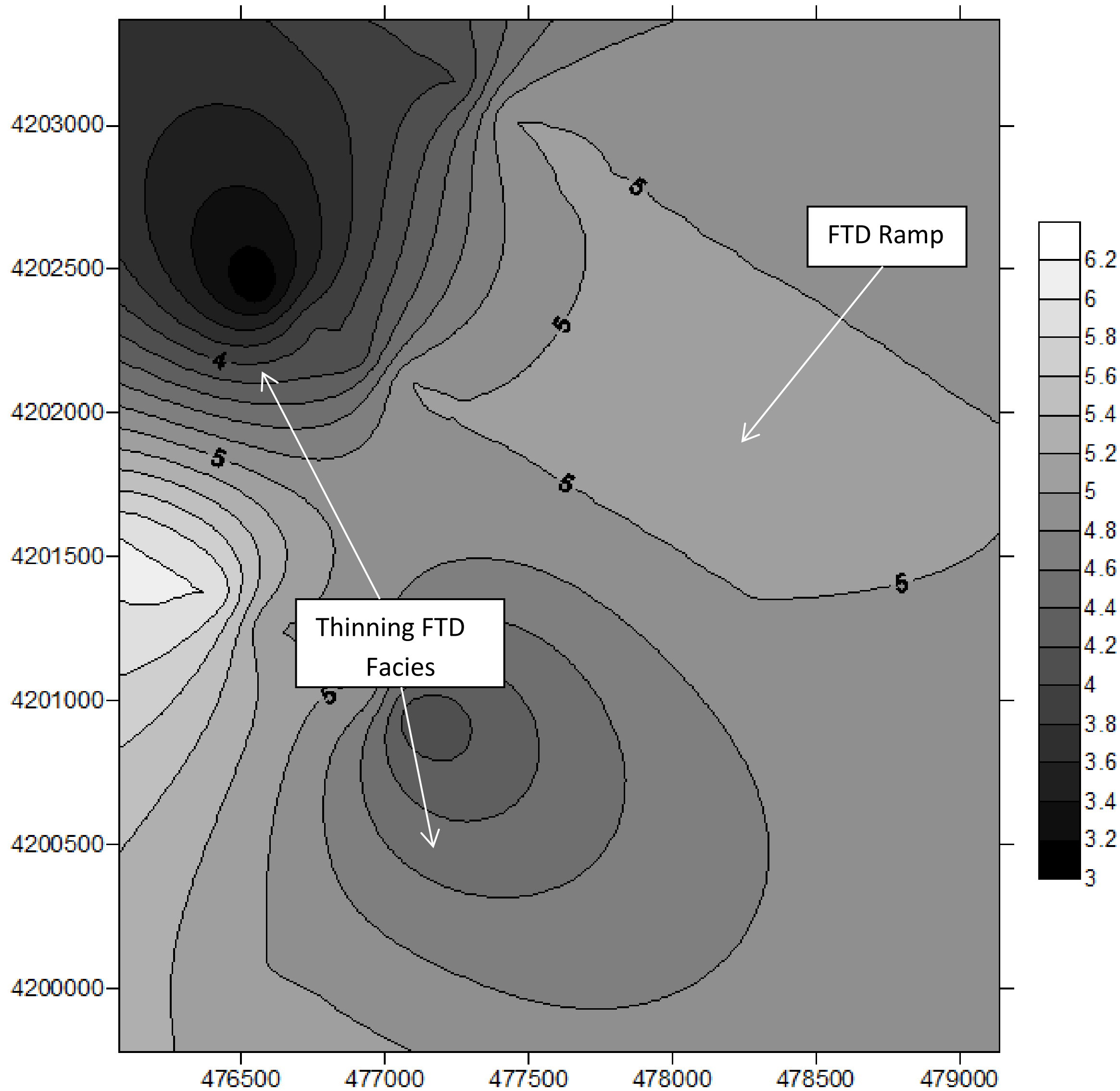


Figure 4-6

Isopach Map of flood-tidal delta complex. Displays thinning facies away from main flood-tidal delta ramp.

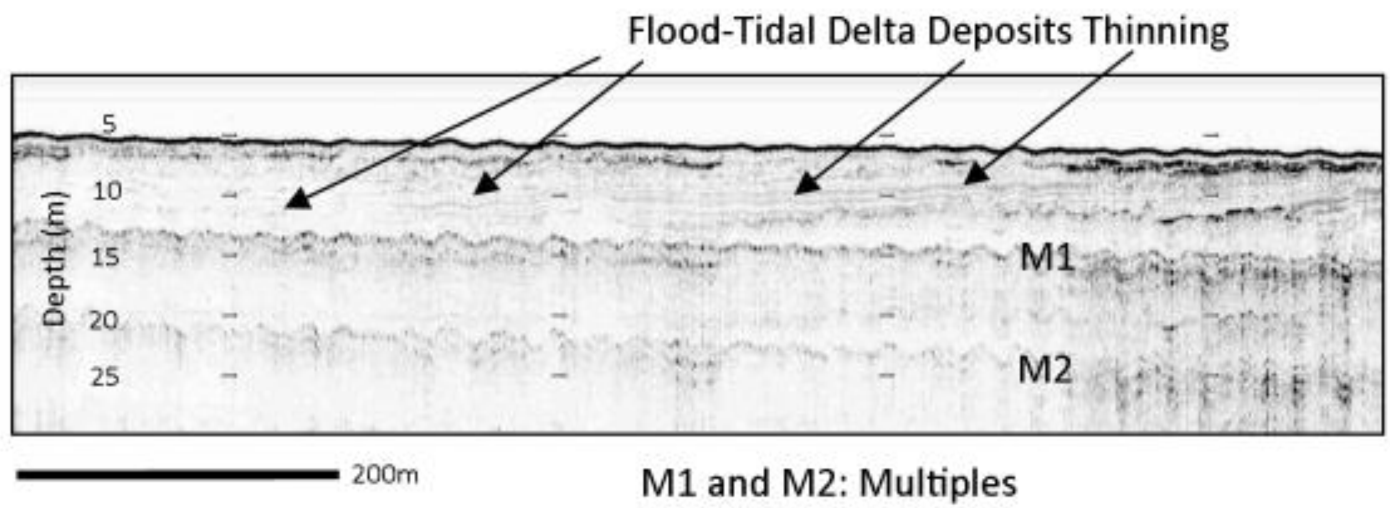


Figure 4-7
Thinning of flood-tidal delta deposits in seismic line 9.

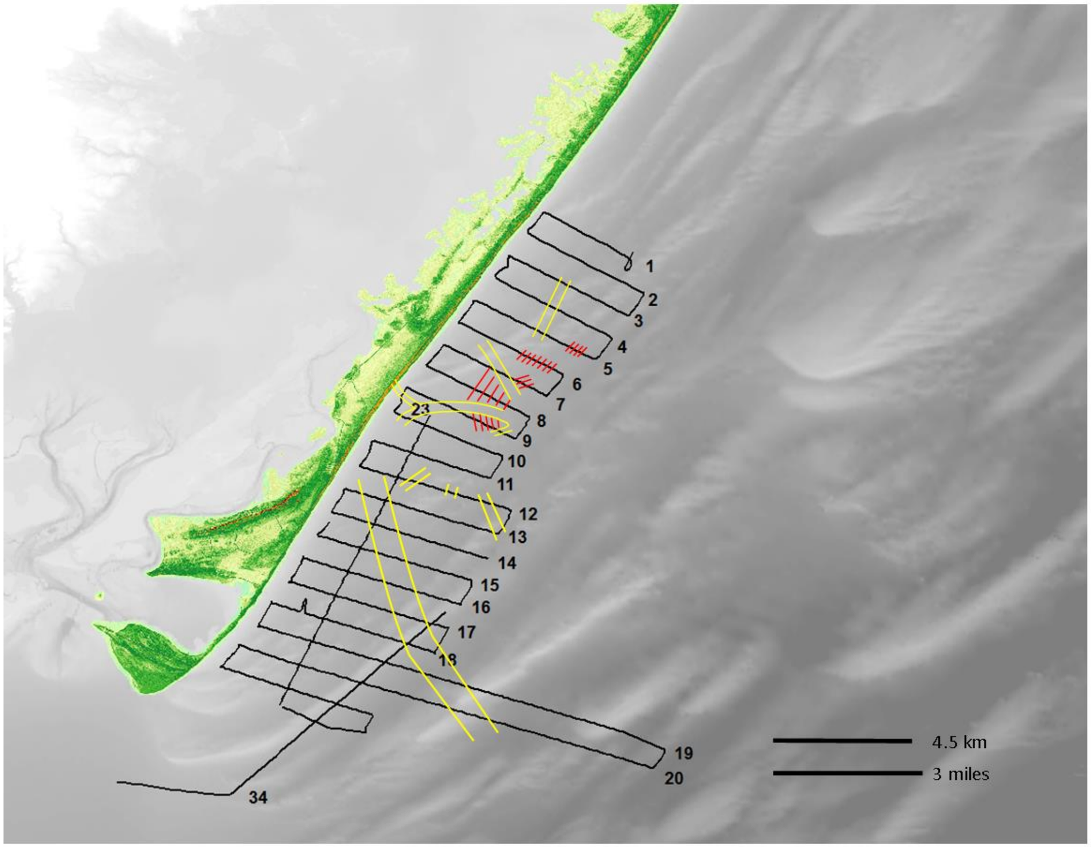


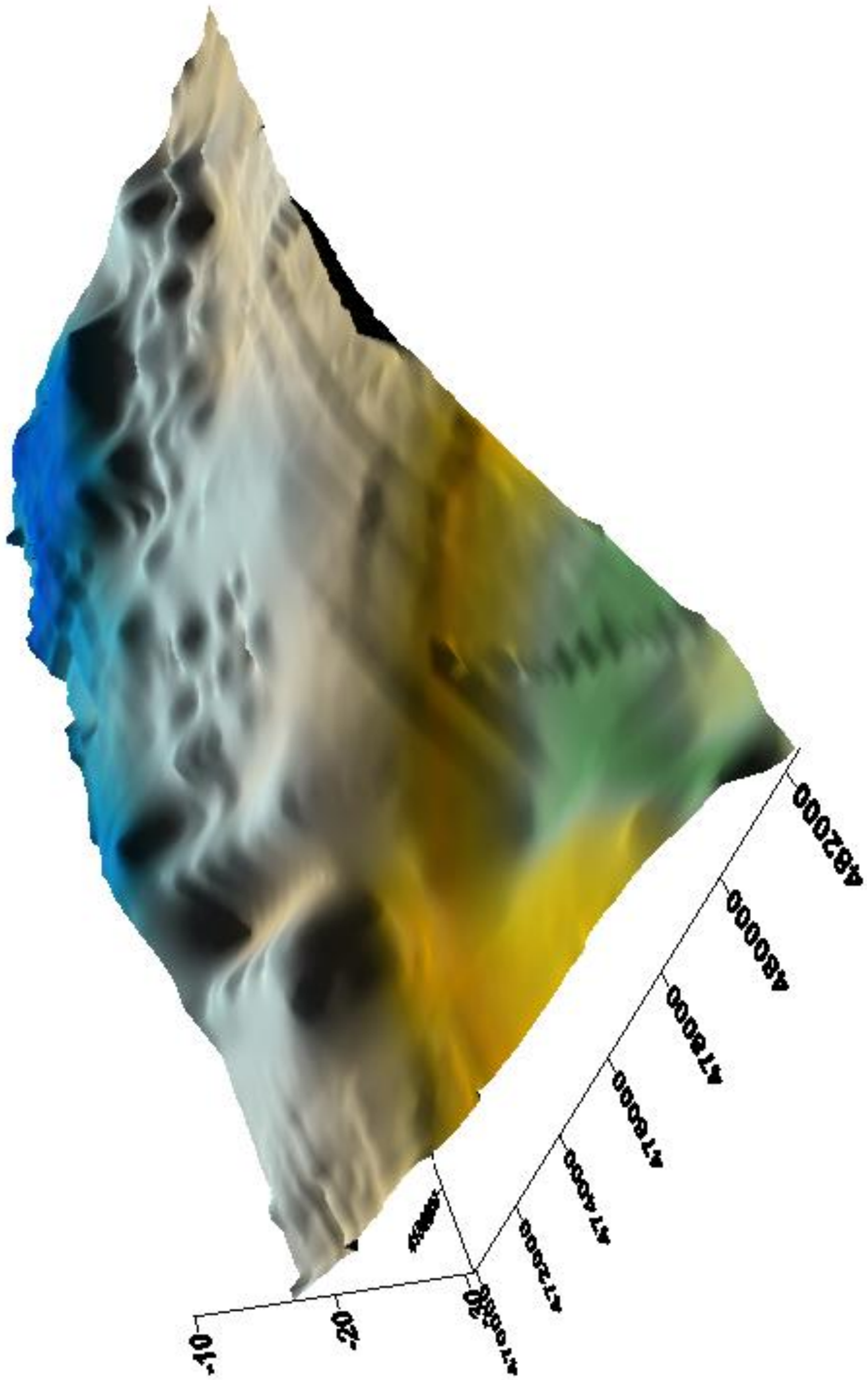
Figure 4-8

Flood-tidal delta complex and large paleochannel seen offshore of present Assateague Island shoreline
Yellow lines represent tidal-inlet channels and red lines represent flood-tidal delta facies.

Figure 4-9 A and B (on next 2 pages)

Surface map of the Holocene/Pleistocene boundary. Composed using interpreted depths of the base of the Holocene sequence from seismic data. Two images presented to see the surface from different perspectives.The general view on each is from the southeast.

Note: Scale in meters on y-axis; horizontal coordinates are UTM meters.


482000
480000
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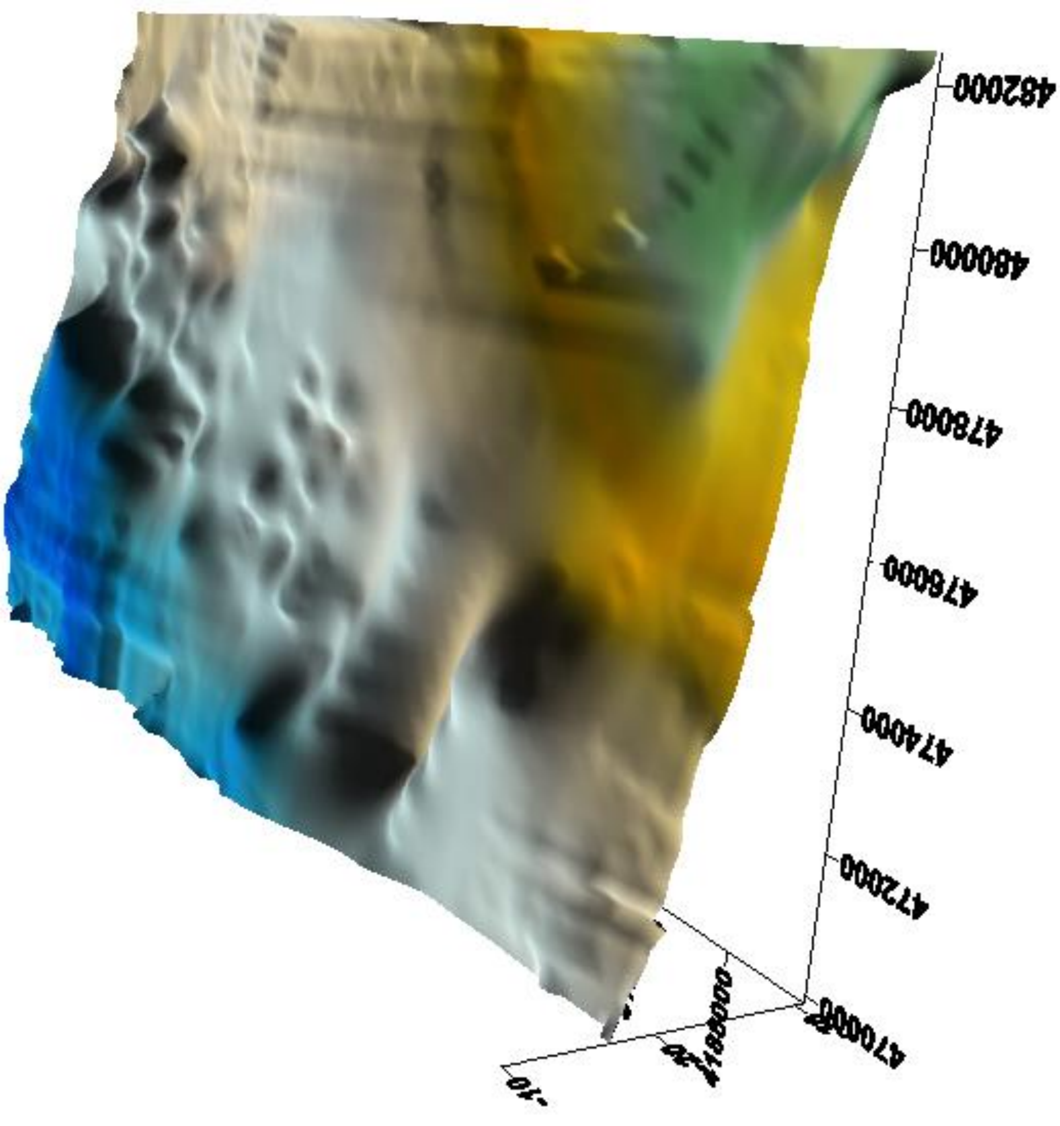
476000
474000
472000
-10
-20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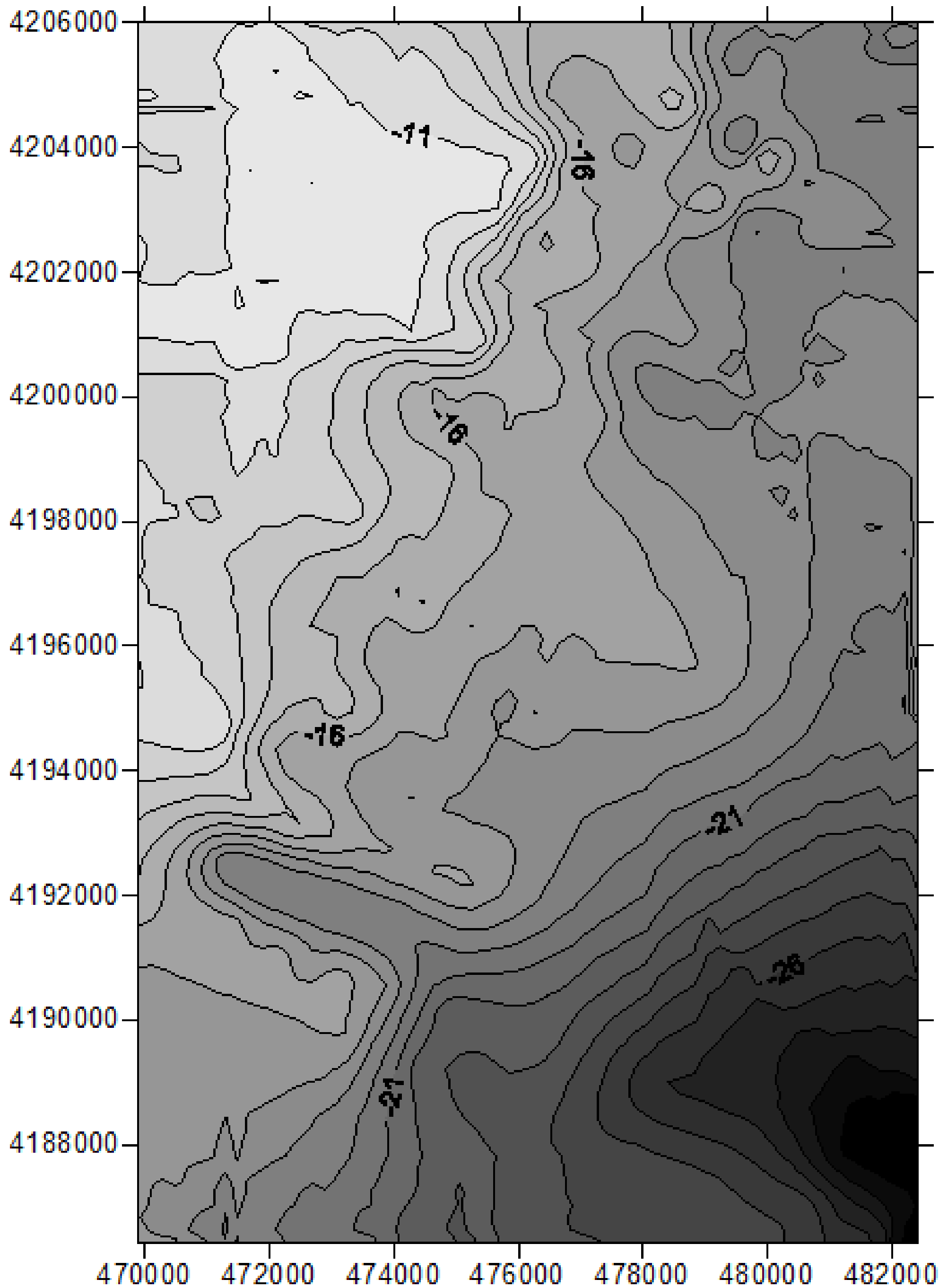

Figure 4-10

Contour map of the Holocene/Pleistocene boundary. Composed using interpreted depths of the Holocene from seismic data. Note: 1-meter contour interval.

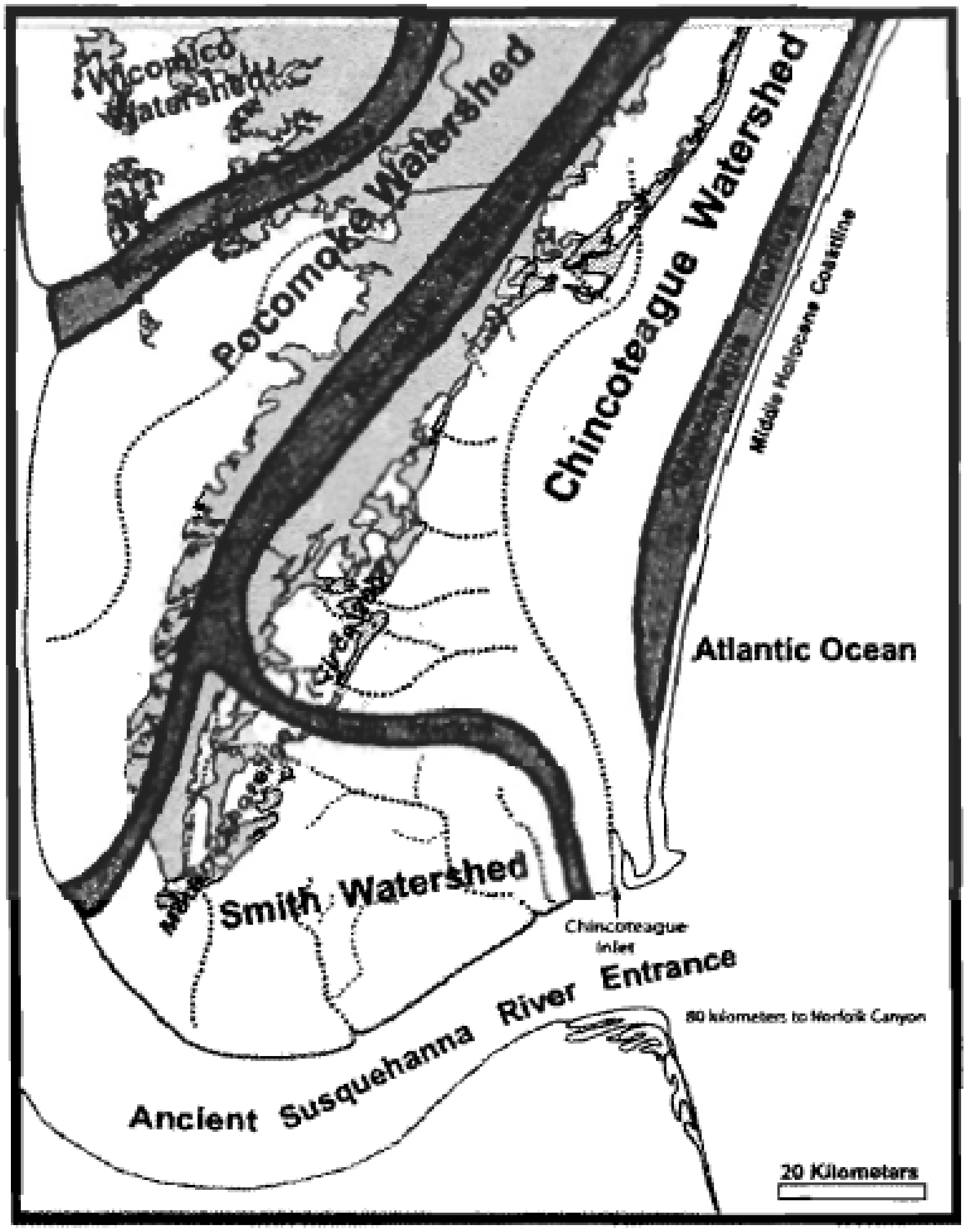

Figure 4-11

Tidal channel and Holocene drainage system (Oertel et al., 2008). Note: the study area is outlined in red.

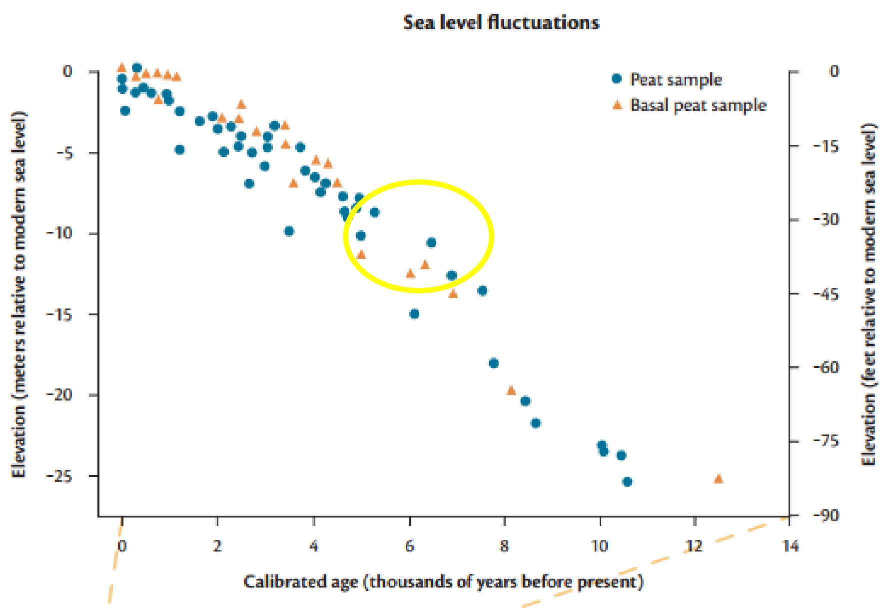

Figure 4-12

Regional sea-level fluctuations through the Holocene; data from Ramsey and Baxter (1996). The yellow circle indicates the elevation of uppermost portions of flood-tidal delta facies and banks of tidal creeks seen in the seismic profiles.

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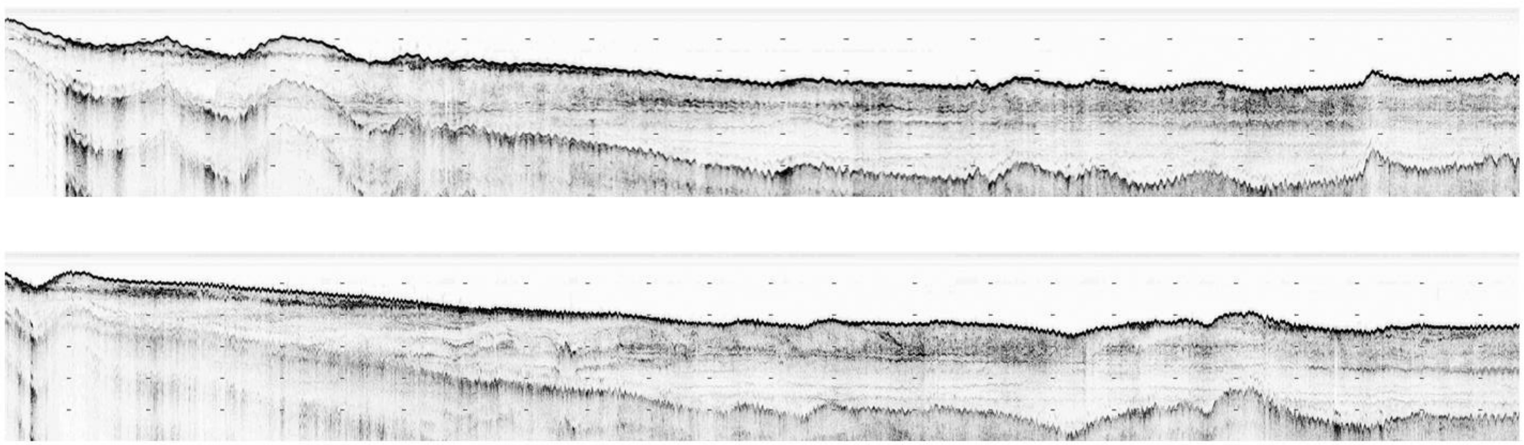
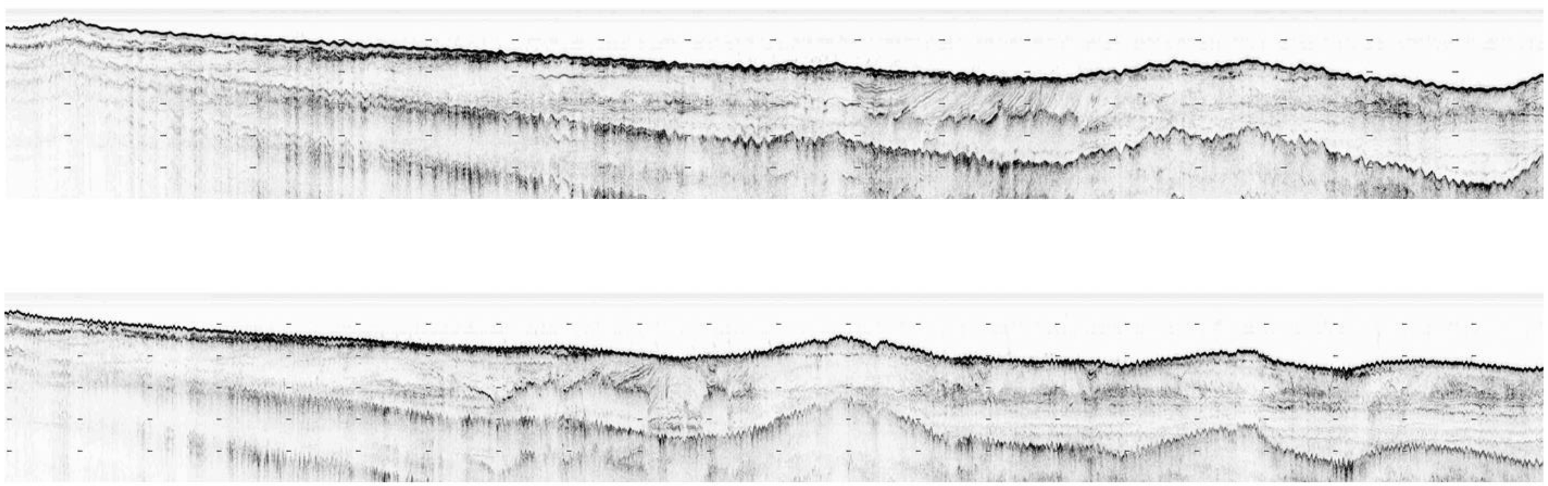
Appendix



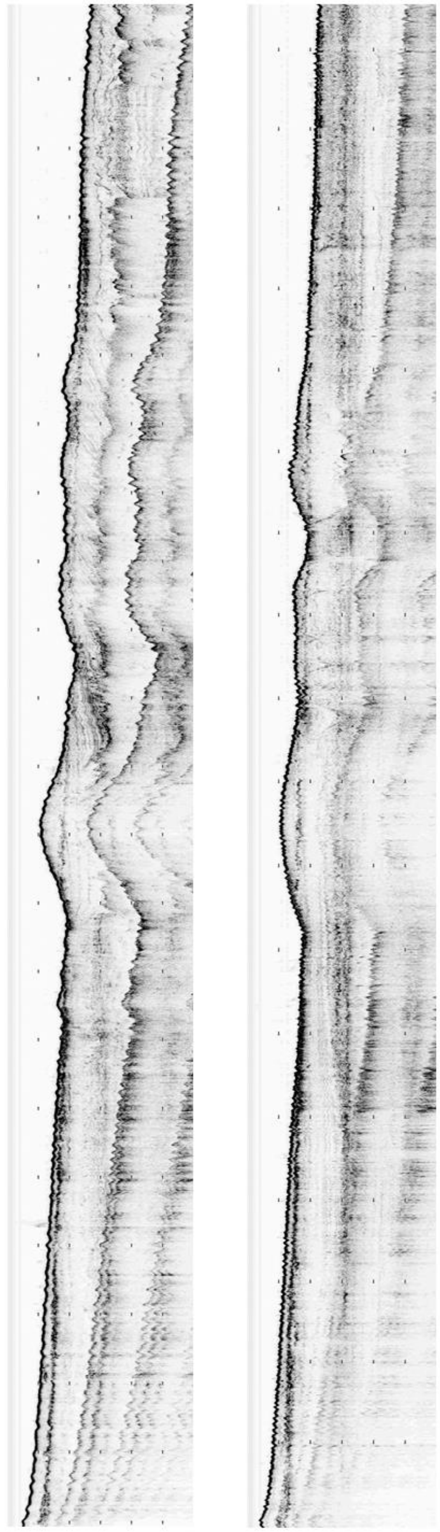
Top: Seismic Line 1

Bottom: Seismic Line 2

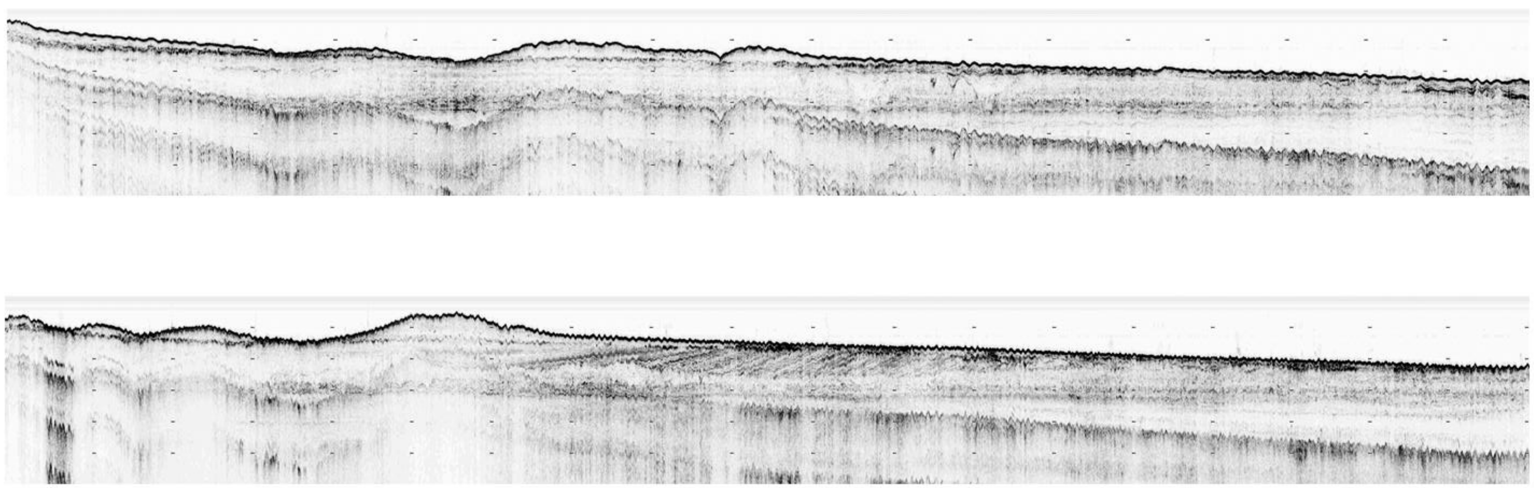


Top: Seismic Line 3

Bottom: Seismic Line 4

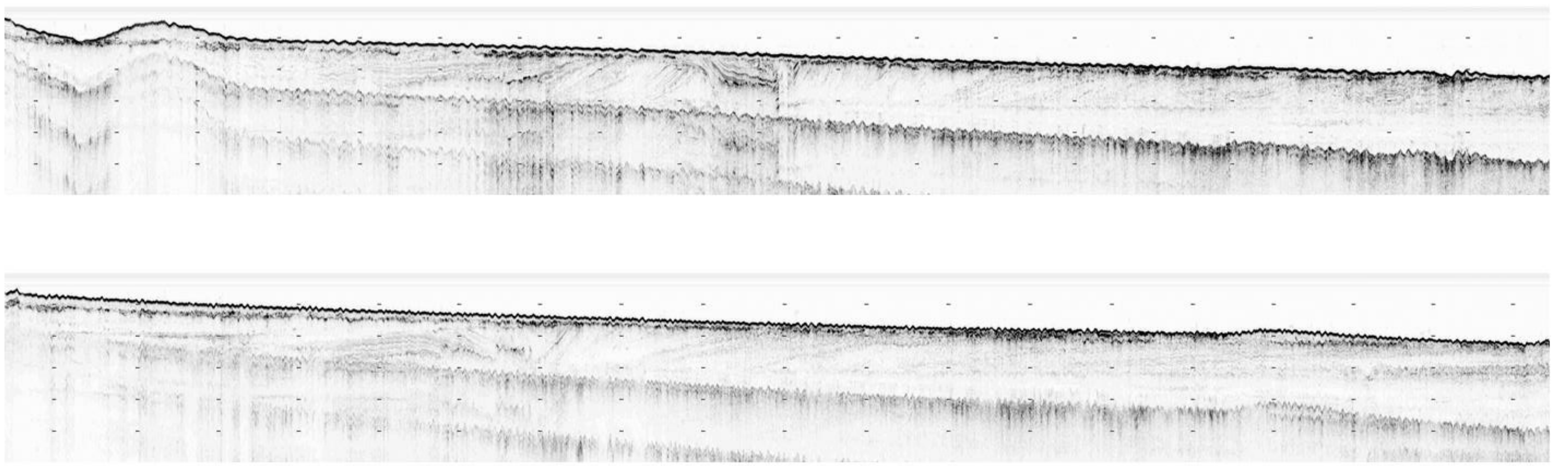


Top: Seismic Line 5
Bottom: Seismic Line 6



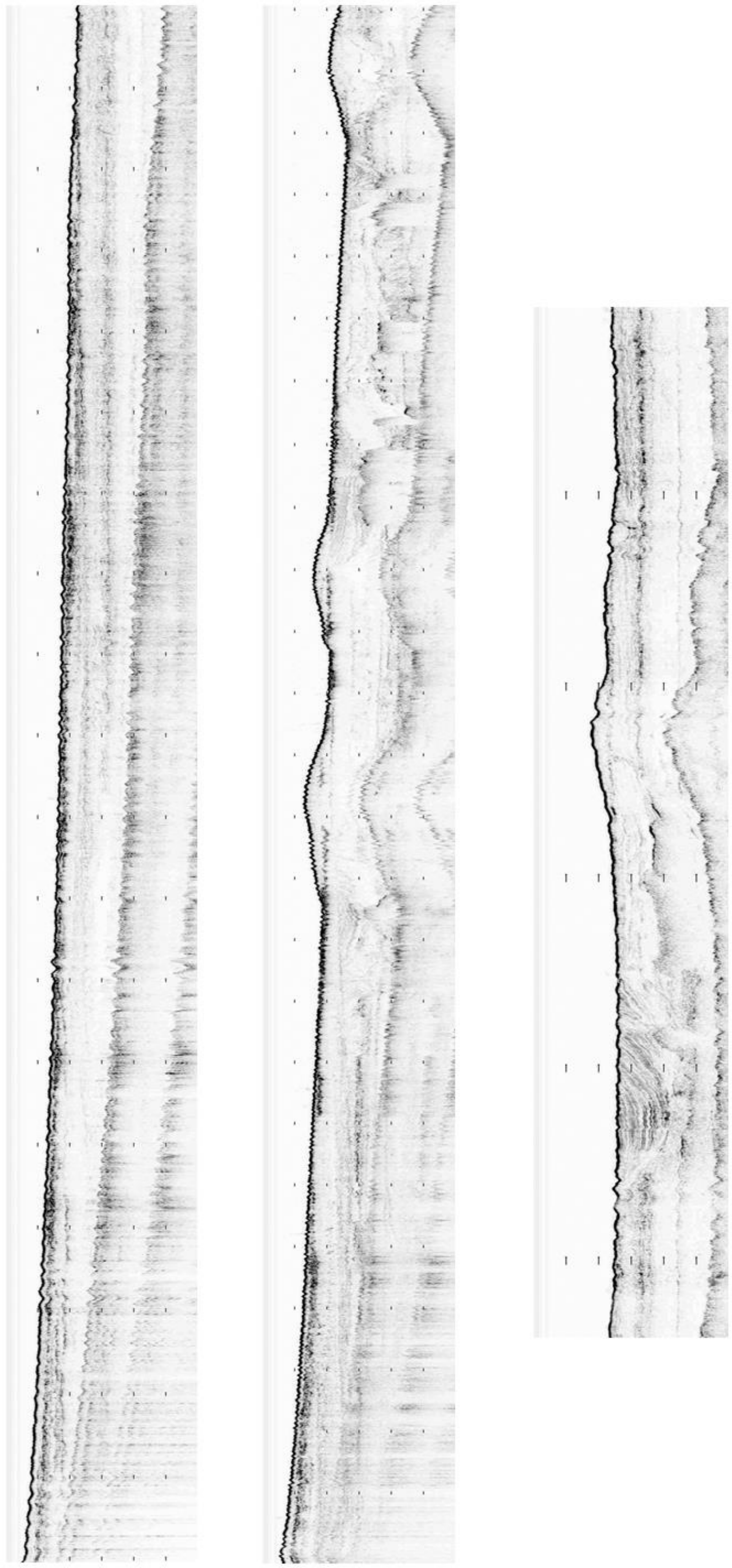
Top: Seismic Line 7

Bottom: Seismic Line 8

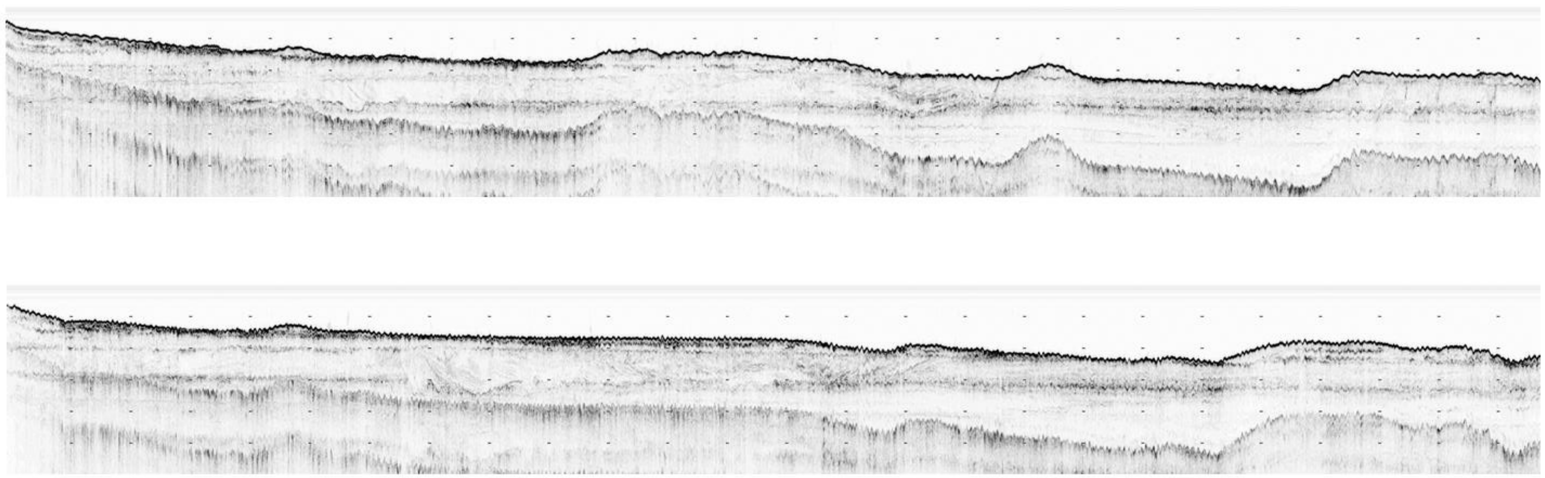


Top: Seismic Line 9

Bottom: Seismic Line 10

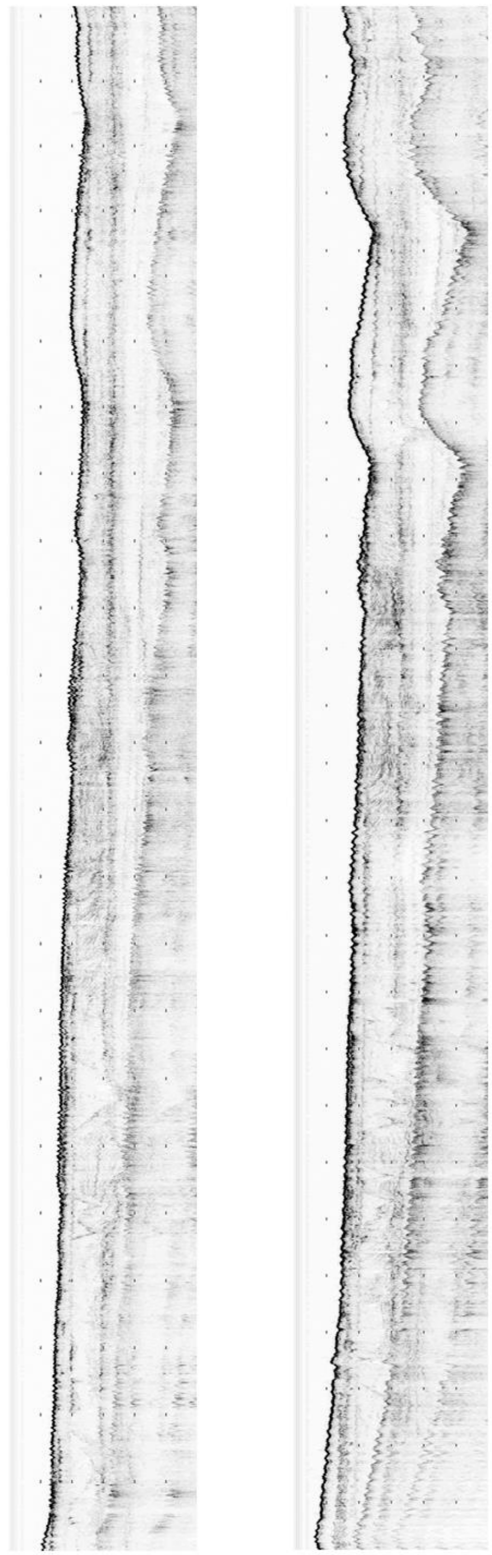


Top: Seismic Line 11
Middle: Seismic Line 12
Bottom: Seismic Line 12 Offshore

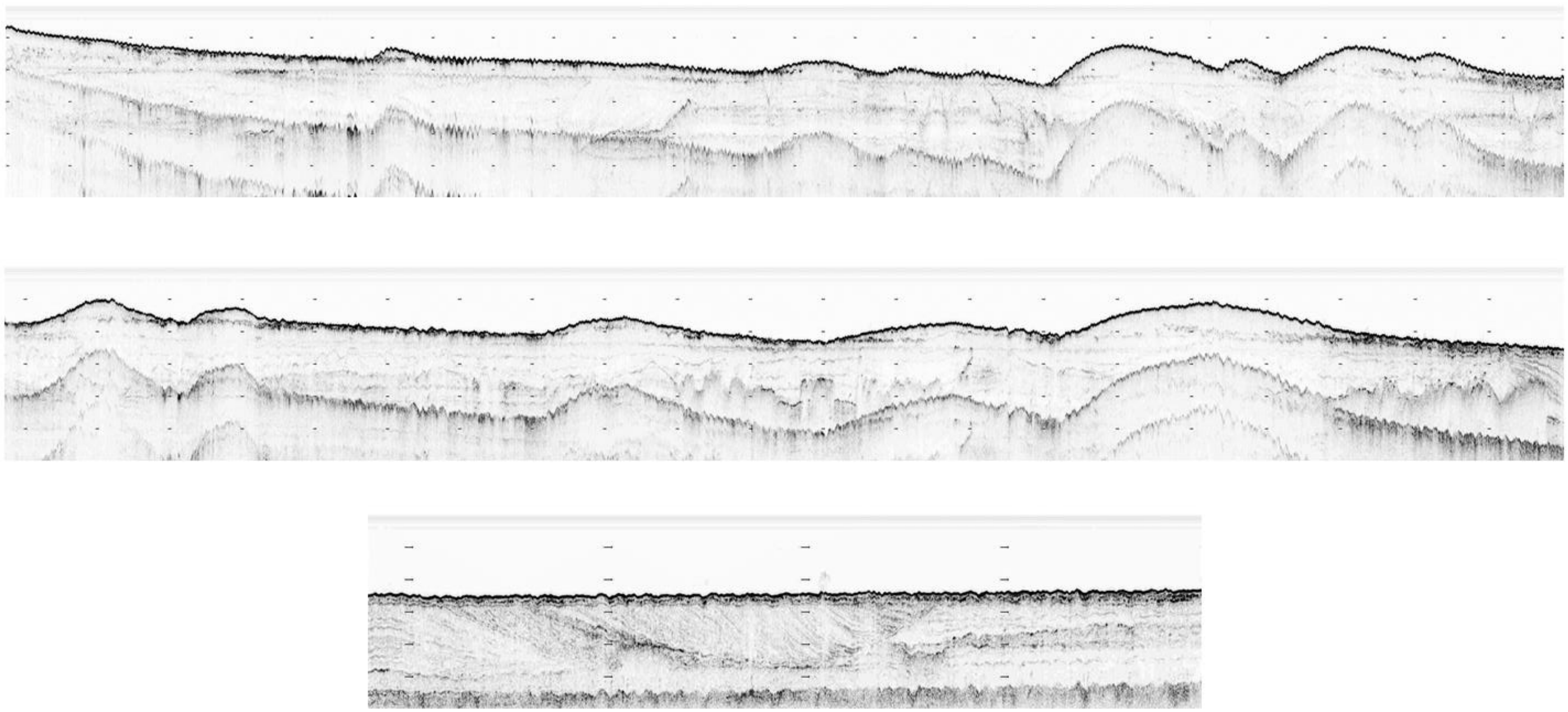


Top: Seismic Line 13

Bottom: Seismic Line 14

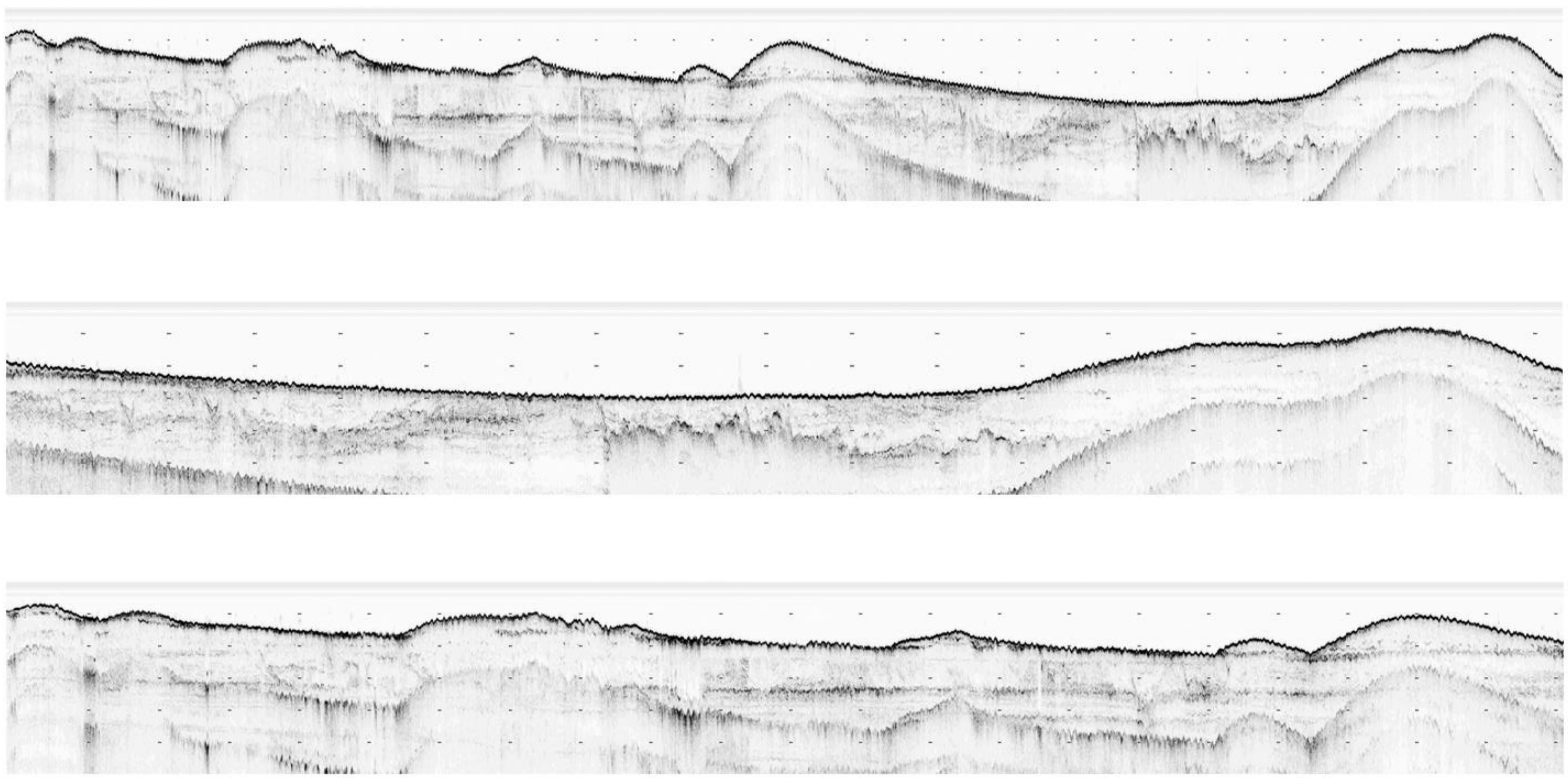


Top: Seismic Line 15
Bottom: Seismic Line 16



Top: Seismic Line 17

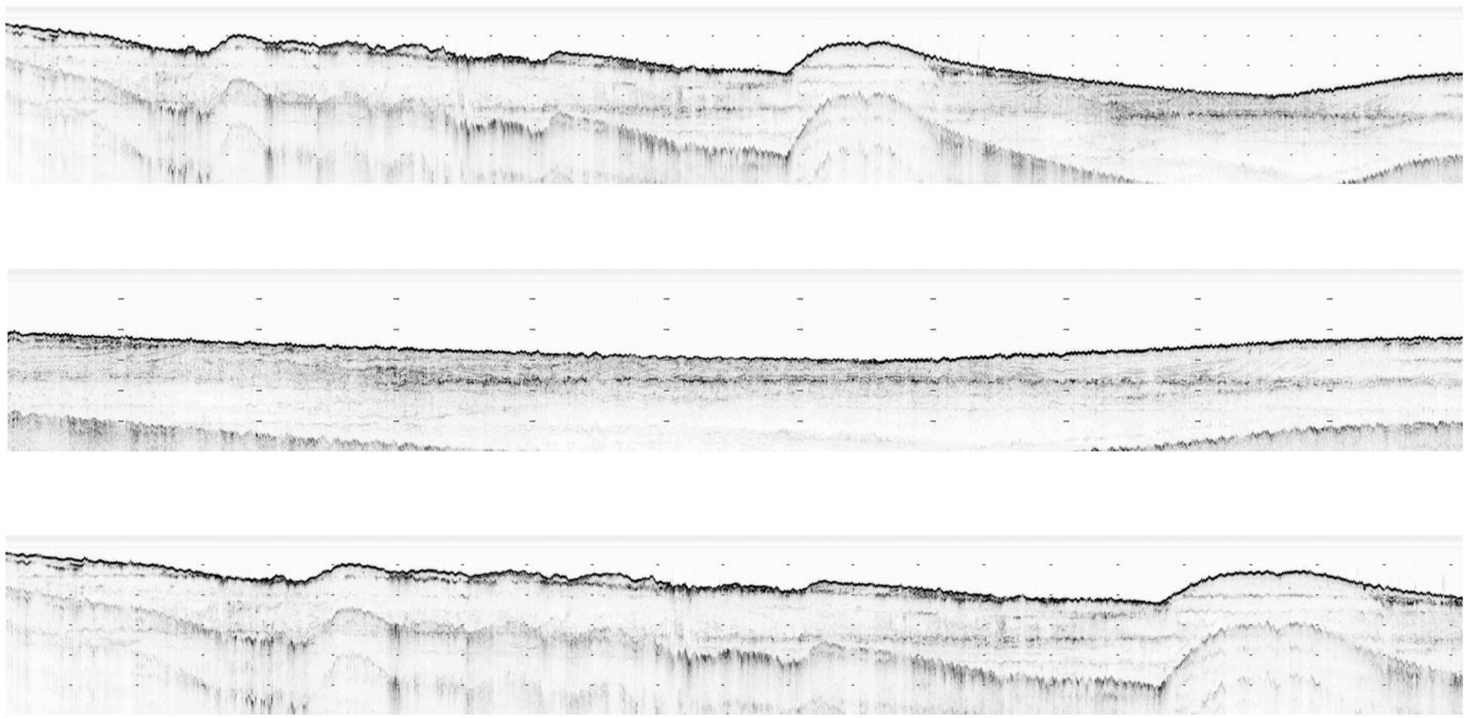
Middle: Seismic Line 18
Bottom: Seismic Line 17 Offshore



Top: Seismic Line 19 East

Middle: Seismic Line 19 Mid

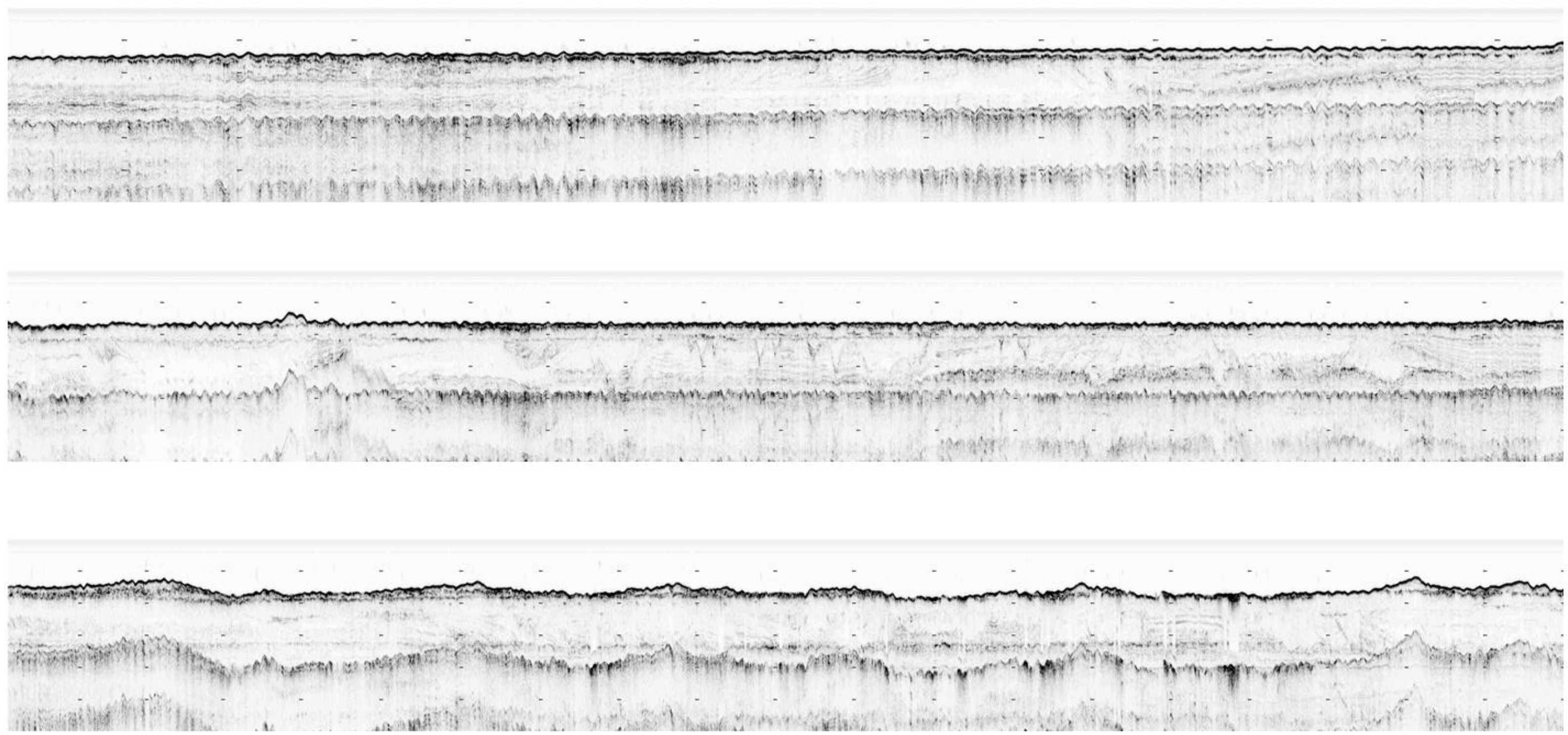
Bottom: Seismic Line 19 West



Top: Seismic Line 20 East

Middle: Seismic Line 20 Mid

Bottom: Seismic Line 20 West



Top: Seismic Line 23 North

Middle: Seismic Line 23 Middle

Bottom: Seismic Line 23 South