LATE QUATERNARY SEQUENCE STRATIGRAPHY OF THE NEW JERSEY CONTINENTAL SHELF

by

JOHN SCOTT CAREY

A Dissertation submitted to the

Graduate School - New Brunswick

Rutgers, the State University of New Jersey

in partial fulfillment of the requirements

for the degree of

Doctor of Philosophy

Graduate Program in Geological Sciences

written under the direction of

Professors Gail M. Ashley and Robert E. Sheridan

and approved by

Gail M. Ashley

Robert E. Sheridan

New Brunswick, New Jersey

October, 1996

ABSTRACT OF THE DISSERTATION

Late Quaternary Stratigraphy of the New Jersey Continental Shelf

by JOHN SCOTT CAREY

Dissertation Directors:

Gail M. Ashley and Robert E. Sheridan

Analysis of high-resolution seismic profiles of the New Jersey continental shelf reveals the stratigraphic architecture resulting from high-frequency, high-amplitude eustatic cycles and nearby continental glaciation. Vibracores™ and the Atlantic Margin Coring Project drill sites provide age and lithologic controls for sequences identified from 1600 km of Uniboom™, Geopulse™, and Minisparker™ profiles. Four sequences can be recognized in the New Jersey shelf stratigraphy, all of which probably post-date stage 6 (~140 ka).

Sequences I and IV likely formed during the major glacial-interglacial sea-level fluctuations (~ 120 m) during isotope stages 6/5e and 2/1, respectively. Sequences II and III record stadial-interstadial sea-level fluctuations from lowstands of -60 to -75 m to highstands of -20 to -30 m. Two interpretations of the age of these sequences are proposed: either sequence II formed during stage 5, and sequence III during stages 4 and 3, or sequence II formed during stage 4 and early stage 3 and sequence III later in stage 3. When eustasy greatly exceeds sediment supply and subsidence, sequences are thin and fragmentary due to intensive fluvial erosion during lowstands and marine erosion during transgressions and highstands. Preservation of sequences is strongly influenced by local factors such as fluvial valley positions and differential

crustal movements.

The development and collapse of a peripheral bulge during glaciation resulted in northward deepening of reflectors in sequences I and IV, and possibly, the northward deflection of the Hudson River to the modern shelf valley shortly before the Late Wisconsin maximum. Sediment supply was greatest during the lowstand and early transgressive intervals of sequences I and IV and throughout sequence III, perhaps reflecting glacial, climatic and vegetational change in the Hudson drainage basin. In the absence of direct evidence of glaciation, similar deposits on ancient shelves could be mistaken for the results of tectonic uplift of the source area.

Seismic studies of passive continental margins can yield useful approximations of global sea level change. Incorporation of sea-level estimates from continental shelves could help to better calibrate models of glacial isostasy. Continental shelf deposits are generally more continuous and more readily dated than terrestrial sections; thus, studies of ice-marginal shelves can shed light on events on the nearby continent.

ACKNOWLEDGEMENTS

I would like to express my gratitude to the staff of the United States Geological Survey at Woods Hole, MA, particularly Ms. Nancy Soderberg and Drs. David Twichell, C. Wylie Poag, and Robert Oldale for their help in identifying and acquiring data that were used in this study. I would also like to thank the Marine Minerals Service, and the New Jersey Geological Survey for monetary support for this work under contract number C29270, and Dr. Peter Rona of Rutgers for extending a research assistantship this spring. I greatly appreciate the work of Dr. John Wehmiller of the University of Delaware in running the mollusk samples in his lab, and for his help in interpreting the results. Finally, I would like to express my deepest thanks to my co-advisers Drs. Gail Ashley and Robert E. Sheridan for their patient reading and re-reading of the critical chapters of this thesis, and for many hours of stimulating discussion. Without all of these people, this work could never have been accomplished.

TABLE OF CONTENTS

LIST OF TABLES

LIST OF FIGURES

CHAPTER ONE: INTRODUCTION

Geologists have long recognized the importance of changes in relative sea level in controlling the lithology and architecture of sedimentary sequences. In the nineteen-seventies, researchers at Exxon found that unconformities and unconformity-bounded depositional sequences could be recognized on seismic records. Analysis of seismic stratigraphic studies of many continental margins led Vail et al. (1977) to postulate that sequence boundaries could be correlated globally and provided a detailed record of eustatic change over time. They produced a global eustatic cycle chart, which was later refined by Haq et al. (1987), and could be used as a geologic time scale for sections where other means of chronological correlation were unavailable.

The validity of the global cycle chart has been challenged. Some researchers have questioned the assumption that eustasy is the dominant control on stratigraphy (e.g. Galloway, 1989). Others have argued that insufficient chronological control exists to demonstrate the global synchrony of regional unconformities (e.g. Miall, 1986). Despite these objections, sequence stratigraphy should be an excellent tool for studying sea-level change on Quaternary passive margins, such as the New Jersey margin (Fig. 1), because of the widespread evidence of globally synchronous, high-amplitude (> 100 m) sea-level changes during the Pleistocene (Bloom et al., 1974; Fairbanks, 1989).

The late Quaternary sea-level history of the New Jersey margin has been the subject of considerable controversy. While there is general agreement that the Sangamon interglacial highstand correlates to oxygen isotope stage 5e (~125 ka; Fig. 2), and reached 5-10 m above modern levels (Mixon et al., 1974; Wehmiller and

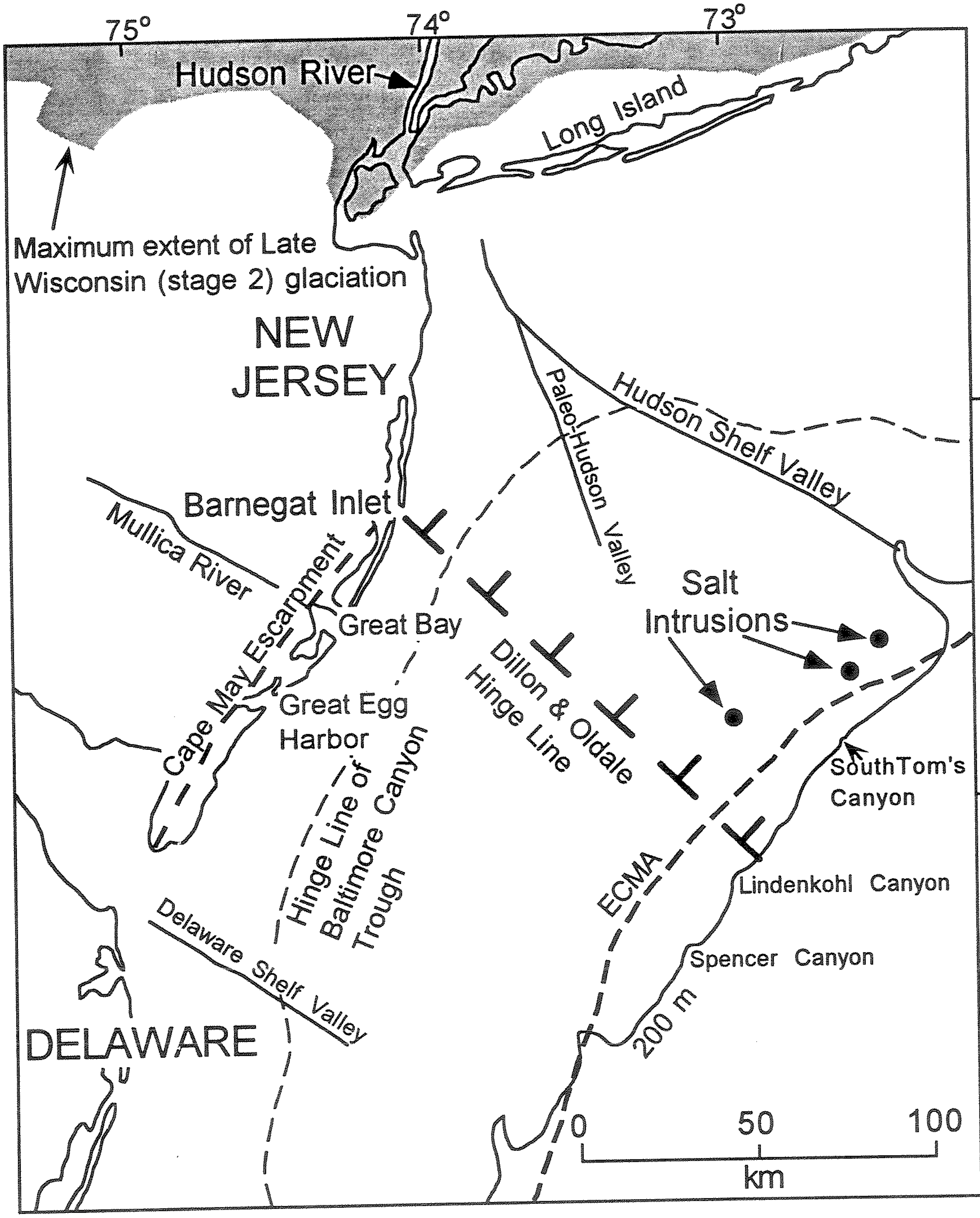


Figure 1. Map of Study Area. Shows selected structural features from Grow et al. (1988), the position of the Laurentide Ice Margin at 18 ka (Dyke and Prest, 1986), and the "hinge zone" of post 15 ka subsidence (Dillon and Oldale, 1978), and various features and structures named in the text.

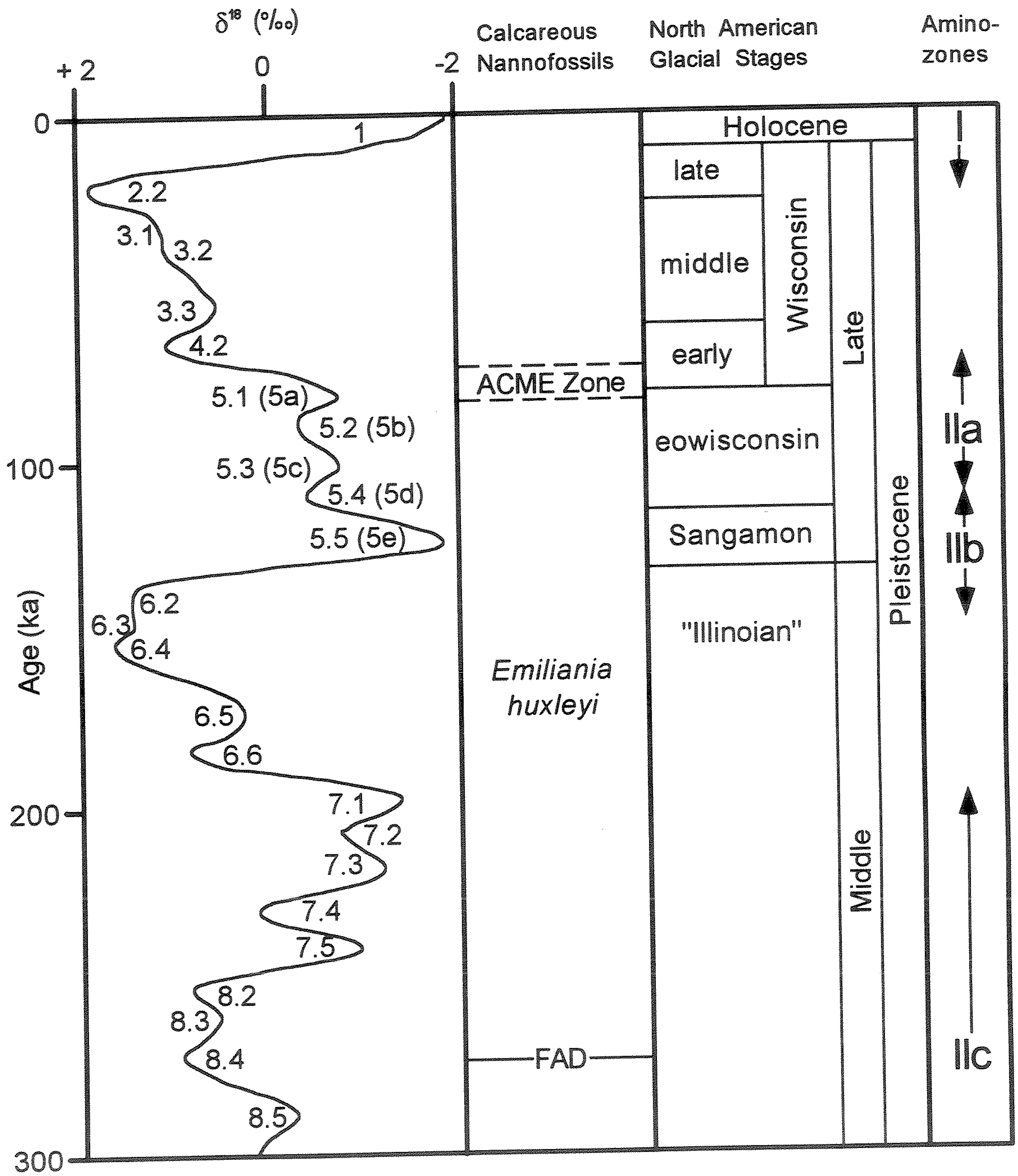


Figure 2. Late Quaternary Timescales. The SPECMAP oxygen isotope curve after Imbrie et al. (1984), along with the approximate ages of the North American glacial stages (after Fulton, 1990), the Atlantic Coastal Plain aminozone chronology (Wehmiller and Belknap, 1982), and calcareous nannofossil biostratigraphy (Berggren et al., 1980).

Belknap, 1982), the elevation of subsequent Pleistocene sea levels is uncertain. Some have argued that middle Wisconsin relative sea level approached modern levels as recently as 36 ka (Milliman and Emery, 1968; Blackwelder et al., 1979; Finkelstein and Kearney, 1988), but these claims have been largely based on radiocarbon dates, which can reflect minor modern contamination in this age range (Thom, 1973; Colman et al., 1989). Ashley et al. (1991) interpreted a -20 m buried shoreline near Barnegat Inlet as resulting from an early stage 3 (Fig. 1) highstand at about 55-60 ka, but Thomas (1992) suggested that it could reflect a stage 5 highstand, and correlated it to events seen on the Gulf Coast (Thomas and Anderson, 1989) and the Delmarva Shelf (Toscano and York, 1992). Analysis of seismic records from the New Jersey shelf indicates that several late Quaternary depositional sequences are preserved on the New Jersey shelf (Sheridan et al., 1993; Carey et al., 1995a). These sequences provide a record of sea-level change on this margin that may help to resolve these controversies.

In addition to shedding light on late Quaternary sea-level change, this study illustrates the stratigraphy developing on a shelf experiencing rates of eustasy that greatly exceed those of subsidence and sediment supply. Several workers (e.g. Thorne and Swift, 1991; Steckler et al., 1993) have used mathematical models to study the effects of rates of sea-level change, tectonic subsidence, and sediment supply on the geometry of stratigraphic successions. Examination of the stratigraphy of the New Jersey shelf provides a test of model predictions for the stratigraphic architecture of a margin where eustasy should be the dominant control.

Another interesting aspect of the New Jersey continental margin is its unique position with respect to the Laurentide Ice Margin (Fig. 1). Glacial processes influenced the stratigraphy both directly, through glacial isostasy (Dillon and Oldale, 1978) and sediment supply changes (Poag and Sevon, 1989), and indirectly through glacio-eustasy. Despite the evident influence of continental glaciation on stratigraphic processes on the New Jersey shelf, it was beyond the Late Wisconsin (stage 2, Fig. 2) ice limit (Dyke and Prest, 1986; Stanford, 1993) and neither glacial till nor ice-rafted sediments are reported from the margin (Mountain, Miller et al., 1994). In the absence of direct evidence of glaciation, certain characteristics of the deposits, such as glacio-isostatic tilting of the basin, and sediment pulses derived from glacial metltwater, could resemble the effects of tectonics. This study examines the attributes of a continental margin influenced by "paraglacial" processes: non-glacial processes that directly result from the actions of data. Possible criteria for the recognition of such margins in ancient deposits are considered.

The purpose of this investigation is to use high-resolution seismic records and available core materials from the New Jersey shelf to construct a late Quaternary sequence stratigraphic framework for the region. Constraints on the ages of these sequences are provided by biostratigraphy, aminostratigraphy, radiocarbon dates, and the global ice-volume signature of oxygen isotope records from deep sea cores (Shackleton, 1967; 1987). Thus, a history of relative sea-level change and sedimentary events on the New Jersey margin is constructed.

The resulting late Quaternary history of the New Jersey margin is used to address three critical problems, which are discussed in chapters 4, 5, and 6. Chapter 4 concerns the effects of high-amplitude, high-frequency eustatic change on the stratigraphic architecture of a slowly subsiding passive margin and assesses characteristics that could be used to recognize similar ancient deposits. Chapter 5 deals with the impact of glaciation on the stratigraphy of the New Jersey margin, and examines criteria for recognizing the influence of paraglacial processes on sedimentary successions. Chapter 6 discusses the age interpretation of the New Jersey shelf sequences, and considers the possible implications for the continental glacial record. It also compares the inferred sea-level record of New Jersey to sea-level curves derived from other continental margins, and to estimates from isotopic records.

CHAPTER TWO: BACKGROUND AND PREVIOUS WORK

2.1 Tectonic Setting

The study area is the NE-SW trending segment of the U.S. Atlantic Coast continental shelf between the Hudson Shelf Valley and the Delaware Shelf Valley from 38° 40' to 40° 30'N and 72° 30' to 74° 40'W (Fig. 1). The region is a classic passive margin formed by the separation of the North American Plate from Africa since the initiation of rifting in the Triassic (Grow et al., 1988). The Baltimore Canyon Trough is the deepest marginal basin on the Atlantic Coast with a thickness of up to 13 km of post-rift sedimentary rocks (Grow et al., 1988). The rate of subsidence increases seaward of the hinge line, which runs parallel to the shoreline approximately 20 km offshore south of Barnegat Inlet through southern New Jersey, then trends offshore to the northeast (Fig. 1). Except for the area around the Great Stone Dome, an early Cretaceous igneous intrusion, the basin gradually deepens from the hingeline toward the East Coast Magnetic Anomaly (ECMA), which marks the transition from transitional to normal oceanic crust and is located on the outer shelf (Fig. 1).

The mean rate of subsidence on the upper slope since the end of the Cretaceous has been ~0.015 mm/yr; nearly half of this is attributed to thermal-tectonic subsidence (Greenlee et al., 1978; Fig. 3). However, subsidence has not been constant throughout this time. During the middle Miocene subsidence rates exceeded 0.05 mm/yr due to sediment loading as a thick wedge of sediment prograded across the shelf (Greenlee et al., 1988). Some authors have attributed the increased sedimentation during the middle Miocene to tectonic uplift of the central Appalachians

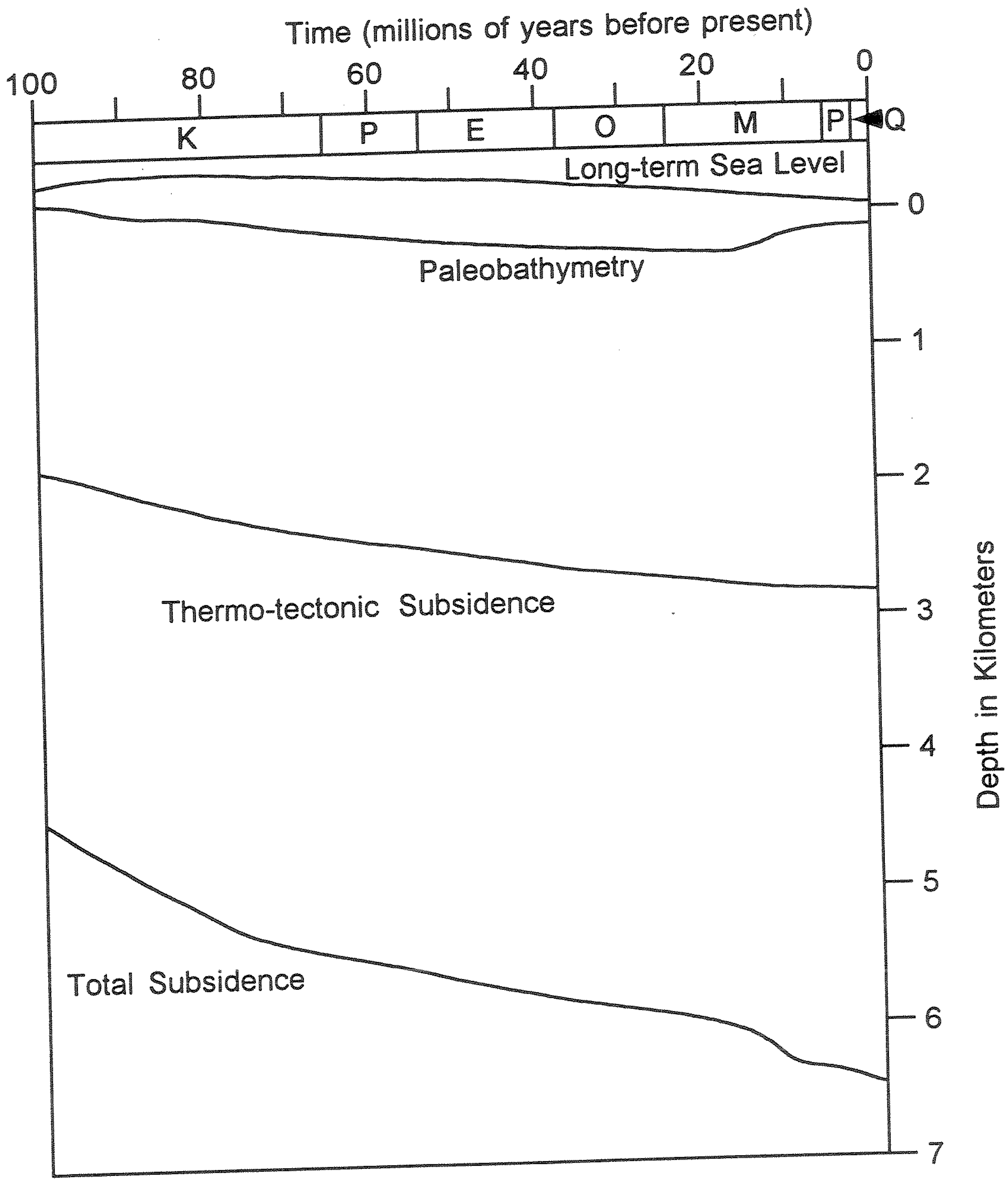


Figure 3. Subsidence History of the New Jersey Margin. Graph of thermal-tectonic subsidence and total subsidence at the COSTB-2 well (Fig. 4) since the initiation of rifting, over the past 100 Ma. Subsidence is corrected for long-term sea-level trends. Modified from Greenlee et al. (1988)

(Mathews, 1975; Hack, 1982; Poag and Sevon, 1989).

Salt tectonics may result in local uplift, altering the trend of increasing subsidence seaward of the hinge line. Salt diapirs and pillows from the early Jurassic evaporite deposits occur along the Atlantic margin within the ECMA (Grow et al., 1988) (Fig. 1). Most of these features are deep swells or small domes, but at least one diapir on the outer shelf pierces the Miocene (Grow, 1980). Sheridan et al. (1993) observed a thinning of the upper Quaternary sediment package on the outer shelf along a high-resolution Uniboom™ seismic record, line 2-16 (Fig. 4) and suggested that it could have resulted from local uplift related to salt movement.

Landward of the hinge line, little or no long-term subsidence has apparently occurred during the late Quaternary. The Cape May Escarpment (Fig. 1) formed at a relative sea-level elevation of 5-10 m above modern sea level; it is speculated that this terrace formed during the last interglacial (Mixon et al., 1974; Wehmiller and Belknap, 1982), which is in close agreement with estimates of stage 5e (Fig. 2) sea level from other stable platforms (Broecker and Thurber, 1965; Ku et al., 1974; Veeh et al., 1979). The outcropping of Cretaceous units on the seafloor off northern New Jersey (Allen-Lafayette, 1996) may imply some uplift north of where the hingeline of the Baltimore Canyon Trough veers offshore (Fig. 1).

2.2 Regional Glacial History

The Hudson River drains much of eastern North America when glacial ice blocks the St. Lawrence River (Fig. 5), as it did for much of the late Pleistocene (Teller, 1987). Major glacial advances can thus be expected to increase the area of the

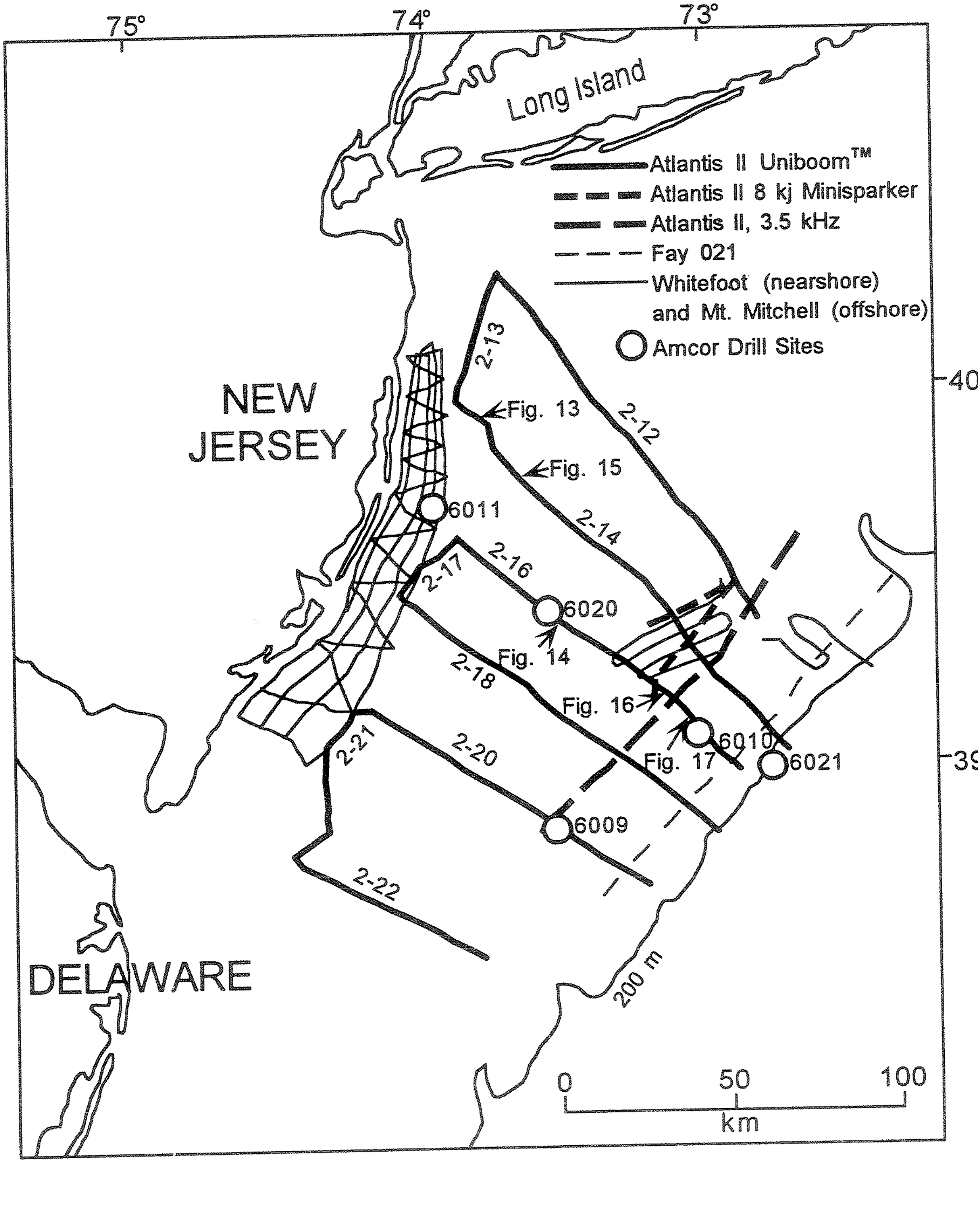


Figure 4. Location map for seismic records and cores used in study.

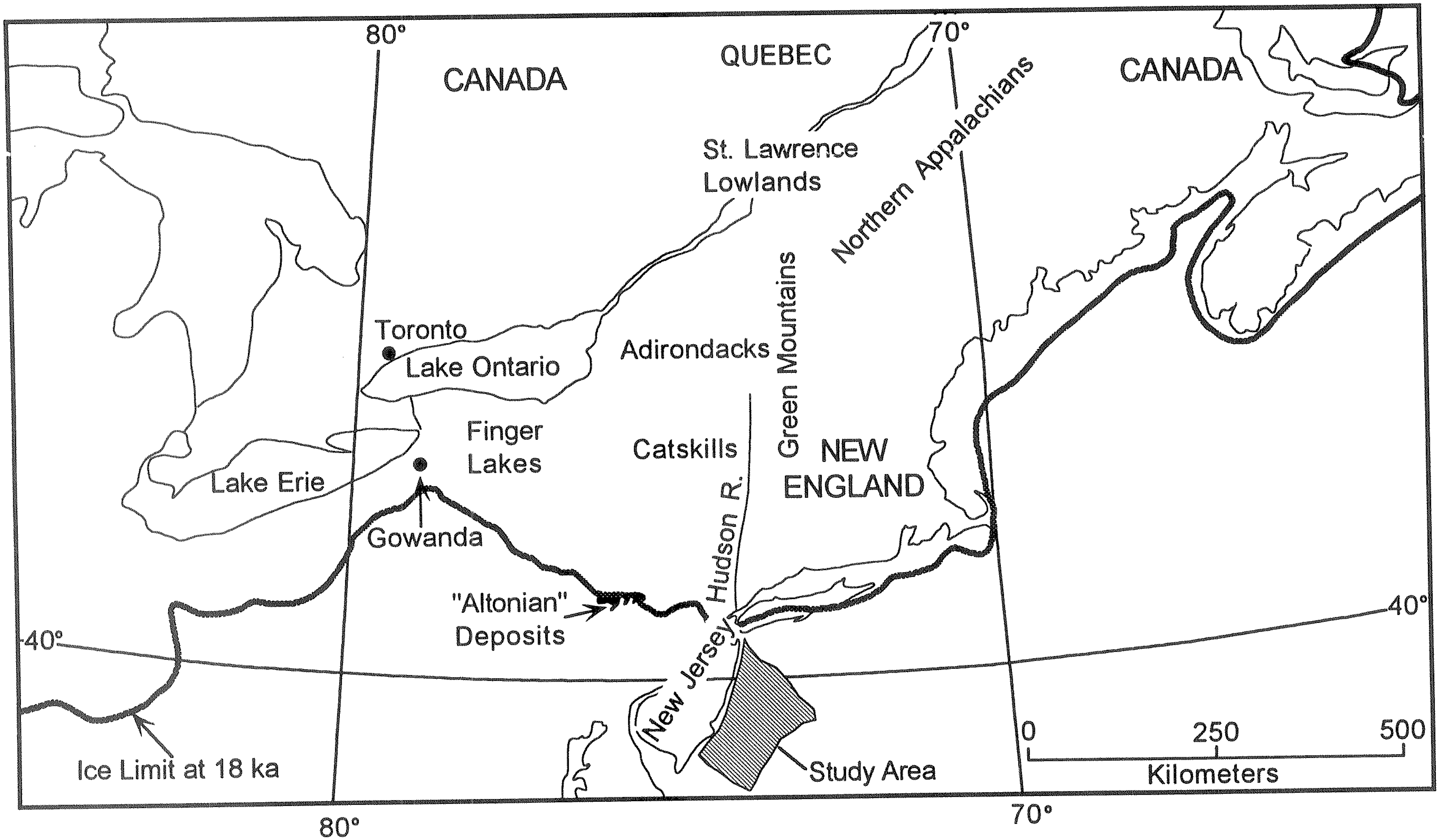


Figure 5. Regional Map. Shows location of Laurentide Ice Margin (after Dyke and Prest, 1986), "Altonian" deposits in Pennsylvania and New Jersey (Ridge et al., 1990), and various locations mentioned in text.

Hudson drainage basin, through capture of the Great Lakes lowlands, affecting its discharge, and the sediment supply to the study area. Oxygen isotope records from deep sea cores indicate that numerous major glacial events occurred during the late Pliocene and Pleistocene (Hays et al., 1975; Imbrie et al., 1984), far more than the four glacial periods traditionally recognized by American glacial stratigraphers (Flint, 1971). However, evidence of glacial events prior to the late Wisconsin (Fig. 2) is sparse in eastern North America, because of the tendency for glaciers to erode the record of older glaciations, and the difficulties in dating material.

Despite the difficulty of assigning ages to Quaternary sediments in the region, there is some consensus on the correlation of the St. Lawrence-Great Lakes lowland (Fig. 5) glacial stratigraphy to the SPECMAP oxygen isotope stages. This study will generally adhere to the terminology of Fulton (1989), which correlates the early, middle, and late Wisconsin to isotope stages 4, 3, and 2, respectively (Fig. 2). The Sangamon, correlated to stage 5 by Fulton (1989), is traditionally described as an interglacial. However, evidence of an ice buildup during this time exists in southern Quebec, where the presence of varved lacustrine clay deposits implies that the St. Lawrence River (Fig. 5) was dammed by ice (Lamothe, 1989; Clet and Occhietti, 1994). For this reason, the use of the term Sangamon in this study is restricted to stage 5e, when conditions were comparable to or warmer than the modern, as suggested by Richmond and Fullerton (1986). For the "late Sangamon", or isotope stages 5a-d, the term "eowisconsin" is used (Fig. 2).

The extent of glaciation in the northeastern United States during the early Wisconsin is uncertain (Muller and Calkin, 1993). Some workers (e.g. Koteff and

Pessl, 1985) interpreted the "Lower Till" in New England as resulting from an early Wisconsin glacial advance. Similar ages have been suggested for tills beyond the late Wisconsin limit in northern New Jersey and Pennsylvania (Crowl and Sevon, 1980) (Fig. 5). Other authors have regarded these tills as stage 6 or older (e.g. Ridge et al. 1990; Oldale and Colman, 1992), which is more consistent with the SPECMAP record, which indicates significantly less ice volume during stage 4 than during stage 2 (Imbrie et al., 1984; Fig. 2).

There is evidence of a significant glacial advance during stage 4 in Canada. The Chaudhière Till in southern Quebec, which may be of stage 4 age, was formed by ice moving out of the Appalachians that was confluent with the Laurentide in the St. Lawrence Lowlands (Karrow and Occhietti, 1990). The Sunnybrook Drift, a diamicton unit in the Toronto area (Fig. 5), has been variously interpreted as a till from an early Wisconsin glacial advance (Karrow, 1967; Hicock and Dreimanis, 1989) or a glacial-lacustrine unit (Eyles and Eyles, 1983). Either interpretation implies that Evidence for early and middle Wisconsin events in New York state is largely restricted to western New York (Muller and Calkin, 1993). Fullerton (1986) regards the "brown till" at Gowanda, New York (Fig. 5) as lower Wisconsin, and Dreimanis (1992) tentatively correlates it to the Sunnybrook drift in the Toronto area. Lacustrine sediments in the Finger Lakes region suggest that ice dammed their drainage basins during the early or middle Wisconsin, but no till units from those times have been clearly identified (Muller and Calkin, 1993).

Ice retreated from the Great Lakes region during the middle Wisconsin, resulting in a falling lake levels and the end of varve sedimentation in the Lake Erie basin (Fig.

5; Dreimanis, 1992). However, lacustrine sedimentation continued in the St. Lawrence Valley during this time, and no evidence of subaerial weathering is seen on the older deposits (Lamothe, 1989), suggesting that ice may have continued to block the St.Lawrence Valley throughout the middle Wisconsin.

During the late Wisconsin, the ice advanced through New York after 25 ka (Muller and Calkin, 1993) and reached as far south as northern New Jersey and Long Island (Fig. 1, Dyke and Prest, 1986; Stanford, 1993). Ice began to disappear from the region around 18 ka, retreating into the Lake Ontario basin (Fig. 5) around 13.5 ka (Teller, 1987). After that time, the Hudson drained the western Great Lakes, which had previously been drained by the Mississippi River. Meltwater from the Great Lakes continued to flow through the Hudson until shortly after 12 ka, when ice retreated from the St. Lawrence Valley (Fig. 5, Teller, 1987).

2.3 Impact of Glacial Isostasy on the Study Area

The degree of isostatic depression and/or uplift for different parts of the state during the late Wisconsin is a subject of ongoing debate. Clark et al. (1978) published a model for global isostasy that divided the planet into regions of similar sea-level history. According to the model, the Atlantic Coast should have been affected by a peripheral bulge as far south as North Carolina. Several subsequent models (e.g. Peltier, 1986; Tushingham and Peltier, 1992) have tended to reinforce the notion of significant isostatic influence at considerable distances from the ice sheet. Tushingham and Peltier (1992) calculated a relative sea level (RSL) curve for Brigantine, New Jersey. Using the 196 km thick lithosphere, their results suggest that the crust was

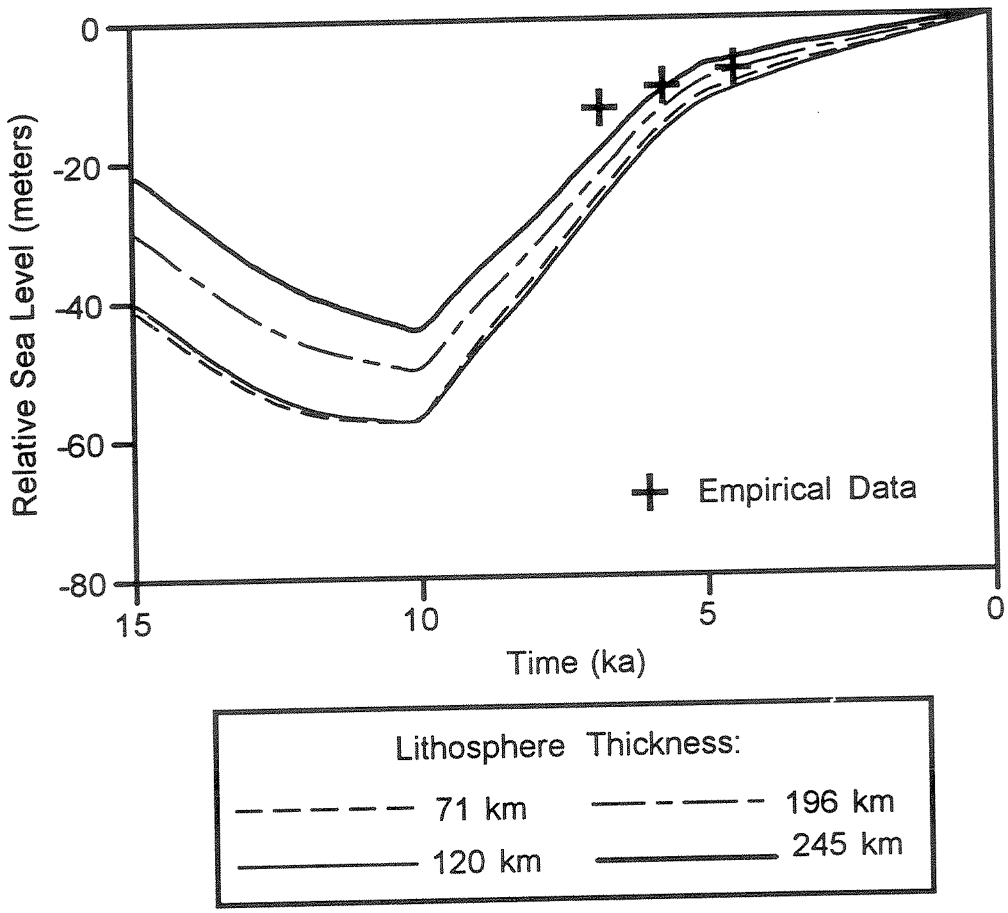


Figure 6. Theoretical Relative Sea Level Curve for Brigantine, NJ. Calculated using the ICE-3G model for four different lithosphere thicknesses (Tushingham and Peltier, 1992). Also shown are radiocarbon dated relative sea-level indicators from the region. Modified from Tushingham and Peltier (1992).

still isostatically depressed at 15 ka, with RSL at approximately -25 m (Fig. 6). Because of the passage of the peripheral bulge, RSL fell to a minimum of roughly -45 m at 10 ka, then rose to modern levels. However, even over the past 8000 years, which is the data base for most of the Atlantic coast locales south of the ice sheet, the rate of observed submergence has been significantly less than predicted (Tushingham and Peltier, 1992). Peltier (1994) unveiled a new ice model that produced better results for some Atlantic Coast sites, but no new curve for New Jersey has yet been published.

Geologic studies of the continental shelf have suggested a more geographically limited extent of glacial-isostatic influence. Dillon and Oldale (1978) demonstrated that late Wisconsinan shorelines preserved on the shelf were essentially horizontal from Virginia to a "hinge zone" running southeastward from Barnegat Inlet, New Jersey (Fig. 7). They suggested that the areas north and south of that hinge zone were on different crustal blocks that responded largely independently to ice loading. Ice loading resulted in relative subsidence of the northern half of the study area since 15 ka (Dillon and Oldale, 1978). In contrast, south of that hinge zone, the shelf has been essentially stable (Dillon and Oldale, 1978; Blackwelder, 1980).

Stable may be a misleading description, however, in view of the evidence of significant modern subsidence along the U.S. Atlantic Coast. Historical tide gauges located on or near bedrock, from New York City and Philadelphia record mean water level rises of 2.6 and 2.7 mm/yr, respectively (Lyles et al., 1987). This is substantially higher than the eustatic rise, estimated at 1.15 mm/yr (Nakiboglu and Lambeck, 1991). Daddario (1961) reported a sea level of about -9 m near Atlantic

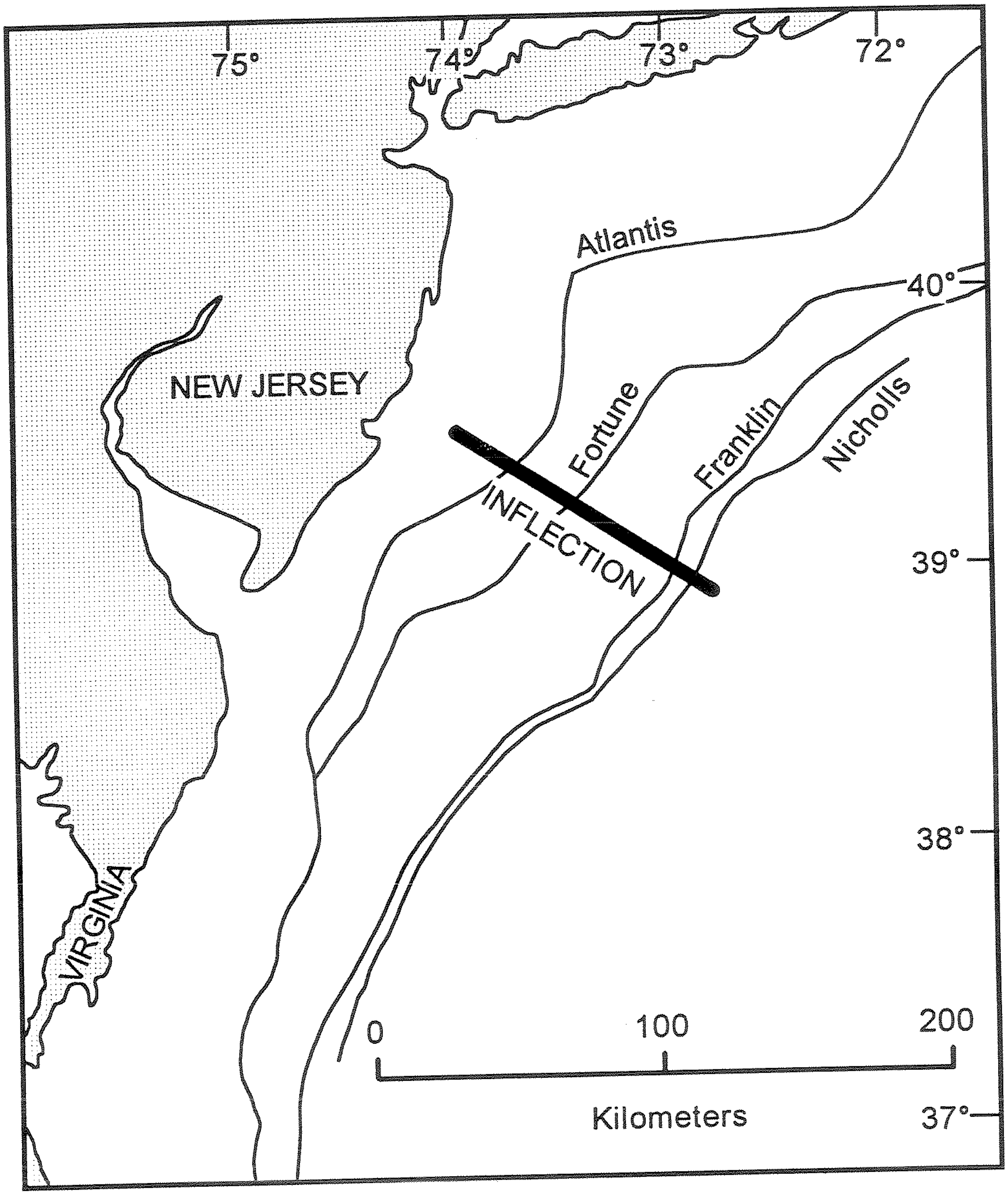


Figure 7. Paleo-shorelines on the Atlantic Coast and the inflection zone identified by Dillon and Oldale (1978). South of the inflection, shorelines are horizontal, north of it they plunge to the south. Modified from Dillon and Oldale (1978)

City that dated at about 5,000 years B.P., while other "stable" regions experienced sea levels of -4 m on the Gulf Coast (Curray, 1963), and +2.5 m in Australia (Fairbridge, 1961; Chappell, 1983). Clark and Lingle (1979) assume that no eustatic change occurred over that time period.

Davis and Mitrovica (1996) were able to generate model predictions that approximated the observed modern sea-level rise on the Atlantic Coast, by using an increased lower-mantle viscosity. While this model predicts current trends well, the collapsing "bulge" is located farther south on the Atlantic coast than New Jersey. In contrast, the larger-amplitude adjustments of the late Pleistocene inferred by Oldale imply a collapsing bulge that was largest to the *north* of New Jersey. Possibly, this reflects crustal responses of different frequencies affecting different geographical areas.

At present, no existing isostatic models successfully reproduce the effects of glacial isostasy in the region over the past 15 ka. Whether the apparent contradiction is due to a failure of the isostatic adjustment model or due to a failure of the input data (ice volume, thickness and location at various times, lithosphere thickness, mantle viscosity) is unclear. Continuing crustal adjustment will be a source of error in estimating past sea levels from the New Jersey data. Variations in the sea levels estimated from different parts of the study area may give some clues to past crustal responses to glacial loading.

2.4 Physiography and Oceanography

The New Jersey shelf is broad and gently sloping ranging in width from approximately 120 km in the south to more than 150 km in the north, and the mean gradient is approximately 0.0001. The steepening of slope associated with the shelf break begins between 120 and 160 m water depth and the width of the shelf ranges from 120-150 km.

The hydrographic regime is mixed energy, with tidal range of 1-2 m and mean significant wave heights of roughly 1 m. The shelf is generally classified as storm-dominated (Duane et al., 1972). Throughout the Quaternary, the Hudson River (Fig. 1) has been the principal sediment source for the study area (Poag and Sevon, 1989). However, the modern shelf receives little sediment supply because of trapping of sediments in lagoons and estuaries (Clarke et al., 1983). The modern Hudson delivers approximately 870,000 metric tons of fine sediment to its lower reaches, but the net flow of sediment is from the ocean into the Hudson estuary (Ellsworth, 1986).

2.5 Quaternary Stratigraphy and Relative Sea Level

There has never been a comprehensive study of regional Pleistocene stratigraphy, but New Jersey has been the site of many investigations of late Pleistocene - Holocene ($\delta^{18}O$ stages 1 and 2) sea-level changes (Stuiver and Daddario, 1963; Milliman and Emery, 1968; Dillon and Oldale, 1978; Blackwelder et al., 1979; Psuty, 1985). The post-glacial stratigraphy has also been described locally in papers dealing with high-resolution seismic studies of sand ridges (Stahl et al., 1974; Stubblefield et al., 1984; Rine et al., 1991; Snedden et al., 1994).

More recently, Milliman et al. (1990) and Davies et al. (1992) have described the internal stratigraphy of upper Quaternary sediment wedges on the middle and outer shelf of New Jersey. Their stratigraphy did not extend below the "R-reflector", a regional unconformity named by McClennen (1973), and believed to have been cut during the lowstand associated with the last glacial maximum ($\delta^{18}O$ stage 2).

Studies of older Quaternary stratigraphy have been few. AMCOR drill sites revealed Pleistocene sediment thicknesses of up to 170 m at site 6010 (Fig. 4), and over 300 m of middle to upper Pleistocene on the upper slope at site 6021 (Poag, 1979). Poor recovery rates (less than 30%) limited the usefulness of these sites for lithostratigraphy, but Poag (1979) reports that sediments from inner shelf cores, such as site 6011 (Fig. 4) were probably marginal marine, lagoonal and fluvial in origin. Outer shelf sites, such as 6010 (Fig. 4) penetrated a complex interfingering of lagoonal, inner shelf and middle to outer shelf strata. Knebel et al. (1979) mapped a large buried valley below the R-reflector beneath the mid-shelf, which he ascribed to a Pleistocene Paleo-Hudson River (Fig. 1). Harris (1983) mapped several reflectors below the R-reflector on the outer shelf in the northern part of the study area.

Groot et al. (1995) examined benthic foraminifera, pollen and amino acid ratios on mollusk shells from several AMCOR sites on the New Jersey margin, particularly hole 6021 on the upper slope (Fig. 4), which they tied into airgun seismic records. They interpreted the sediments at this site as representing an essentially continuous 220 m thick section deposited over the past 500 ka, and were able to distinguish glacial and interglacial deposits on the basis of pollen. Mountain et al. (1994) used nannofossil biostratigraphy and physical properties to correlate a nearly continuous section deposited between 500 ka and 73 ka at ODP sites 902-904 (Fig. 4) on the New Jersey slope, but sediments from stages 4 and 3 were absent at these sites. A prominent reflector, p1, traced from Site 902 to the shelf edge was found to correlate to the stage 7 / stage 6 transition.

Milliman and Emery (1968) produced the first RSL curve for the U.S. Atlantic continental shelf that extended beyond the late Wisconsin maximum (Fig. 8). Based on radiocarbon dates from a variety of sea-level indicators (shells, saltmarsh oolites) collected from Maine to Florida, they concluded that sea level for the lowstand from 21 to 15 ka was approximately 130 m below modern levels and that RSL was near its modern elevation from 30 to 35 ka. Because similar results had been derived from several other continental shelves, they concluded that the RSL curve for the U.S. Atlantic Coast approximated the eustatic curve.

MacIntyre et al. (1978) questioned the timing of the maximum lowstand on the Milliman and Emery curve, arguing that the curve was based on many dates from shell material that could have been transported or was of doubtful sea-level significance. Dillon and Oldale (1978) revised the curve based on an examination of the elevation of paleo-shorelines on the continental shelf. Their corrected curve (Fig. 8) inferred that the northern samples gave erroneously deep results and suggested that RSL never fell below -100 m over the past 25,000 years.

Bloom (1983) reviewed the evidence for high sea levels during the middle Wisconsin and concluded that there was no rigorous evidence of ESL within 15 m of modern levels during that time. Reiterating the arguments of Thom (1973), Bloom (1983) discounted all the middle Wisconsin radiocarbon dates obtained as suspect, and said that the evidence was not sufficiently compelling to cast doubt on the ice volume estimates obtained from the $\delta^{18}O$ curve.

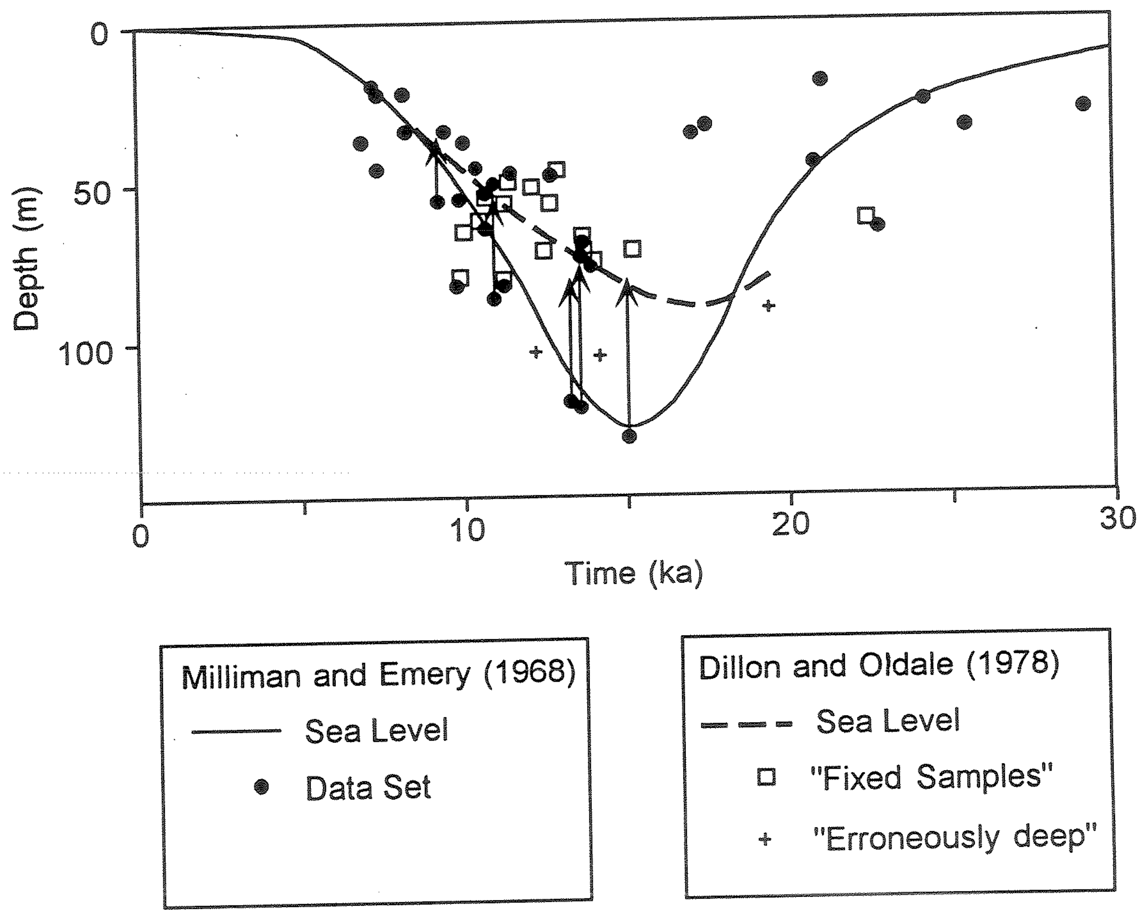


Figure 8. Late Quaternary Sea-Level Curves for the Atlantic Continental Margin Over the Past 30 ka. Shows the Milliman and Emery (1968) and Dillon and Oldale (1978) curves, along with the data these curves were based on. Arows indicate correction for glacial isostasy on northern samples.

Yet the controversy over the elevation of mid-Wisconsin sea-level highstands refuses to go away. Finkelstein and Kearney (1988) interpreted radiocarbon dates on Maryland peats as evidence of a mid-Wisconsin high stand ca. 30 ka. However, Colman et al. (1989) and Toscano (1989) challenged the validity of their dates, arguing that there was evidence of contamination by carbon from modern rootlets.

Recent studies have identified a submerged shoreline on high resolution seismic lines obtained 2 km offshore of Barnegat Inlet, New Jersey at a depth of -20 m, which has been assigned a middle Wisconsin age (Ashley et al., 1991; Wellner et al., 1993). They based their interpretations on its stratigraphic position unconformably overlying the Sangamon, and on a correlation to the most recent -20 m sea level stand identified from the uplifted reef terraces of New Guinea (Bloom et al., 1974). Thomas (1992) suggested that the shoreline formed during a late stage 5 highstand, arguing that sea level was never higher than -40 m during stage 3 based on evidence from the Gulf Coast (Thomas and Anderson, 1989).

Deposits from a -20 m sea-level stand have been reported from many areas of the U.S. East Coast (Blackwelder et al., 1979; Ashley et al., 1991; Toscano and York, 1992), Gulf Coast (Thomas and Anderson, 1989) and the southern coast of Australia (Murray-Wallace et al., 1992). The age of these deposits and their synchrony is still controversial. Blackwelder et al. (1979) reported a series of sea-level indicators at -20 m offshore of South Carolina that gave radiocarbon ages ranging from 33 to 21 ka, but radiocarbon dates on shell material of this age may result from minor modern contamination (Thom, 1974), and should be considered suspect unless supported by other, independent age estimates. Murray-Wallace et al. (1992) cross-checked their radiocarbon dates with amino acid ratios, and interpreted their sediments as deposited during a 35-40 ka high stand, but the amino acid ratios are also consistent with an earlier Stage 3 age (~55 ka), such as that inferred by Ashley et al. (1991). Amino acid racemization was also used by Toscano and York (1992) on shells from Maryland, yielding age estimates suggesting late stage 5. Due to the limits of radiocarbon dating and the inaccuracy of amino acid dating the ages of these deposits remain questionable. Wehmiller, 1984 considers it nearly impossible to resolve age differences of less than 25%. Coral reef terraces in Barbados and New Guinea suggest sea levels near this elevation during isotope stages 5c, 5a and early stage 3 (Bloom et al., 1974). Thus, the various -20 m shorelines reported may record events of varying ages.

Sheridan et al. (1993) presented the most detailed summary of Wisconsin RSL history from the Atlantic Coast to date. Based primarily on sequence stratigraphic analysis of USGS Uniboom™ line II-16 (Fig. 7), we interpreted the series of sequence and systems tract boundaries and inferred sea-level changes summarized in Table I.

Table I. Wisconsin Sequence Stratigraphy of the New Jersey Shelf*

Age	Glacial Age	Systems Tract	O^{18} Stage	Reef Peak	Depth	Distance Offshore
14-18 ka	Late Wisconsin	LST	2		~120 m	~110 km
28-40 ka	Mid Wisconsin	HST	3a	II, IIIa,	~ 35 m	~ 35 km
~ 45 ka	Mid Wisconsin	LST	3b	IIIb	~ 60 m	~ 75 km
~ 55 ka	Mid Wisconsin	HST	3c		~ 20 m	~ 2 km
~ 70 ka	Early Wisconsin	LST	4a	IV	~ 70 m	~ 55 km
~ 75 ka	Early Wisconsin	LST	4c		~ 75 m	~ 60 km

* Source: Sheridan et al. (1993)

This study traces out the surfaces identified by Sheridan et al. (1993) across the

New Jersey shelf, and other reflection surfaces recognized above the unconformity believed to represent the Stage 6 low stand. This verifies the regional nature of these unconformities, distinguishing them from local surfaces created by marine erosion or delta abandonment. Mollusk shells from the AMCOR sites are used to obtain age estimates using amino acid racemization ratios to help constrain the timing of these events, helping to verify or reject the Sheridan et al. (1993) chronology.

CHAPTER THREE: DATA BASE AND METHODS

3.1 Seismic Records

Seismic data for the New Jersey shelf is publicly available from the United States Geological Survey (USGS), which performed several high resolution surveys of the region. The cruises used in this study are summarized in Table II. The primary data used in this study are the Atlantic II Leg 2 lines 12-14, 16-18 and 20-22, which comprise about 750 km of good quality Uniboom™ data collected by the USGS (Fig. 4). These lines were linked only on the inner shelf and on the slope, however. The Fay 021 line was selected as a useful outer shelf tie line. The ties in the mid-shelf were obtained through use of the Atlantis II 3.5 kHz data and Mount Mitchell data for uppermost reflectors, and the Atlantis II Mini-sparker data for deeper reflectors. Whitefoot data were used primarily to tie the lines into the onshore stratigraphies developed by Ashley et al. (1991) and Wellner et al. (1993).

Table II. Seismic Records Used in Study

Cruise	Dates	Type of Data	Location	Length
Atlantis II, Leg 2 Atlantis II, Leg 2 Atlantis II, Leg 3	05/09/75 - 05/11/75 05//04/75	Uniboom™ 3.5 kHz bathymetry 8 kJ Mini-sparker	regional mid-shelf mid-shelf	750 km 75 km 45 km
Fay 021		8 kJ Mini-sparker	outer shelf	150 km
Whitefoot	06/13/81 - 06/17/81	Uniboom™	inner shelf	250 km
Mount Mitchell	05/19/76 - 05/21/76	Uniboom™	mid-shelf	300 km

Of all the sound sources used, Uniboom™ provided the best data for stratigraphic study in the range of upper Pleistocene sediment thicknesses and water depths found on the New Jersey shelf. In mid-shelf water depths (50-80 m), it typically picked up strong reflectors as much as 40-75 m below the sediment surface, and vertical resolution was approximately 1 m. Clarity and penetration were generally good on the Atlantis II data, but not as good on the Whitefoot and Mount Mitchell cruises. The 3.5 kHz data provide even better resolution (~0.5 m) but has much poorer penetration (10-20 m at best), while the mini-sparker data have better penetration (>100 m) but poorer resolution (~4 m). These data were obtained through analog collection and filtering.

3.2 Core Data

Although there are large numbers of shallow Vibracores from the inner shelf, most of the units described by Sheridan et al. (1993) pinch out farther offshore or are too deeply buried. Thus, to obtain material for age estimates on the older units, the only source of data was the AMCOR core samples (Fig. 4). The USGS took cores from five sites in the study area summarized in Table III.

Poor recovery rates (rarely greater than 30% in the upper Pleistocene) limit the usefulness of the cores for lithostratigraphy. However, cores 6010, 6011 and 6020 lie close to the Atlantis II Leg 2 Line 16, and core 6009 lies near line 18. Sheridan et al. (1993) linked core 6021, which is from the upper slope using air-gun data. Thus, the samples in the cores can be correlated to the correct seismic depositional sequence. If suitable material is available, it can be used to date the sequence.

3.3 Geophysical Logs

The USGS collected geophysical logs for the AMCOR holes. These logs provided clues to the lithologies of sediments drilled through but not successfully recovered, helping to identify the original elevations of cored sediments, and the lithologies present in different sequences at those locations. The logs available for each AMCOR site on the New Jersey margin are summarized in Table III.

Table III. AMCOR Sites in Study Area

Site	Latitude	Longitude	Water Depth (m)	Logs[1]
6009B	38° 51.27' N	73° 35.47' W	59	C,E,FD,G,N,S
6010	39° 03.70' N	73° 05.90' W	76	C,E,FD,G,N,S,V
6011	39° 43.50' N	73° 58.60' W	22	C,E,FD,G,N,S,T
6020	39° 25.41' N	73° 35.63' W	39	None
6021C	38° 57.92' N	72° 49.20' W	298	C,E,FD,G,N,S,T

[1]C = Caliper Log; E = Electric log, FD = Formation Density Log, G = Gamma Ray Log, N = Neutron Log, S = Sonic Log, T = Temperature, V = Sonic Variable Density Log

Source: Hathaway et al., 1976

3.4 Seismic Interpretation Methods

3.4.1 Sequence Stratigraphic Model

There are currently three competing theories of stratigraphic analysis: sequence stratigraphy (Mitchum et al., 1977a; Van Wagoner et al., 1988), the genetic stratigraphic sequence model (Galloway, 1989), and allostratigraphy (Walker, 1992).

Allostratigraphy is difficult to apply without detailed lithofacies data and is, therefore, inappropriate for this study. The other two models were both designed to derive stratigraphic information from stratal geometries, but reflect different views of the controls on stratigraphic architecture.

In sequence stratigraphy, sequence boundaries are defined as surfaces resulting from subaerial and submarine erosional truncation and their correlative conformities (Mitchum et al., 1977a). Within the sequences are parasequences, which are bounded by marine flooding surfaces (Van Wagoner et al., 1988). Parasequences can be aggradational (build upward without landward or seaward shift in facies), progradational (seaward shift in facies) or retrogradational (landward shift in facies). Sequences are broken up into depositional systems tracts (Posamentier and Vail, 1988). The lowstand systems tract (LST) is bounded on the bottom by the sequence boundary and on the top by the transgressive surface, which is a marine flooding surface. The transgressive systems tract (TST) is underlain by the transgressive surface, comprises a set of retrogradational parasequences, and is overlain by the maximum flooding surface. The highstand systems tract (HST) overlies the TST and maximum flooding surface, and is a set of progradational parasequences terminated by the sequence boundary.

The fundamental difference between the Galloway (1989) genetic stratigraphic sequence model and sequence stratigraphy is that different bounding surfaces are used to define the sedimentary sequences. Genetic stratigraphic sequences are separated by marine flooding surfaces rather than erosion surfaces. Unlike the Exxon researchers, who explicitly regard eustatic fluctuations as the dominant control on stratigraphy (Van

Wagoner et al., 1988), a view reflected in the systems tract terminology, genetic stratigraphy follows the view of Frazier (1974) that many stratal surfaces reflect changes in amount and location of sediment delivery to the basin (Galloway, 1989). Although his sequence *boundaries* may retain time-significance, the sequences themselves are bound together by sediment provenance. Sediments in transgressive intervals are formed by reworking of sediments from the underlying "offlap" interval. Galloway (1989) cited several advantages to genetic stratigraphic sequences: the greater ease of obtaining biostratigraphic information from maximum flooding surfaces, the questionable presence of the rapid eustatic falls implied by Type I sequences, and the difficulty of recognizing Type 2 sequences in the distal parts of the basin, and the identification of autocyclic processes.

While there is merit to the Galloway (1989) view, particularly in supply-dominated basins, his objections to sequence stratigraphy seem irrelevant in the late Pleistocene in New Jersey. In view of the overwhelming evidence of rapid eustatic changes during that time, the assumption that stratigraphy will be dominated by eustatic falls is appropriate. Furthermore, the lower rates of sedimentation and subsidence than those observed on the Gulf Coast suggest that autocyclic processes, such as delta switching, should be less important. For these reasons, the more widely recognized and understood sequence stratigraphic terminology of Van Wagoner et al. (1988) is used in this study.

3.4.2 Mapping of Seismic Surfaces

Sequence stratigraphic interpretation relies on the recognition of "surfaces of discontinuity", which are seismic reflections against which other reflections terminate (Mitchum et al., 1977a). Four basic types are recognized: onlap, downlap, toplap and truncation (Fig. 9). Onlap is updip termination of reflections onto a lower bounding surface. Downlap is downdip termination of reflections onto a lower bounding surface. Toplap is termination of reflections against an overlying, non-depositional surface with only minor erosion. Erosional truncation is termination of reflections against an overlying surface where significant erosion has occurred. Truncations can also be structural in origin, in which case they are interpreted as faults rather than unconformities. Mapping of the onlap, downlap, toplap and truncation surfaces on the seismic records allowed definition of sequence boundaries and systems tracts.

Owing to the vertical resolution (approaching 1 m), small valley incisions are readily recognizable in the stratigraphy. As a result, significant incision and erosional truncation is recognizable on shoreface ravinement surfaces associated with marine flooding, as well as on sequence boundaries. These two surfaces were distinguished primarily on the basis of the characteristics of valley fill deposits associated with them and the relative elevation of onlapping reflectors in the underlying and overlying units. Sequence boundaries are characterized by a seaward shift in coastal onlap across the boundary, while ravinement surfaces are characterized by a landward shift in coastal onlap. Furthermore, small to moderately-sized valleys initiated by fluvial incision during a lowstand are characteristically filled with muddy, acoustically transparent sediments (Ashley and Sheridan, 1994). In contrast, tidal inlet incisions associated with shoreface ravinement are typically filled with sand deposits that appear as

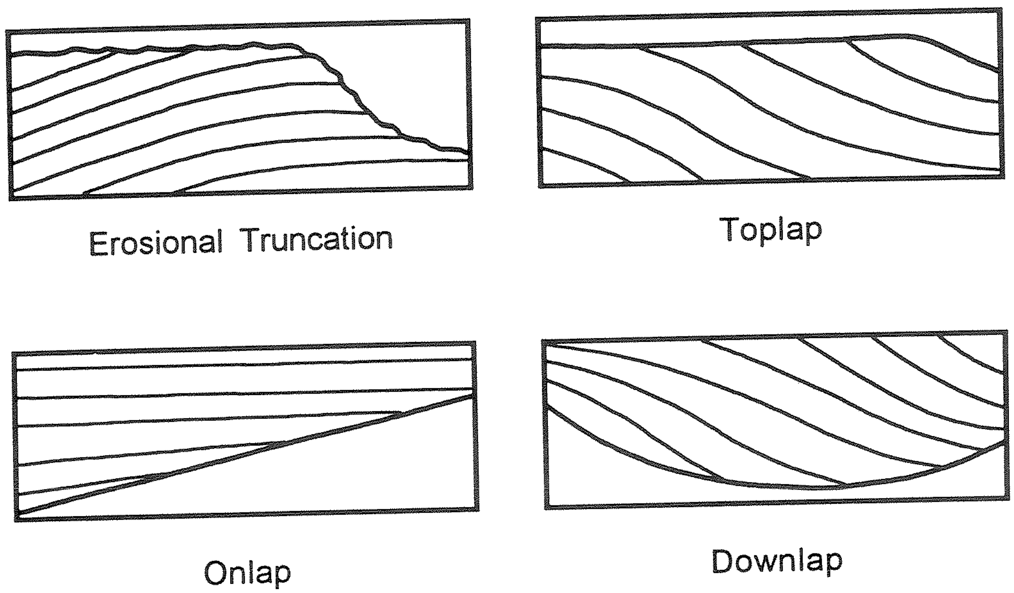


Figure 9. Illustration of Surfaces of Discontinuity Identified on Seismic Records. Modified from Mitchum et al. (1977a).

coincide with marine flooding surfaces. On the outer shelf, sequence boundaries are typically downlapped by a series of clinoforms, which are truncated by an overlying flooding surface. The clinoform deposits are interpreted as the LST, and the overlying flooding surface as the transgressive surface.

The maximum flooding surface, which separates the TST and HST, is identified as a prominent downlap surface separating the onlapping TST from the prograding clinoforms of the overlying HST. Steckler et al. (1993) predicted that on shelves experiencing high rates of eustasy relative to subsidence and sediment supply, the HST would be poorly preserved. In such cases, the marine flooding surface is not recognized and the overlying sequence boundary cuts directly into the TST.

3.4.3 Determination of Sea Level

Three indicators of past sea levels were recognized on the seismic lines: onlapping lowstand wedges, prograding shorelines, and valley-fill deposits. For lowstand wedges, the deepest onlapping reflection horizon in the lowstand wedge was assumed to approximate the lowest sea level. This surface may actually represent fluvial aggradation slightly above sea-level, or marine onlap slightly below sea level in some cases.

The shallowest reflection horizon in the LST, where the wedge is truncated by the transgressive surface, was also useful in estimating sea level. Assuming the truncation surface is a wave-cut ravinement surface resulting from shoreface translation, the depth of erosion during transgression should be roughly 10 m (Kraft et al., 1987; Oertel et al., 1991). Thus, the elevation where the LST is truncated by the

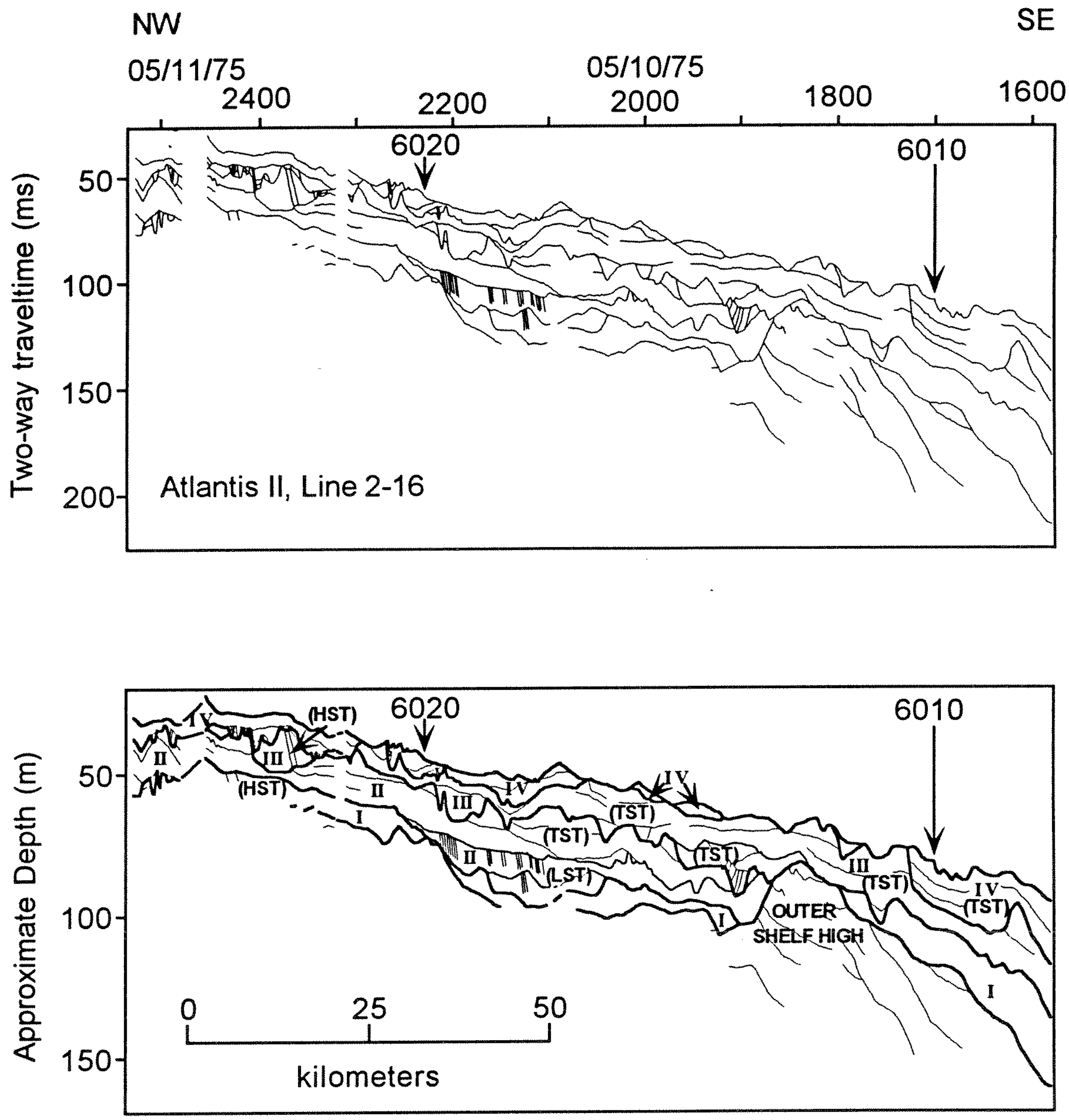


Figure 11. Seismic-stratigraphic Section Along Line 2-16. An uninterpreted line diagram of prominent reflectors seen in the data (above), and the sequence stratigraphic interpretation (below). See Fig. 4 for track line. Depth scale based on velocity of sound in water (1500 m/s). Date and times of data collection are indicated at top.

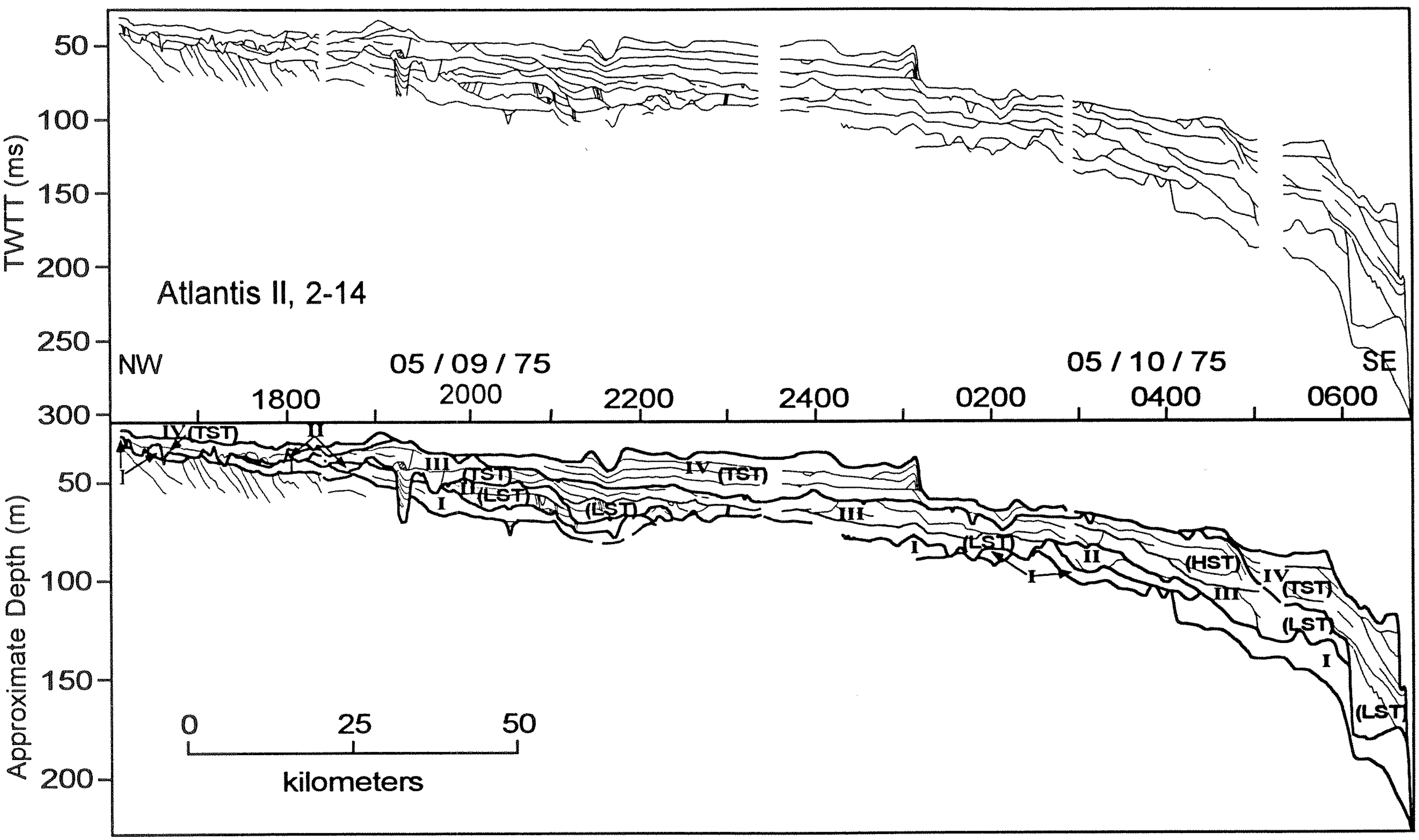


Figure 12. Seismic-stratigraphic Section Along Line 2-14. An uninterpreted line diagram of prominent reflectors seen in the data (above), and the sequence stratigraphic interpretation (below). See Fig. 4 for track line. Depth scale based on velocity of sound in water (1500 m/s). Date and times of data collection are indicated at top.

data from the inner shelf along line 2-14 (Fig. 4). The base of Sequence I erosionally truncates a series of dipping reflections with an apparent seaward gradient of approximately 0.007.

The base of sequence II is a prominent reflection surface that erosionally truncates the sequence I boundary in Figure 15, a section along line 2-16 from the mid-shelf, near AMCOR Site 6020 (Fig. 4). This surface also truncates clinoform reflections within sequence I on the inner shelf along line 2-16 (Fig. 11). Channelling is relatively minor along this surface, but onlapping of internal reflections within sequence II on the mid-shelf, well downdip of clinoforms in the underlying sequence I, provides evidence of significant downward shift of coastal onlap. Sequence II is absent from the outer shelf on both lines.

Numerous valleys are incised into the lower boundary of Sequence III, some of which are shown in Figure 14. The most prominent incision into this surface is the >30 m deep Paleo-Hudson valley described by Knebel et al. (1979), which crosses line 2-14 about 35 km offshore (Fig. 15). Line 2-16, crosses the seaward end of this valley about 75 km offshore (Fig. 16). The prominent incised valleys and widespread erosional truncation of underlying reflections at this surface (Figs. 11, 12) indicate that this surface is a sequence boundary.

The uppermost surface of regional erosional truncation is the base of sequence IV, which largely corresponds to the R-reflection identified by McClennen (1973). This sequence boundary has considerable relief, with channels up to about 10 m deep giving evidence of subaerial erosion (Figs. 11, 12), but is discontinuous, especially on the middle to outer shelf. Figure 17, from the outer shelf along line 2-16, shows a

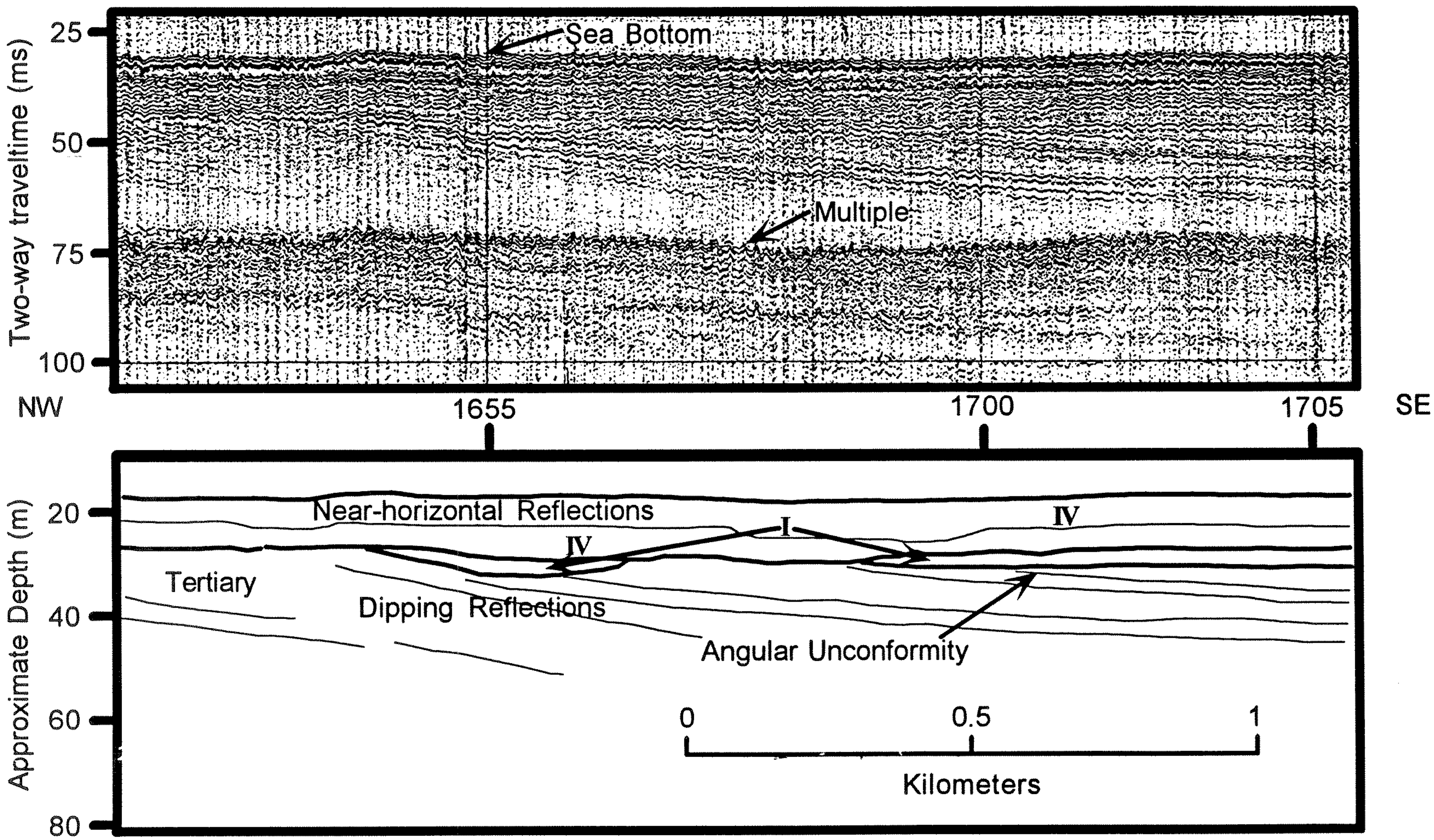


Figure 13. Seismic Section Along Part of Line 2-14. Part of seismic line 2-14 (above) and its sequence stratigraphic interpretation (below). See Fig. 4 for location. Horizontal "scale" indicates times of data collection on May 10, 1975. Roman numerals refer to sequences. Subbottom depths based on a velocity of 1700 m/s.

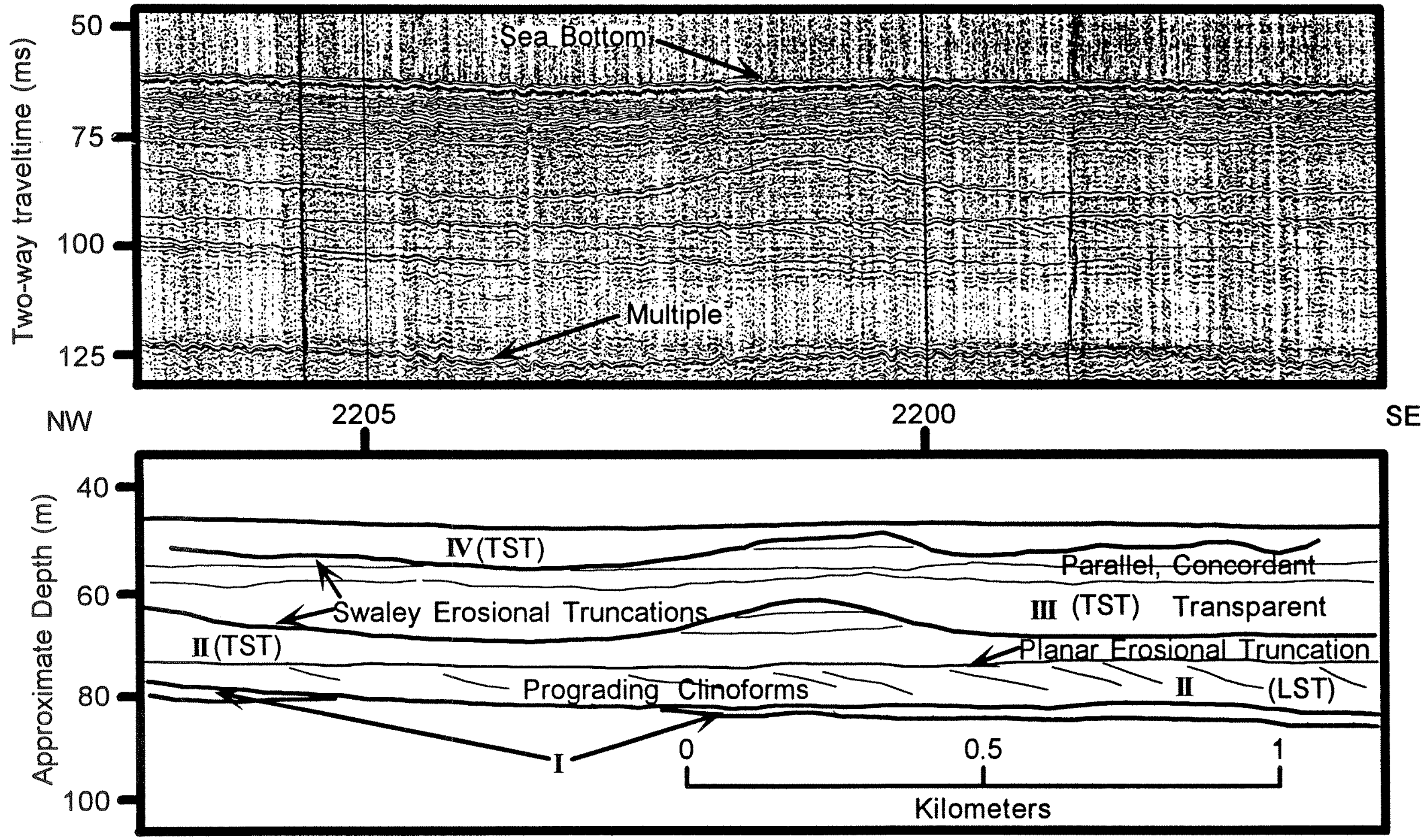


Figure 14. Seismic Section Along Part of Line 2-16. Part of seismic line 2-16 (above) and its sequence stratigraphic interpretation (below). See Fig. 4 for location. Horizontal "scale" indicates times of data collection on May 10, 1975. Roman numerals refer to sequences. Subbottom depths based on a velocity of 1700 m/s.

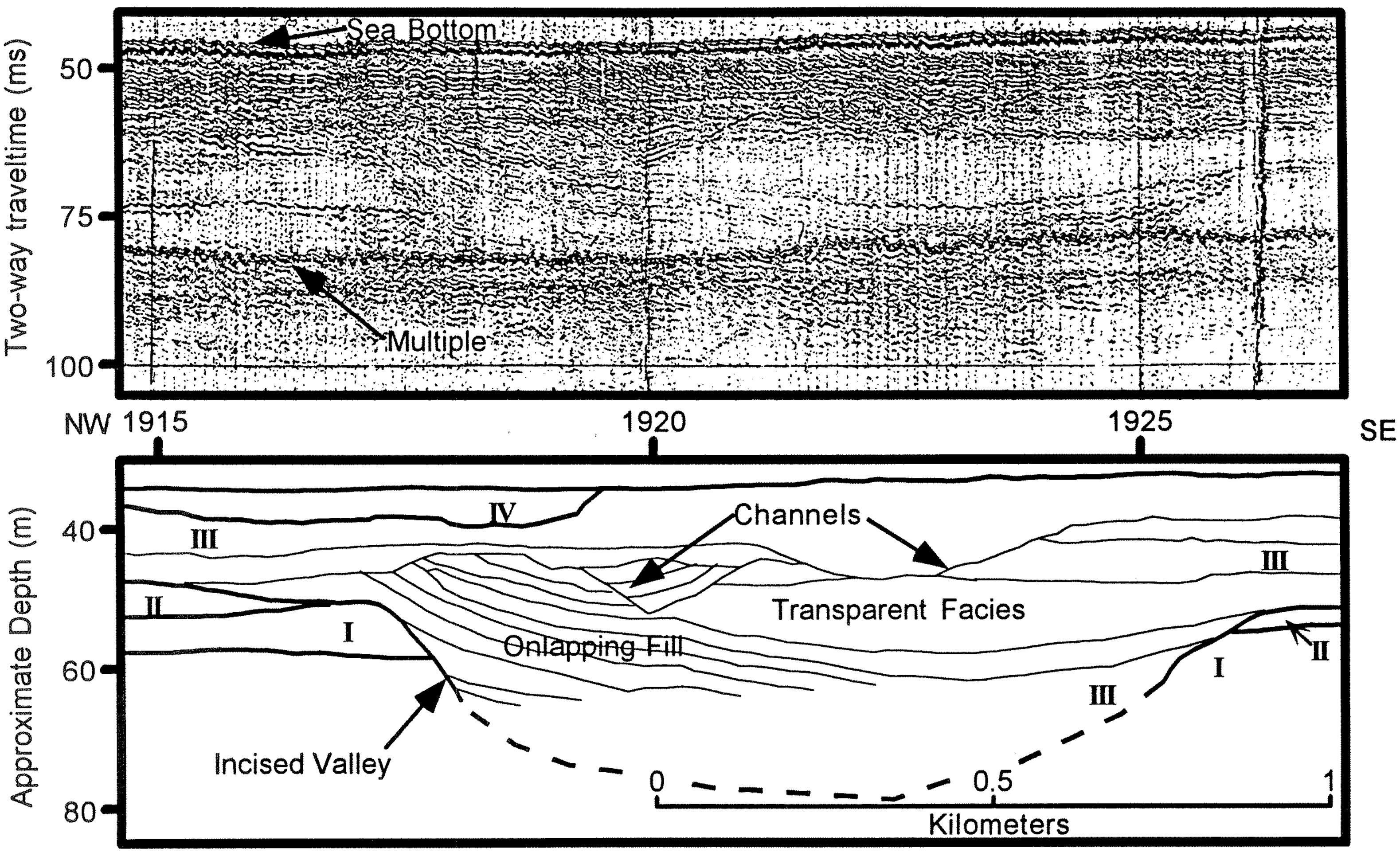


Figure 15. Seismic Section Along Part of Line 2-14. Part of seismic line 2-14 (above) and its sequence stratigraphic interpretation (below). See Fig. 4 for location. Horizontal "scale" indicates times of data collection on May 9, 1975. Roman numerals refer to sequences. Subbottom depths based on a velocity of 1700 m/s.

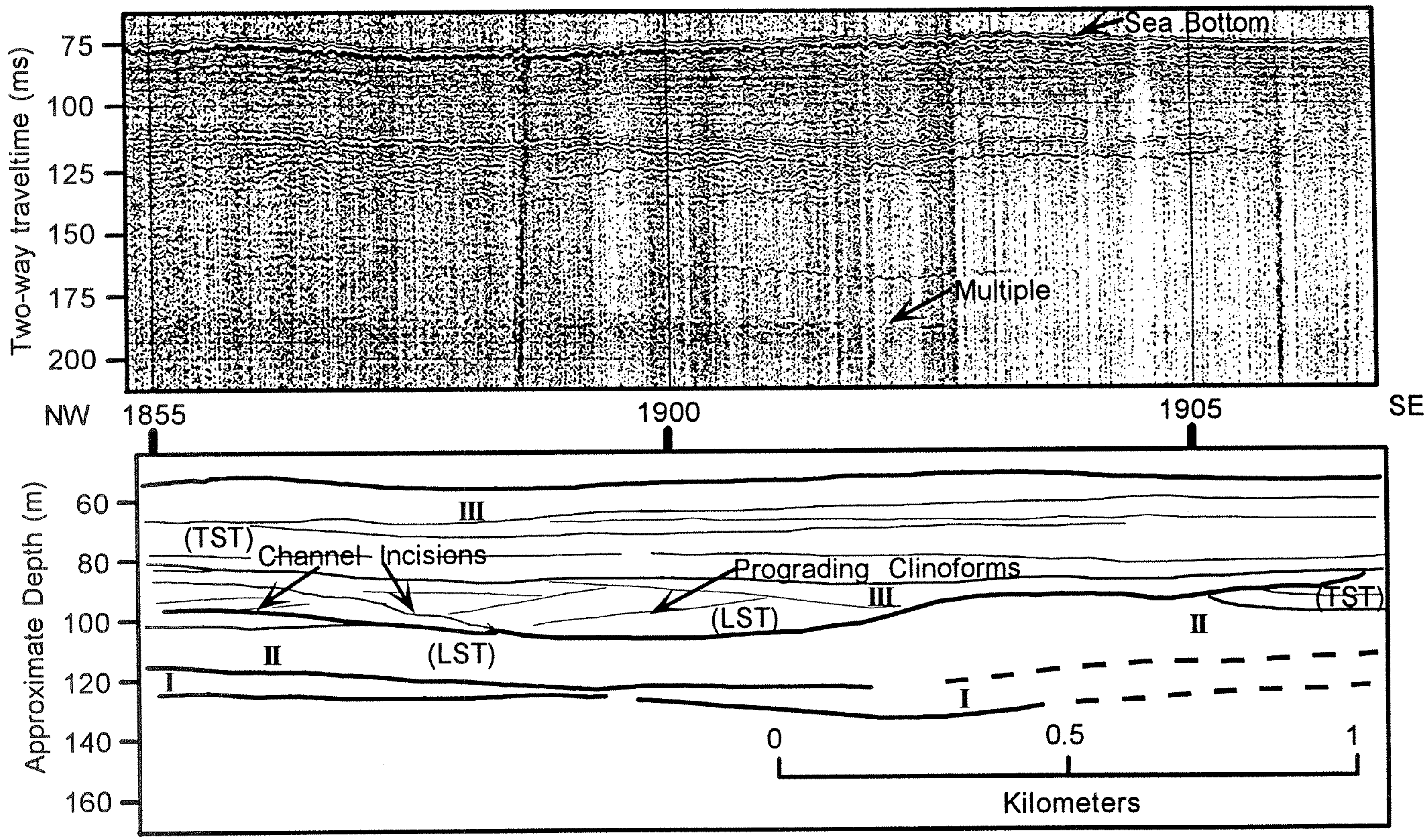


Figure 16. Seismic Section Along Part of Line 2-16. Part of seismic line 2-16 (above) and its sequence stratigraphic interpretation (below). See Fig. 4 for location. Horizontal "scale" indicates times of data collection on May 10, 1975. Roman numerals refer to sequences. Subbottom depths based on a velocity of 1700 m/s.

series of reflections onlapping the sequence boundary, that are below and seaward of clinoforms in the underlying sequence III (Fig. 11). This suggests a downward shift in coastal onlap across the boundary.

4.4.2 Line 2-16

4.4.2.1 Description of reflection surfaces. The base of sequence I emerges from beneath the multiple approximately 40 km offshore and dips gently seaward (approx. ~0.0005) on the inner shelf (Fig. 11). Local relief on this surface, of up to 10 m in this region, reflects significant channelling. Approximately 55 km offshore, it descends into a broad swale approximately 35 m deep relative to its flanks, with an apparent width of 40 km across the mid-shelf. The sequence onlaps a prominent outer-shelf high in the older Pleistocene sediments. Sequence I reappears farther offshore as a seaward- thickening wedge onlapping the seaward flank of this outer shelf high, with a gradient at the base of approximately 0.005.

Internal reflections within sequence I are generally weak, probably due to the loss of energy of the signal as it passes through the overlying sediments. A few descending reflections can be seen on the inner shelf (Fig. 11)

On the inner shelf, the lower boundary of sequence II generally parallels the seafloor at a subbottom depth of 15-20 m and a seaward slope of approximately 0.0005, and is incised by several small valleys (< 1 km wide and 8 m deep). This surface can be traced beneath a highstand shoreline identified by Ashley et al. (1991) 3 km offshore of Barnegat Inlet (Fig. 1) using intersecting seismic lines from the Whitefoot cruise (Fig. 4). On the mid-shelf, it descends into the swale developed in

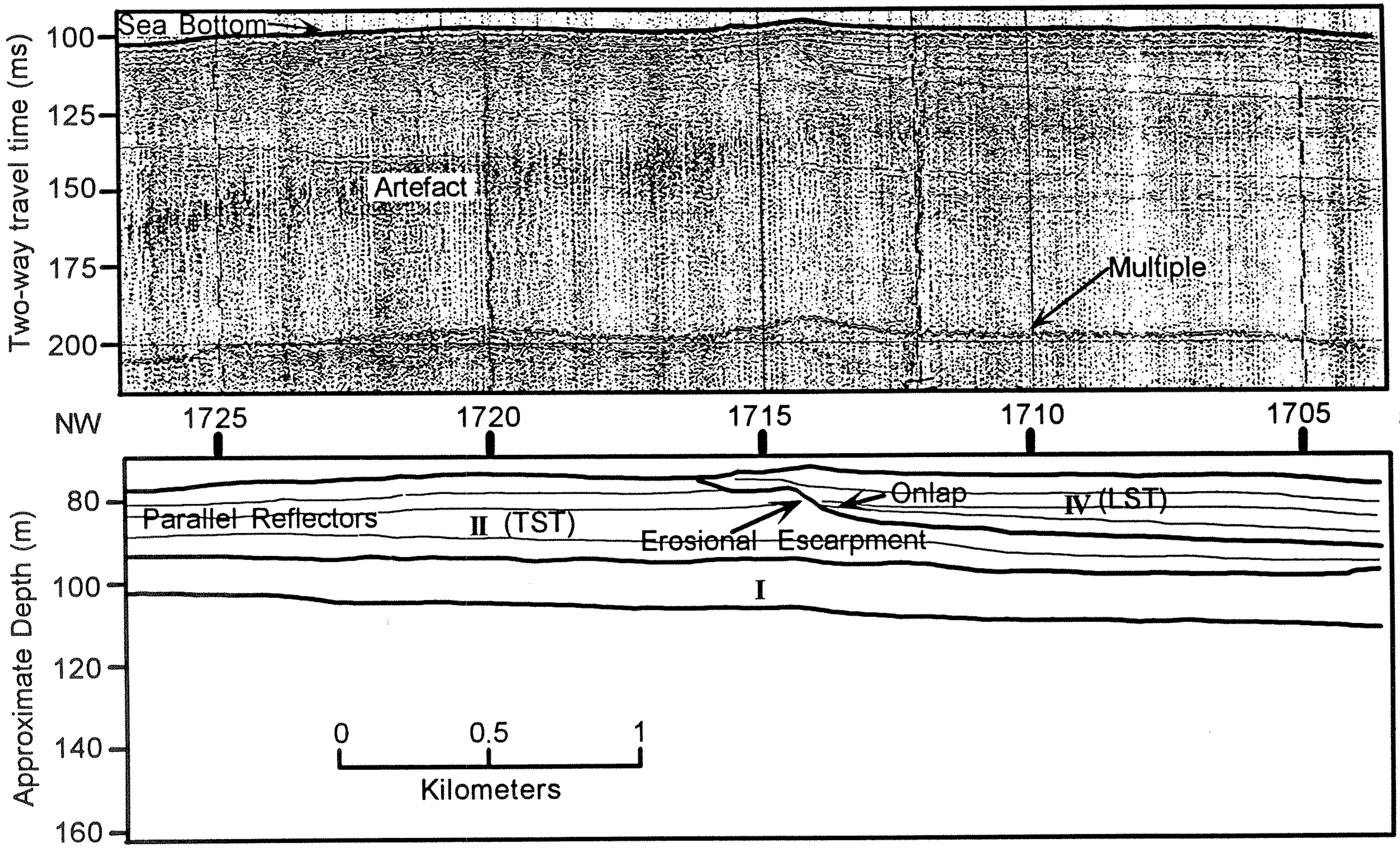


Figure 17. Seismic Section Along Part of Line 2-16. Part of seismic line 2-16 (above) and its sequence stratigraphic interpretation (below). See Fig. 4 for location. Horizontal "scale" indicates times of data collection on May 10, 1975. Roman numerals refer to sequences. Subbottom depths based on a velocity of 1700 m/s.

the underlying sequence I, though the swale is slightly wider and shallower (Fig. 11). The sequence boundary truncates against the outer shelf sediment high, and cannot be recognized reappearing on the outer shelf on line 2-16. However, possibly correlative units appear on the upper slope in some of the profiles of Harris (1983). The swale in the base of the sequence is filled by two distinct packages of prograding, inclined reflections truncated by a largely planar, gently seaward dipping surfaces (Fig. 11). The upper of these two units, and the surface that truncates it are shown in Figure 14. These units have a combined thickness of approximately 15 m. The surface truncating the upper unit is overlain by generally faint but continuous parallel reflections that can be traced onto the inner shelf, where they onlap the underlying sequence boundary.

Sequence III is absent within 30 km of shore and discontinuous throughout the inner shelf (Fig. 11), where it is present only within incised valleys in the base of the sequence, or beneath highs in overlying units. The largest of these valleys is approximately 40 km offshore, more than 10 m deep and 5 km wide in this section. Shallower incisions are seen on the inner shelf in Fig. 14. Seaward of 2300 hrs, the sequence is continuous. The seaward gradient of the sequence boundary on much of the shelf is approximately 0.001, but it increases to about 0.002 beneath the outer shelf. The seaward end of the Knebel's Paleo-Hudson Valley is seen in Figure 16.

The stratigraphically lowest deposits in this sequence are the valley fills. The Paleo-Hudson Valley is composed of two discrete incisions filled by apparently landward-inclined reflectors (Fig. 16). The fill of the shallower incisions shown in Figure 15 is nearly acoustically transparent. A sharp reflection truncates these valleys and is overlain by weak, but continuous parallel reflections that onlap the underlying surface farther onshore.

The base of sequence IV is mostly continuous in the inner shelf, but marked by numerous small incisions; it is truncated by the sea bottom only in a few places (Fig. 11). However, on the mid-shelf, sequence IV is largely absent. The sequence boundary reappears on the outer shelf seaward of a prominent scarp that truncates against the seafloor (Fig. 17). Seaward of the scarp, the sequence boundary has a seaward gradient of approximately 0.0015.

The lowest deposits in this sequence consist of a seaward thickening wedge that onlaps this prominent scarp. This unit is isolated from the inner and middle shelf, where much of this sequence consists of valley fills with a variety of fill patterns, as described by Ashley and Sheridan (1994). These valley fills are truncated by nearly planar surfaces that parallel the seafloor and outcrop in swales in the sea bottom, particularly between sand ridges (Fig. 11).

4.4.2.2 Interpretation of Systems Tracts. The paucity of internal reflections in sequence I makes it difficult to distinguish the systems tracts, but the descending reflections on the inner shelf could be the downlapping clinoforms of a prograding highstand unit (Fig. 11). The seaward-thickening deposit on the outer shelf may represent an onlapping lowstand wedge.

As there is little channelling and no sign of significant downward shift in coastal onlap along the planar surfaces truncating the progradational packages in sequence II (Fig. 11), they are interpreted as flooding surfaces separating parasequences within the LST rather than as sequence boundaries. The significant landward shift in onlapping reflectors above the upper parasequence suggests a transgressive surface, implying that the overlying, largely concordant reflectors belong to the TST.

Because the Paleo-Hudson valley is oblique to the seismic section, the apparently landward-inclined reflections in the Paleohudson valley could represent filling of a baymouth by southward longshore drift. Ashley and Sheridan (1994) found that multiple incision and fill by spit progradation are common features of large incised valleys. Alternatively, the clinoforms could represent a flood tidal delta prograding into an estuary during transgression. The acoustically transparent fill of the small valleys is probably muddy sediment; the seismic facie resembles those of small, fluvial valleys on the Atlantic coastal plain that filled with mud during the late Pleistocene-Holocene transgression. The considerable apparent width of these valleys suggests that the line may be cutting obliquely across the same sinuous valley. The landward shift in coastal onlap above the truncating reflection suggests that it is a ravinement surface, resulting from subsequent transgression.

The onlapping wedge on the outer shelf in sequence IV is the "outer shelf sediment wedge" described by Milliman et al. (1990) and Davies et al. (1992). Cores from the outer shelf wedge indicate that most of the wedge was deposited in inner to middle shelf water depths (Davies et al., 1992). This suggests that only the basal sand may have been deposited in the lowstand, while the remainder of the wedge is transgressive in origin. The seafloor in this region and the outer shelf is a transgressive surface. Above the sequence boundary, where it reappears nearer shore, is the thin transgressive shelf blanket described by Swift et al. (1972). The landward increase in continuity of sequence IV sediments could reflect the slowing of the rate of sea-level rise during the Holocene. The planar surfaces truncating the valley fills likely represent wave-cut ravinement surfaces.

4.4.3 Line 2-14

4.4.3.1 Description of Reflection Surfaces. The base of sequence I is a strong continuous reflection demarcating the sharp angular unconformity shown in Fig. 13. Apparent dips of the underlying units are as high as 0.014; the sequence boundary slopes seaward with a gradient of about 0.0007, similar to the modern seafloor. Sequence I is discontinuous on the inner shelf because of incisions in the overlying sequence IV boundary. About 25 km from the start of the line, the sequence boundary gradient increases to about 0.002 and the sequence thickens to as much as 10 m. Though largely smooth, there is one narrow, prominent incised valley (~ 0.5 km apparent width, 8 m deep) (Fig. 12). The sequence boundary disappears below the multiple on the mid-shelf, making it difficult to recognize the seaward flank of the swale. The sequence reemerges on the outer shelf, where it first reappears as thin and discontinuous, but thickens seaward. The sequence boundary gradient steepens to more than 0.02 near the shelf break, as the line passes into South Tom's Canyon (Figs. 1, 4). Sequence I outcrops on the upper slope within the canyon at water depths of 190 m to 230 m. As on line 2-16, few internal reflections can be seen in this unit; those which are present parallel its base. Elsewhere on the outer shelf, Sequence II is eroded and not preserved.

Sequence II is absent within 25 km of the shoreline and discontinuous on the inner shelf, where it is present only within small valleys or beneath highs in the overlying sequence boundaries (Fig. 13). On the mid-shelf, it dips more steeply seaward (gradient ~0.004 into a mid-shelf swale as on line 2-16 (Fig. 12) but is truncated by an overlying reflection surface at about 2200 hrs. On the outer shelf, there is local evidence of incision between the sequence I boundary and an overlying reflection displaying erosional truncation; this valley fill and adjacent deposits are tentatively included in sequence II.

The stratigraphic lowest part of the sequence is contained within the mid-shelf swale, where two packages, each capped by a largely planar, seaward-dipping surface, onlap the sequence boundary (Figure 11). The upper of these two packages has well-developed higher-angle reflections that downlap the lower planar surface and truncate against the upper one; a few similar reflections are seen in the lower package. Above the upper planar surface, there are faint, parallel reflections similar to those observed in the upper part of sequence II on line 2-16.

Sequence III is absent within 30 km of the shoreline on line 2-14, but thickens seaward into the Paleohudson Valley identified by Knebel et al. (1979) (Figs. 12,15). Although the deepest part of this valley is only 1-2 km wide, the entire swale in the basal sequence III boundary extends for 10 km. On the mid-shelf, the sequence boundary slopes very gently seaward with a gradient of about 0.0002 except for a 15 m scarp. On the outer shelf, the seaward gradient steepens to 0.003, and the sequence boundary has considerable relief, until it is truncated by a prominent steeply plunging reflection near the shelf edge (Fig. 12).

Seaward of the 15 m mid-shelf scarp in the underlying sequence boundary are two packages of progradational clinoforms with a combined thickness of approximately 10 m separated by a planar surface similar to that seen on line 2-16 in sequence II, and shown in Figure 14.

The prominent valley shown in Figure 15 has a complex fill pattern, that can be broken into three components. The lowest unit contains a series of strong continuous reflections that dip gently seaward and onlapping the valley wall. Overlying this unit is one that is nearly acoustically transparent. The upper part of the valley fill consists of strong, "noisy" reflections with multiple incisions. Elsewhere the upper part of the sequence contains reflections that are largely parallel and concordant, similar to those seen on line 2-16. On the outer shelf, the uppermost of these parallel reflections is downlapped by a series of steeply dipping reflections that characterize a seaward-thickening wedge (Fig. 12).

The base of sequence IV displays a similar character to the more southerly line across the inner shelf, with numerous small incisions. It truncates against the seafloor above the large valley in sequence III (Fig. 12). Farther offshore, however, the sequence boundary reappears and descends with a seaward slope of approximately 0.001 to approximately 2300, then flattens out and outcrops on the seafloor on the outer shelf. It reappears near the shelf-edge, where it descends in a series of steep, curved steps with gradients of up to 0.03 (Fig. 12). The sequence boundary outcrops on the wall of South Tom's Canyon at a depth of approximately 190 m.

On the outer shelf, the lower part of sequence IV along line 2-14 can be divided into an acoustically chaotic deposit at the base in the curve of the steepest "step" in the underlying sequence boundary (Fig. 12). This unit is downlapped by a series of steeply dipping reflections that abruptly terminate against the canyon wall. This unit comprises most of the Outer Shelf Sediment Wedge on line 2-14. Sequence IV on the inner shelf along this line closely resembles line 2-16, but the mid-shelf is dramatically different. Situated above a largely horizontal reflection surface on the mid-shelf is a seaward-thickening wedge of sediment up to 20 m thick. Internal reflections in this wedge largely parallel the bottom, except at the extreme seaward end, where they downlap an underlying surface.

4.4.3.2 Interpretation of Systems Tracts. Little internal detail can be seen within sequence I, but the thickening on the outermost shelf resulting from the increased gradient of the basal reflector probably corresponds to a lowstand wedge. The steeply dipping inclinations of the reflections below the sequence boundary on the inner shelf suggest these may be Tertiary units that subcrop near the seafloor landward of the hingeline, where little or no long-term subsidence is occurring.

The two packages onlapping the base of sequence II within the mid-shelf swale may correlate to the two lowstand parasequences observed on line 2-16. Above these packages, the largely concordant onlapping reflectors are similar to those seen on line 2-16, and are also interpreted as transgressive. The deposition of the lowstand wedge was apparently limited to the mid-shelf; porgradation ceased in this area before reaching the outer shelf, apparently, accounting for the absence of Sequence II there.

The two packages seaward of the scarp in the lower boundary sequence III are similar to the features seen in the mid-shelf swale on line 2-16. This suggests that they may also be lowstand deposits truncated by a transgressive surface. The presence of shell fragments in the uppermost part of the valley fill (Knebel et al., 1979) suggests that the valley was transgressed as it filled. The three distinct seismic facies within the valley could correspond to the succession predicted for a transgressive valley fill deposit by Dalrymple et al. (1992) and Ashley and Sheridan (1994). The well-stratified lower unit may represent progradation of a bay-head delta complex, overlain by acoustically transparent muds from the estuarine basin, and capped by a "noisy" surface resulting from multiple tidal current incisions during transgressed of the baymouth. The downlap surface on the outer shelf could represent a maximum flooding surface, implying that the downlapping clinoforms represent a prograding highstand wedge.

The acoustically homogenous deposit at the base of the steep, curving scarp in the lower boundary of sequence IV on the outer shelf could represent a slump deposit. Overlying reflections are steep, and suggestive of progradation during a lowstand. These deposits are truncated by a thin, discontinuous blanket similar to the TST on line 2-16. However, within the mid-shelf there is a thick, prograding sediment wedge that has been described in more detail by Milliman et al. (1990). In view of the subsequent truncation of this deposit by the seafloor, this is interpreted as a parasequence within the TST.

4.4.4 Age Constraints on Depositional Sequences

Analysis of the seismic data revealed four depositional sequences, which we have numbered I to IV. Deeper units were visible on many lines, but could not be correlated over significant distances using the Uniboom™ data. Though absolute age estimates are scanty, some limits can be placed on the age of the sequences from information from AMCOR wells 6010 and 6020, which lie within 1 km of line 2-16.

The base of sequence I lies above the first appearance datum of the calcareous nannofossil *Emiliania huxleyi* in AMCOR 6010 (Harris, 1983), which first appeared during stage 8, ~275 ka (Thierstein et al., 1977; Berggren et al., 1980). Furthermore, a shell from just above the boundary in core 6020 gave an amino acid ratio of 0.21, which is consistent with a stage 5 age (Fig. 2), if the shell racemizes at a rate comparable to *Mercenaria* (Groot et al., 1995). This suggests that the sequence I boundary may have been cut during the stage 6 lowstand (Fig. 2).

The base of Sequence IV corresponds to the R-reflection of McClennen (1973). The lower boundary of this sequence is reasonably well-constrained by radiocarbon dates, and is believed to correspond to the stage 2 lowstand.

If sequences I and IV correspond to the last two major glacial / interglacial cycles as suggested (stages 5/6 and 1/2 respectively) (Fig. 2), Sequences II and III must, therefore, correspond to the smaller-scale eustatic events, such as stages 4, 5.2 and 5.4, controlled by the obliquity and precession cycles between stages 2 and 6. Thus, the stratigraphy of the New Jersey Shelf records several smaller-scale fluctuations, along with the major glacial / interglacial cycles.

4.5 Discussion

The overall pattern of the stratigraphy of the two lines is similar. Both sections show a broad mid-shelf swale in the sequence I lower boundary; this swale trends obliquely to the shoreline so that it is 55-95 km offshore on line 2-16, but only 35-60 km offshore on line 2-14. The swale widens and deepens as it trends southward, and was the most likely a result of migration of the Hudson River as it flowed across the shelf during a low sea-level stand. The swale remained as a broad shelf-valley that was filled by prograding sequence II LSW sediments. Along Line 2-14, a second valley was incised into the lower boundary of sequence III, eroding the outer part of this unit. The Hudson had apparently changed position to a deeply incised valley nearer shore, that did not extend as far as line 2-16. Sequence III sediments overtopped the relative high on the outer shelf, and the Hudson shifted to the modern Hudson shelf valley prior to the deposition of Sequence IV, so the Paleohudson Valley was not reincised. The timing of this change of course suggests that it may have resulted from the weight of the advancing Laurentide ice sheet depressing the crust in front, and creating a bulge farther south. The two distinct positions of the Hudson may partly reflect underlying structural control, as the three salt intrusions identified by Grow et al. (1988) lie between the modern Hudson Shelf Valley and the transect of the Paleohudson Valley (Fig. 1).

Another feature of the stratigraphy peculiar to the region is the sharp angular unconformity on the inner shelf on Line 2-14. The dips of the underlying units (up to 0.014) suggest that they are of at least Pliocene age. This unconformity arises because the hinge line of the Baltimore Canyon Trough veers offshore parallel to Long Island off northern New Jersey (Fig. 1). Flexure about this hinge by subsidence in the Baltimore Canyon Trough may result in slight uplift landward of the hinge. Hence this region was subject to extensive erosion, and Cretaceous through Tertiary strata are exposed (Allen-Lafayette, 1996).

Many of the units are fragmentary, particularly on the inner shelf where preservation is often restricted to incised valley fills. The sequences were typically less than 10 m thick on much of the shelf. If thin sequences such as these were deeply buried, the regional airgun surveys typically used for sequence stratigraphic analysis would be hardpressed to resolve them, and might underestimate the number of sea-level cycles recorded. In fact, at lower resolution, the repeated channel incisions might result in chaotic reflections. Though it seems likely that sea-level changes of largest magnitude would be better preserved regionally, this need not be true locally. For example, large rivers may incise 20-30 m into the shelf. On a slowly subsiding shelf, the accommodation space created within these valleys could equal that created by subsidence over several 100 ka. Thus, the sequences preserved in a particular area may be controlled as much by its position relative to major rivers at the time of valley incision as by the scale of sea-level fluctuation.

In view of the fragmentary stratigraphy and strong influence of fluvial erosion, attempting inter-basin correlations in times of high frequency eustasy, such as the three third-order Pleistocene cycles shown on the Haq et al. (1987) curve, could be a risky proposition on shelves with relatively low rates of subsidence and sediment supply. Thorne and Swift (1991) predicted that high-amplitude, high-frequency eustasy superimposed on a lower-frequency eustatic curve would result in a complex stratigraphy that would be difficult to unravel; this appears to be correct. Both regional coverage and dense seismic spacing would likely be necessary.

Steckler et al. (1993) predicted that the stratigraphy of margins experiencing large eustatic fluctuations should be dominated by type 1 sequence boundaries. This appears to be true for the New Jersey margin; all surfaces across which a significant seaward shift of coastal onlap occurred were prominently incised and associated with extensive truncation beyond the underlying clinoform inflection point. These characteristics suggest widespread subaerial erosion.

I present a schematic model sequence formed during major interglacial/glacial cycles derived primarily from the most recent sequence, sequence IV (Fig. 18). The internal stratigraphy of Sequence I was difficult to resolve, but the gross characteristics of the sequence were similar: it also consisted of a fairly thin (5-10 m thick) , discontinuous sheet over most of the shelf, thickening to wedge near the shelf edge. This pattern is consistent with the prediction of Steckler et al. (1993) that the LST and TST would dominate margins where rates of eustasy are much greater than rates of sediment supply and subsidence. No shelf-margin systems tracts were identified in any sequence, and highstand units were poorly preserved. While this pattern is similar to that described from the Rhône Delta by Tesson et al. (1990), it differs from the shelf-edge deposits from the Gulf of Mexico, which are dominated by thick, late highstand deposits (Sydow and Roberts, 1994). This suggests that the HST may be well-developed under conditions of high-amplitude, high-frequency eustasy, only if sediment supply and subsidence are particularly high. Evidence from ancient sequences formed during periods of substantial continental glaciation also suggests decreased sequence thickness and diminished importance of the HST on slowly subsiding margins. For example, Pekar and Miller (in press) report that shelf sequences formed during the early Oligocene, the time of the largest sea-level fluctuations of the Paleogene, are thin and dominated by transgressive deposits. In contrast, sequences formed during the smaller sea-level fluctuations of the Eocene and late Oligocene, are thicker and have well-preserved highstand deposits.

The TST is generally described as typified by retrogradationally stacked

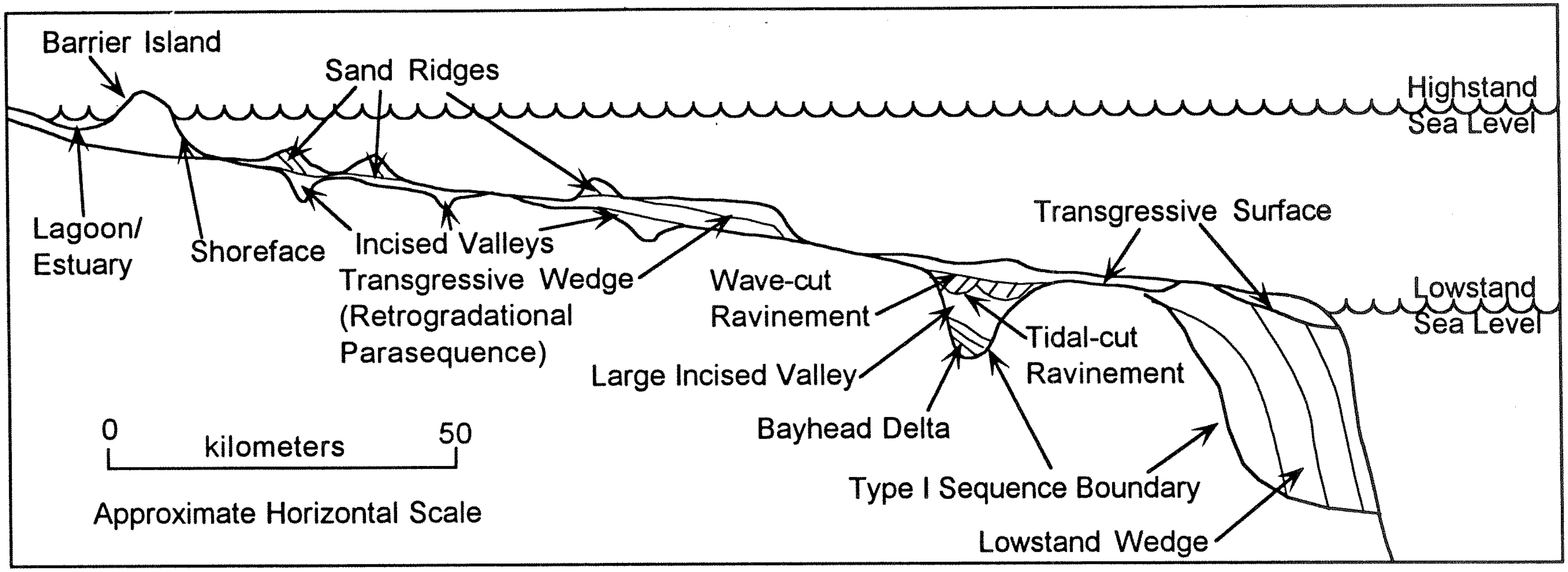


Figure 18. Schematic model of depositional sequence formed during a major glacial / interglacial cycle.

parasequences (Mitchum et al., 1977b; Van Wagoner et al., 1988), but this did not hold true for the study area. In most cases, the TST was restricted to incised valley fills and a thin, discontinuous blanket of sands reworked by shallow marine processesinto ridges and swales (Fig. 18), because of sediment trapping within coastal estuaries (Swift et al., 1972; Clarke et al., 1983). Backstepping parasequences may exist within incised valleys, or as isolated bodies on the shelf, but generally not as stacked sets of regional extent (Figs. 12, 18). Erosional stripping and reworking of sediments into sand ridges during transgression, and continuing marine erosion during highstands, result in discontinuous sediment cover above the ravinement surface. The combination of rapid sea-level rise and low sediment input prevent progradation back to the position of the previous parasequence. In contrast, the stage 2 - stage 1 transition in the Mississippi delta is characterized by backstepping parasequences (Suter et al., 1987), suggesting that their development requires a higher ratio of sediment supply to sea-level rise than has generally been experienced on the New Jersey shelf during the Quaternary.

An important exception is the mid-shelf sediment wedge, which has been attributed to a pulse of sediment supply from the Hudson associated with deglaciation (Milliman et al., 1990). The Hudson River currently supplies approximately 870,000 metric tons of sediment per year to the lower Hudson (Ellsworth, 1986). At this rate of sediment supply, it would take approximately 50,000 years to produce a 10 m thick wedge of sediment covering 2500 km^2. Thus, the mid-shelf sediment wedge could not have been created by modern sedimentation rates; it must have resulted from higher rates of sedimentation during the late Pleistocene. Radiocarbon dates place the age of the Fortune shoreline, which is associated with this sediment wedge at approximately 11.5 ka (Dillon and Oldale, 1978). Fairbanks (1989) reports evidence from Bermuda reefs of a slowing of the rate of sea-level rise between 12 ka and 10 ka corresponding to the younger Dryas event. This slowing of sea-level rise also likely contributed to the development of the mid-shelf sediment wedge.

Figure 18 illustrates the importance of incised valley fills to shelf sequences. Incised valley fill sequences are commonly attributed to the LST in the Exxon model (e.g. Posamentier and Vail, 1988). However, as noted by Ashley and Sheridan (1994), fluvial deposition is generally a minor component of incised valleys on the U.S. Atlantic margin. This likely reflects the low rate of sediment input relative to sea-level rise that tends to result in transgression rather than aggradation or progradation once sea level rise begins again. In general, the stratigraphy of the large valleys on the New Jersey shelf was consistent with fill occurring mostly during transgression. Incised valley fills in the Gulf Coast generally have more fluvial material infilling them during transgression (Suter et al., 1987), another effect of the higher sediment input in that region.

Sequences II and III, which developed under stadial / interstadial sea-level fluctuations, display slightly different characteristics from each other, as well as from the glacial / interglacial model shown in Figure 18. Sequence II is primarily preserved within a fluvial plain on the mid-shelf, and is largely absent from the outer shelf. In contrast, sequence III, while also thickening prominently in the mid-shelf region, includes a thick highstand unit (Fig. 12). The preservation pattern of sequence II differs from sedimentary sequences generated in numerical models, which typically have a layer-cake structure with sequences thickening offshore to the shelf-break, and sequence preservation improving with distance from shore (e.g. Jervey, 1988). The preservation of sequence II on the mid-shelf may partly reflects the low subsidence rate; the amount of accommodation space created by fluvial erosion during lowstand can exceed that created by subsidence on the outer shelf.

Other factors, such as shoreline position and sediment transport processes, can also favor better stratigraphic preservation of stadial / interstadial sequences on the mid-shelf. In a margin with a relatively low sediment supply, progradation does not approach the shelf edge during these intermediate stands of sea level. As a result, the outer shelf can experience net erosion even during highstands. Vincent et al. (1981) reported currents capable of significant sediment transport at depths of 91 m. This suggests that continuing marine erosion during highstands may be a stratigraphically significant process that limits the preservation of stadial / interstadial sequences on the outer shelf. The preservation of the HST of III on the outer shelf suggests that sediment supply must have been greater, possibly reflecting the advance of ice into the Hudson River drainage as sea-level fell during stage 2.

4.6 Conclusions

In view of the unique position of the New Jersey continental margin at the edge of the Laurentide ice sheet, and the limited extent of strike-oriented seismic lines, caution is advisable in applying the results of this study to other areas. Nevertheless, several characteristics of the stratigraphy seem likely to be typical of continental shelves experiencing high frequency, high amplitude eustasy in combination with low subsidence and sediment supply. Large-scale continental glaciations capable of producing such eustatic conditions occurred not only during the Pleistocene, but also during the late Tertiary, the Permo-Carboniferous, and late Ordovician and late Precambrian (Frakes, 1979).

Short time intervals (100 ka or less) between lowstands, low sediment supply, intensive fluvial incision and marine erosion during transgressive and highstand periods combine to produce thin and fragmented sedimentary sequences that are difficult to correlate. Preservation of sequences does not necessarily improve seaward, but can be influenced by local factors such as shifting fluvial valley positions, and differential tectonism related to hinge lines, salt movement and glacial isostasy.

Type 1 sequence boundaries predominate, and most sediment is deposited in transgressive or lowstand systems tracts. During transgressive and highstand intervals, such as stage 5, sediments may be trapped in estuaries, starving the offshore region of sediment. As a result, much of the TST is deposited in incised valley fills; sediments above the ravinement surface are discontinuous, and parasequences tend to be isolated rather than stacked on top of one another. The outer shelf is subject to long intervals of marine erosion and the record from transgressive and highstand times may be poorly preserved. Preservation of highstand deposits is generally dependent on increased sediment supply and subsidence, which in New Jersey is related to glacial fluctuations. However, similar effects could be produced by climatic changes or drainage changes on continental margins that are not ice-marginal.

CHAPTER FIVE:
PARAGLACIAL INFLUENCE ON LATE QUATERNARY STRATIGRAPHY OF THE NEW JERSEY SHELF: A COMPARISON OF THE EFFECTS OF GLACIATION AND TECTONICS

5.1 Introduction

The stratigraphy of sedimentary basins is commonly used to evaluate regional tectonic settings. Examination of the structure of the basin and the rates and patterns of sediment supply and subsidence yields information about where the basin lay relative to plate margins and the types of plate interaction occurring during deposition. These data are used by basin analysts to classify the basin, and interpret the regional geologic history.

Surficial, or "exogenic" processes (Ritter, 1986), such as glaciation, can also produce changes in sediment supply and subsidence that may appear to mimic the effects of deeper crustal tectonic processes. The weight of a large continental ice sheet can depress the crust beneath it by several hundred meters and produce an uplifted "peripheral bulge" around its margins. These effects could be confused with tectonic uplift and subsidence associated with plate movement or thermal subsidence. Glaciers can also increase sediment supply to an area, both from sediment transported directly from the glacier by meltwater during deglaciation, and from glacial and proglacial sediment eroded and transported by rivers. Furthermore, glacial-isostasy and the blockage of existing rivers by advancing ice can result in significant drainage pattern changes that may affect the sediment flux. Increased sediment loads associated with glaciation could be mistaken for the result of tectonic uplift of the source region.

Church and Ryder (1972) used the term "paraglacial" to refer to non-glacial processes that directly result from the actions of glaciers. These processes can have significant stratigraphic consequences. The New Jersey shelf is a region where such "paraglacial" processes should have had significant impact. Endogenic tectonic effects (those driven by deep crustal processes; Ritter, 1986) should be small because of its location on a relatively old, stable passive margin, allowing recognition of paraglacial effects. During the Late Wisconsin (stage 2), the Laurentide Ice Sheet advanced as far south as northern New Jersey (Dyke and Prest, 1986) (Fig. 1). Furthermore, the Hudson River, which has been the primary sediment source for the New Jersey shelf throughout the Pleistocene (Poag and Sevon, 1989), drained much of the eastern margin of the Laurentide Ice Sheet. Thus, continental glaciation must have impacted on the shelf through sediment supply changes and isostatic adjustment.

The objectives of this study are to use the extensive coverage of existing seismic data and the limited core data collected by the USGS to construct a sequence stratigraphy for the upper Quaternary section of the New Jersey continental shelf. The resulting stratigraphy will be used as an example of a "paraglacial" continental margin, and to examine the relationship of sediment supply changes on the shelf to events on the adjoining continent. The observed relationship between glaciation and shelf stratigraphy allows an assessment of criteria for distinguishing the effects of tectonic processes from those resulting from paraglacial processes.

5.2 Geologic Setting

The study area is the NE-SW trending segment of the U.S. Atlantic Coast continental shelf between the Hudson Shelf Valley and the Delaware Shelf Valley from

38°40' to 40° 30'N and 72°30' to 74°40'W (Fig. 1), an area of approximately 25,000 km^2. The continental shelf of New Jersey is broad (120-150 km) and gently sloping (< 0.001). The steepening associated with the shelf break begins between the 120 and 160 m isobaths. The shelf is generally classified as storm dominated, and has a mixed energy shoreline, with tidal range of 1-2 m and mean significant wave heights of roughly 1 m.

The region is part of the Baltimore Canyon Trough, a passive margin with a hinge line running essentially parallel to the shoreline approximately 20 km offshore south of Barnegat Inlet (Fig. 1; Grow et al., 1988). Farther north, the hingeline trends parallel to Long Island (Fig. 1). The rate of thermal subsidence has averaged less than 0.01 mm/a over the past 80 Ma in the center of the trough (Greenlee et al., 1988). However, significantly higher rates of total subsidence have sometimes occurred due to sedimentary loading. For example, under the influence of increased sediment input during the middle Miocene, which has been attributed to tectonic uplift of the central Appalachians (Mathews, 1975 Hack, 1982; Poag and Sevon, 1989), total subsidence rates exceeded 0.05 mm/a (Greenlee et al., 1988).

The Hudson River has been the primary sediment source for the area throughout the Quaternary (Poag and Sevon, 1989). Current sediment supply to the shelf is minimal, because the sediment is trapped in lagoons and estuaries (Swift et al., 1972; Clarke et al., 1983). However, the average rate of sediment accumulation on the margin has been even higher during the Quaternary than during the middle Miocene, because of glaciation (Poag and Sevon, 1989).

During isotope stage 2 (Fig. 2), the Laurentide ice sheet advanced as far south as northern New Jersey (Fig. 1; Dyke and Prest, 1986). The weight of the ice sheet caused substantial isostatic adjustment of the crust in the region. Estimates of the extent of isostatic adjustment from geophysical models appear to conflict with observed relative sea-level data from the mid-Atlantic. Existing isostatic models have generally predicted significant isostatic adjustment as far south as North Carolina (Clark et al., 1978; Tushingham and Peltier, 1992), but evidence from the Atlantic Coast suggests a more geographically limited effect. Dillon and Oldale (1978) demonstrated that Late Wisconsinan shorelines preserved on the shelf were essentially horizontal from North Carolina to an inflection zone running southeastward from Barnegat Inlet, New Jersey (Fig. 7). They suggested that the areas north and south of that hinge zone were on different crustal blocks that responded largely independently to ice loading. Ice loading resulted in relative subsidence of the northern half of the study area since 15 ka (Dillon and Oldale, 1978, Fig. 7). Peltier (1994) unveiled a new model in which the Laurentide Ice Sheet is 35% thinner than in his previous models, which may address this problem, but a new relative sea-level curve for sites in the region has not yet been published.

5.3 Data Sources and Methods

The seismic data used were collected on several USGS cruises between 1974 and 1982. The principal cruise data used were the Atlantis II Leg 2 lines 12-14, 16-18, and 20-22 that comprise about 750 km of Uniboom™ data collected in 1975 (Fig. 4). An additional 300 km of Uniboom data from the 1976 Mount Mitchell cruise, and about 120 km of 3.5 kHz and 8kJ Mini-sparker data collected on the Atlantis II cruise, provided tie-lines in the mid-shelf region. On the outer shelf, 150 km of the Fay 021 8 kJ Mini-sparker data collected in 1974 and the seismic reflection stratigraphy developed from sparker data by Harris (1983) were examined. The Atlantis II data were tied to the Barnegat Inlet stratigraphy established by Ashley et al. (1991) through use of about 250 km of Uniboom™ and 3.5 kHz data collected on the Whitefoot cruise in 1981.

There are large numbers of Vibracores penetrating the upper 6-9 m of the sediment, particularly on the inner shelf, and several wells drilled by oil companies have cored deeper deposits. However, groundtruth for the upper Pleistocene section on the shelf below the surficial sediments is limited to the five holes drilled by the Atlantic Margin Coring Project (AMCOR) (Fig. 4.). Poor recovery rates (rarely more than 30% in the upper Pleistocene) limit the usefulness of the cores for lithostratigraphy (Hathaway et al., 1976), but the cores have provided material for determining the approximate age of the deposits through foraminiferal and calcareous nannofossil biostratigraphy (Poag in Hathaway et al., 1976; Valentine in Harris, 1983), palynostratigraphy and aminostratigraphy (Groot et al., 1995).

In general terms, sequences and systems tracts in the seismic data were interpreted in a fashion consistent with sequence stratigraphic methods (Mitchum and Vail, 1977; Mitchum et al., 1977a; 1977b). Some of the data were interpreted somewhat differently, however, due to the higher vertical resolution of the data used, and due to known characteristics of the New Jersey shelf sediments. For example, incised valley fill deposits are generally assigned to the lowstand systems tract (LST) in the Exxon model (e.g. Posamentier and Vail, 1988). However, valley fills on the mid-Atlantic shelf are largely filled with Holocene lagoonal-estuarine deposits, and contain little or no alluvial material, clearly indicating that the fill is transgressive in origin (Oertel et al., 1989; Thorne and Swift, 1991, Ashley and Sheridan, 1994). Lowstand deposits within valley fills appear to be confined to the basal fill of the most seaward valleys (Belknap and Kraft, 1985). For this reason, I interpret most of the valley fill deposits as belonging to the transgressive systems tract (TST).

All depths quoted in the text and appearing on the structural contour maps were determined by assuming a 1.5 km/s sound wave velocity for the water column, and a 1.7 km/s sound wave velocity for the sediments. The velocity in the sediments was chosen as a typical value from the sonic logs of the upper Pleistocene section in the AMCOR drill sites.

5.4 Seismic Stratigraphy

5.4.1 Description of Depositional Sequences

Using the criteria described above, the stratigraphic section was broken into four depositional sequences. The prominent reflections seen on Atlantis II lines 2-14 and 2-16 are shown in line diagram form in figures 11 and 12 respectively. These lines are chosen as examples of the shelf stratigraphy because of the relatively good quality and continuity of the seismic data, and because of the different characteristics which they feature. Line 2-16 provides a tie-line for AMCOR wells 6010 and 6020, and is along the same roughly shore-perpendicular transect as holes 6011 and 6021 (Fig. 4). The seismic line is approximately 100 km in length and is continuous except for two gaps in the data on the inner shelf (Fig. 12) with a combined length of less than 4 km.

Line 2-14 runs slightly oblique to line 2-16 so that the distance between the lines declines from 22 km on the inner shelf to 15 km on the outer shelf (Fig. 4). The line begins approximately 14 km offshore and is 134 km in length; it is largely continuous, but there are four short (< 2 km) gaps (Fig. 12).

5.4.1.1 Sequence I. This unit lies immediately above the deepest observed reflection surface of regional erosional truncation that is recognized across a large part of the shelf. The evidence of erosion along this surface is vividly expressed in Fig. 13, which shows approximately 2 km of data from the inner shelf along line 2-14 (Fig.4). The base of Sequence I is the lower of two strong reflections that erosionally truncate a series of dipping reflections with an apparent seaward gradient of approximately 0.007. This reflection is discontinuous in this area, because of truncation by an overlying sequence boundary, but generally parallels the seafloor with a gradient of approximately 0.0005. Thus, this reflection marks a prominent angular unconformity that is adequate evidence of a sequence boundary.

The topography of the sequence I boundary shows a generally seaward increase n depth and one major valley. In the southern part of the study area, it emerges clearly from beneath the multiple 40 km offshore, as on line 2-16. Farther north, it is discontinuous, except in the vicinity of the Hudson Shelf Valley (Fig. 19b), where it deepens to more than 60 m, before being truncated by the base of sequence IV.

Apart from this deepening in the extreme north of the study area, the structural contours roughly parallel the modern shoreline to a distance of approximately 60 km offshore (Fig. 19a). The surface dips gently seaward (approx. ~0.0005) but is marked by numerous small valleys with local relief of up to about 10 m, and apparent widths

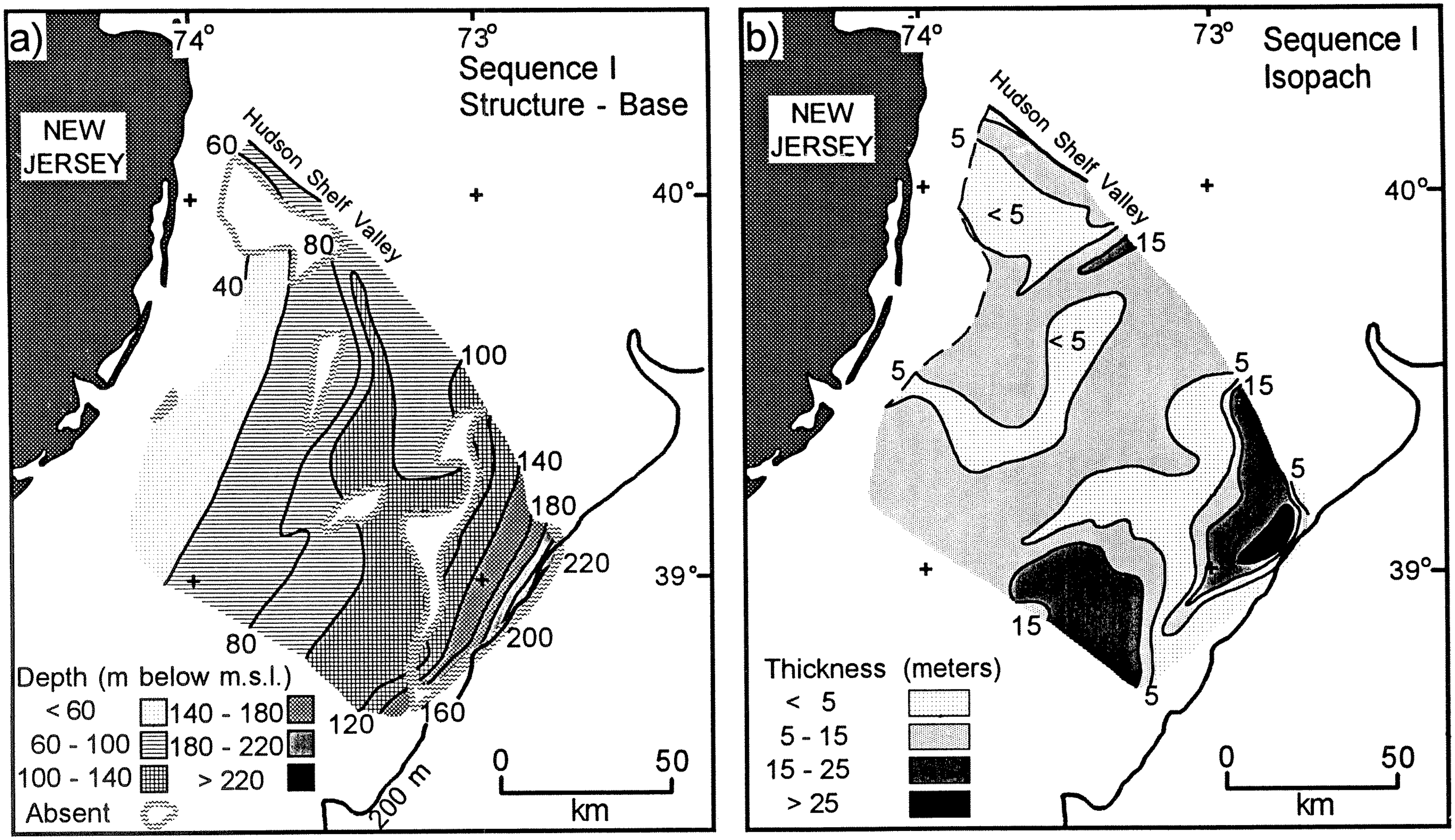


Figure 19. Generalized structure (a) and isopach maps (b) for sequence I. All thicknesses assume a seismic velocity of 1.7 km/s through the sediment.

of up to 2 km. In the southern part of the study area, this gentle seaward dip continues across the mid-shelf, but elsewhere the sequence boundary dips steeply (gradient ~0.0025) into a prominent swale with a maximum depth of approximately 35 m relative to its flanks (Fig. 11, 19a). This swale becomes deeper and wider southward from the Hudson Shelf Valley to a maximum width exceeding 30 km.

On the outer shelf, the sequence onlaps a prominent high in the older Pleistocene sediments, and is thin to absent in the northern and central parts of the outer shelf (Fig. 19b). Farther south, however, the sequence boundary continues to slope gently seaward to the outermost shelf, where the gradient steepens to roughly 0.005. Elsewhere, sequence I reappears farther offshore with a similar gradient which extends to the shelf edge. The unit thickens substantially offshore, locally exceeding 30 m in thickness. On the upper slope, the unit thins abruptly, where it is either truncated by overlying reflections or outcrops at depths ranging from 190 (line 2-18) to 230 m (line 2-14) as in Fig. 12.

Internal reflections within sequence I are generally weak, probably due to the loss of energy of the signal as it passes through the overlying sediments. Those that are present generally parallel its base, but there are a few descending high angle reflections on the inner shelf on line 2-16 (Fig. 11). The one exception is the most southerly line, line 2-20 (Fig. 4), on which a strong reflection horizon can be traced from the mid-shelf, where it onlaps the sequence boundary at an elevation of approximately -75 m, for a distance of roughly 60 km to the outer shelf. This reflection largely parallels the underlying sequence boundary, and lies 5 to 7 m above it until being truncated by the overlying sequence boundary on the outer shelf at an elevation of -105 m. About 5 km seaward of AMCOR site 6009 (Fig. 4), this surface is downlapped by a set of steeply-inclined (gradient ~.004), 15 m high reflections.

5.4.1.2 Sequence II. The sequence boundary at the base of sequence II is the prominent reflection surface that erosionally truncates the sequence I boundary in Figure 14, which is from line 2-16 near the location of AMCOR drill site 6020 (Fig. 4). Further evidence of erosion along this surface is provided by the truncation of clinoform reflections in Sequence I on the inner shelf (Fig. 11), and several small valleys seen on line 2-17 (Fig. 4). Overlying reflections begin to onlap this surface at a point far below and seaward of the Ashley et al. (1991) shoreline, implying a significant downward shift in coastal onlap.

On the inner shelf, this sequence boundary is present only within small valley fills in the northern part of the study area (Fig. 12). The structural contours largely parallel the shoreline except between Barnegat Inlet and Great Bay (Fig. 20a), where the surface is incised by several channels, the deepest of which has a maximum depth of approximately 8 m, and is located 10 km south of Barnegat Inlet. This sequence boundary can be traced beneath the highstand shoreline identified by Ashley et al. (1991) using intersecting seismic lines from the Whitefoot cruise (Fig. 4).

As in the underlying sequence I, a prominent swale is developed in the mid-shelf region in the northern and central parts of the study area (Fig. 20a). This swale extends from the Hudson Shelf Valley southward to a point approximately 80 km southeast of Barnegat Inlet. This swale is filled with a sediment package that thickens from 15 m in the north to up to 30 m in a triangular shaped region in the central part of the study area. Outside of this swale, the sequence boundary slopes gently

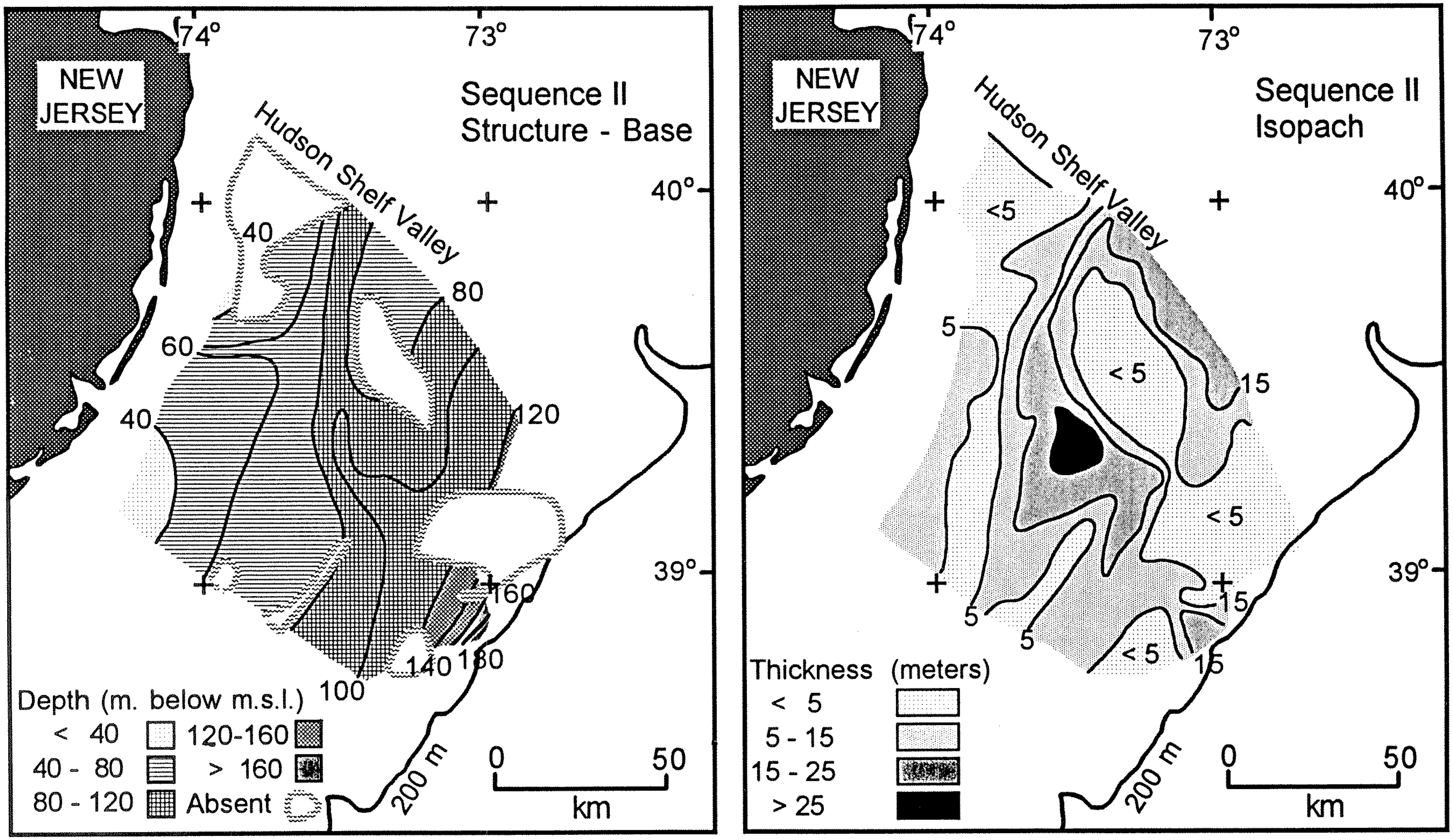


Figure 20. Generalized structure (a) and isopach maps (b) for sequence II. All thicknesses assume a seismic velocity of 1.7 km/s through the sediment.

seaward, where present.

On the outer shelf, the sequence is largely absent in the northern part of the study area and thickest south of the seaward end of the swale (Fig. 20a, b). On the outermost shelf, the sequence is present only in the Spencer Canyon area, where the sequence boundary dips steeply offshore (gradient ~0.005). Elsewhere the sequence is erosionally truncated by overlying sequence boundaries.

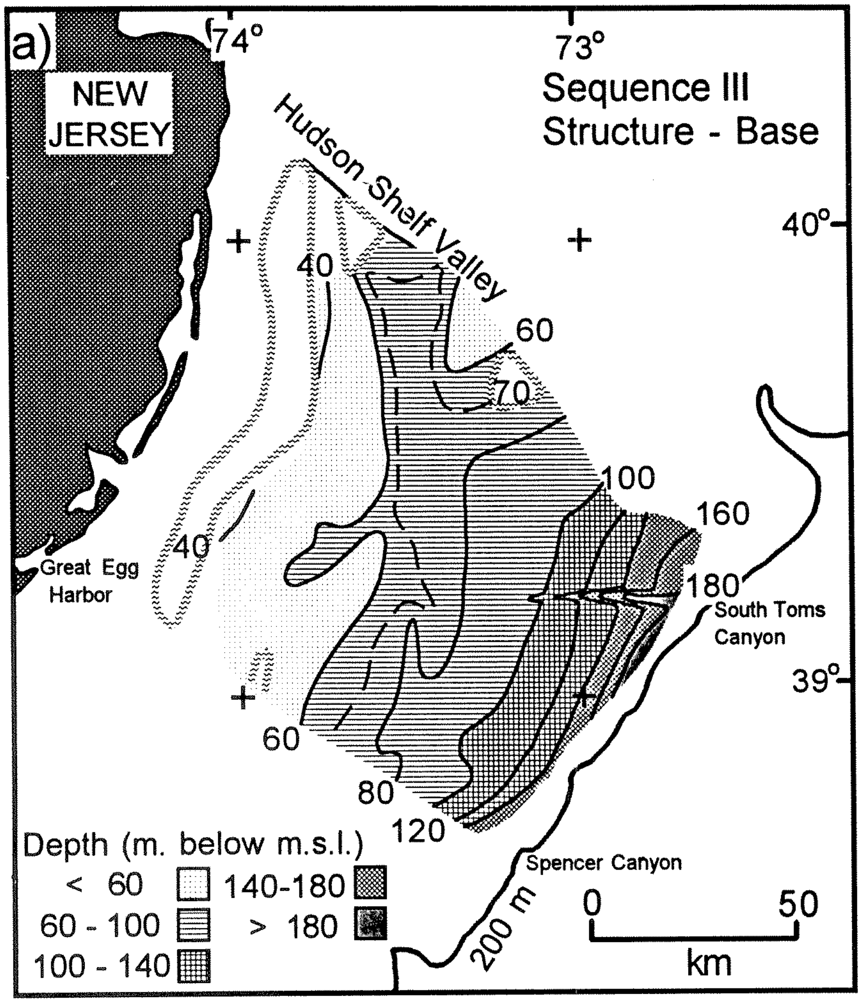
Above the sequence boundary and within the mid-shelf swale, there are two distinct packages of steeply seaward-inclined reflections (gradients > 0.05) separated by a prominent, planar to swaley surface (relief up to ~5 m) that truncates the reflections in the lower package (Fig. 11b, 12b). The upper package is shown in Fig. 14. These packages have a combined thickness of up to 15 m. This truncation surface dips down-valley with a gradient of approximately 0.001, though it appears nearly horizontal on shore-perpendicular lines.

The upper package of inclined reflections is truncated by the strong, continuous, largely planar reflection shown in Figure 14. This surface appears to slope gently (0.0002) seaward on the seismic lines, while having a down-valley gradient of approximately 0.0008. Above this prominent reflection are several reflections that are parallel to slightly oblique to it, and onlap the sequence boundary on the inner shelf. Outside of the mid-shelf swale, sequence II is nearly everywhere comprised by relatively faint, but continuous, concordant reflections onlapping a gently sloping sequence boundary. The only exception to this is seen on the outermost shelf on line 2-18, where the sequence boundary is a prominent scarp truncating the underlying sequence I, and is overlain by largely seismically transparent sediments.

5.4.1.3. Sequence III. The base of sequence III is incised by numerous valleys, the largest of which is the prominent Paleo-Hudson valley described by Knebel et al. (1979). Figure 16 shows this valley where it crosses line 2-14. The depth of incision of this valley (up to 35 m) is evidence of fluvial entrenchment, although the valley was probably widened and reworked by marine processes subsequently. A series of shallow valleys ($<$ 10 m) with apparent widths of up to 5 km is seen on line 2-16 (Figs. 11, 14).

Sequence III is absent within 30 km of the shore, and the erosional edge trends eastward in the northern part of the study area (Fig. 21a), parallelling the hinge line of the Baltimore Canyon Trough (Fig. 1). Throughout the inner shelf, the sequence is discontinuous, present only within incised valleys in the base of the sequence, or highs in overlying units. The large Paleo-Hudson valley and a possible tributary valley extending southwestward toward Great Egg Harbor can be seen (Fig. 21a). Several smaller valleys can be seen on the isopach map (Fig. 21b) owing to its finer scale (5 m, rather than 20 m). From 30 to 75 km offshore, the structural contours parallel the shore except where incised by valleys, and the overall seaward gradient of the surface is approximately 0.0008. The unit also generally thickens seaward, exceeding 15 m about 75 km offshore, and within a broad swath around the Paleo-Hudson valley and its tributary valleys.

Farther offshore, the gradient of the sequence boundary steepens to about 0.0025. A deep ($>$40 m) valley trends westward from South Toms Canyon, noted by Harris (1983). The unit is generally more than 15 m thick on the outer shelf (Fig. 21b), except below lows in the overlying sequence IV boundary. Near the shelf



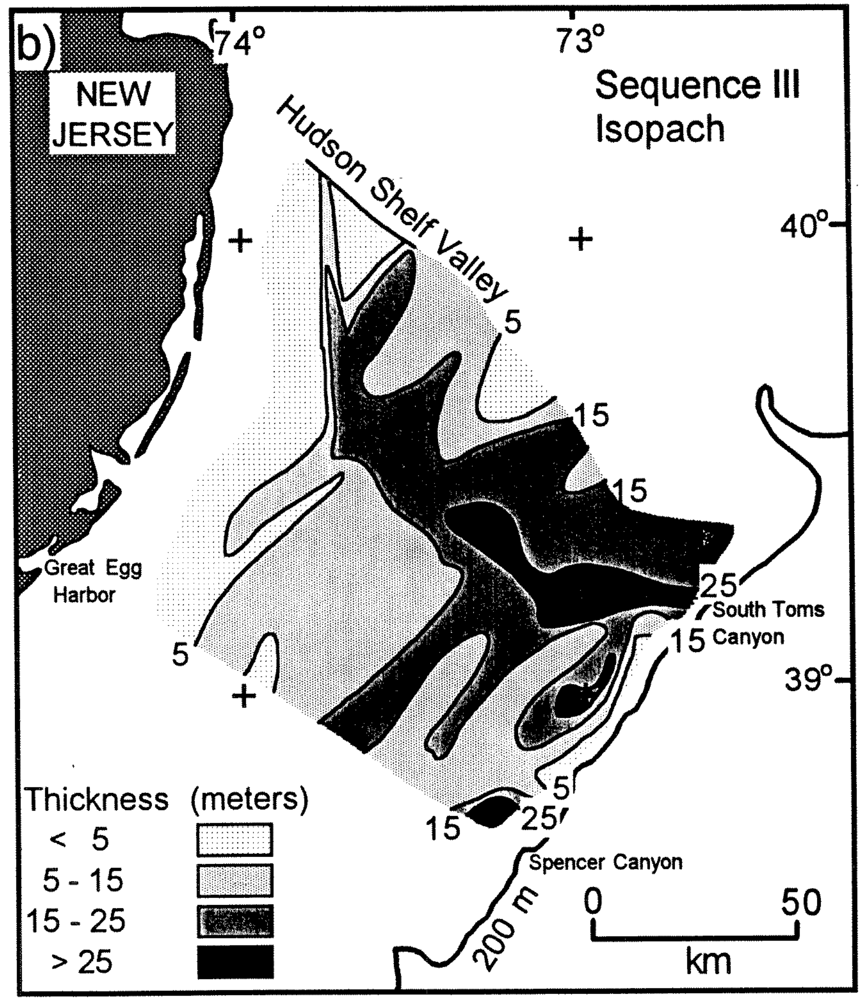


Figure 21. Generalized structure (a) and isopach maps (b) for sequence III. All thicknesses assume a seismic velocity of 1.7 km/s through the sediment.

break, it is typically truncated by the overlying sequence IV boundary (Fig. 12).

The internal reflections within this sequence display a complex and varied morphology. On the outermost shelf, there is a series of strong, concordant, gently seaward sloping reflections that onlap the underlying sequence boundary at elevations ranging from as deep as -140 m on line 2-20 to as high as -80 m on line 2-16. On the middle and inner shelf, the deepest reflections are those preserved within valley fill deposits. Small valleys, such as those in Figure 15, are generally filled with acoustically transparent seismic facies, but the Paleo-Hudson valley fill has a more complex fill pattern. Line 2-16 intersects the seaward end of this valley approximately 80 km offshore (Fig. 16), where the valley is roughly 15 m deep and filled by landward-inclined reflections that are truncated by a second valley incision. This second valley is similarly filled, then truncated by a largely planar, gently seaward dipping reflection.

Nearer shore along line 2-14, the Paleo-Hudson valley fill stratigraphy is even more complex (Fig. 15). The lowermost part of the valley fill consists of a series of continuous parallel reflections with gentle southeastward dips onlap the valley walls. In the eastern half of the valley these reflections are overlain by a 10 m thick acoustically transparent facies, which is truncated by a prominent undulating reflection that onlaps the upper valley wall. One of the channels in this surface is filled with inclined reflections with apparent northwest dips of approximately 4°.

Above the prominent reflections which truncate all the valley fills, the sequence is largely characterized by parallel, concordant reflections that onlap the sequence boundary at elevations up to ~ -40 m. This is the youngest seismic unit in the sequence within the study area, except along the outer shelf of line 2-14, where the concordant reflections steepen seaward and are downlapped by high angle (gradient ~0.01) seaward-dipping reflections.

5.4.1.4. Sequence IV. Sequence IV has been investigated by other authors, such as McClennen (1973), Milliman et al. (1990), and Davies et al. (1992). The lower boundary essentially corresponds to the R-reflection described by McClennen (1973), and has considerable relief, with channels up to 10 m deep across the entire shelf (Figs. 11, 12) in addition to the deeper incision of the Hudson Shelf valley. The presence of this unit is sporadic on most of the shelf, largely restricted to the sand-ridge fields and incised valleys that comprise the "Transgressive Shelf Blanket" described by Swift et al. (1972). There are only two areas of the shelf where thick, continuous deposits exist above the Sequence IV boundary: the outer shelf, where the "Outer Shelf Wedge" is developed, and a triangular region of the mid-shelf (the "Mid-Shelf Sediment Wedge"; Milliman et al., 1990).

The Outer Shelf Wedge is visible on both Figures 11 and 12, though only the most landward part of it can be seen in Figure 11. It consists of a wedge of reflections that onlap a prominent scarp in the underlying sequence boundary and thickens to a maximum thickness of as much as 60 m at the shelf edge. The unit is truncated by a nearly horizontal planar reflection on line 2-14 (Fig. 12) and by the seafloor on line 2-16 (Fig. 11) at approximate elevations of -95 m and -75 m respectively. This unit can be seen on all the lines that reach the outer shelf; on lines 2-18 and 2-20, it reaches a similar elevation to that on line 2-16.

The mid-shelf wedge can be seen most prominently on line 2-14 (Fig. 12), where this unit thickens seaward to a thickness of up to 20 m, then thins abruptly in a prominent scarp. Strong, continuous internal reflections onlap the sequence boundary landward and downlaps a prominent planar reflection surface a few meters above the sequence on the seaward side.

5.4.2 Age Constraints on the New Jersey Shelf Sequences

Though absolute age data are scanty, some limits can be placed on the ages of the sequences from information from AMCOR wells. The base of sequence I lies above the lowest occurence of the calcareous nannofossil *Emiliania huxleyi* in AMCOR 6010 (Harris, 1983), which first appeared during stage 8, ca 275 ka (Thierstein et al., 1977; Berggren et al., 1980). An unknown shell from just above the boundary in AMCOR site 6020 gave an amino acid ratio of 0.21, which is consistent with a stage 5 age (Fig. 2), if the shell racemizes at a rate comparable to *Mercenaria* (Groot et al., 1995).

Knebel and Spiker (1977) reported several radiocarbon dates of > 40 ka from surficial sediments on the outer shelf just seaward of the Mid-Shelf Sediment Wedge that probably belong to sequence III. Other radiocarbon ages from this unit ranging from 28 ka to > 40 ka were reported by Knebel et al. (1979) and Davies et al. (1992).

The base of Sequence IV corresponds to the R-reflection of McClennen (1973). The lower boundary of this sequence is reasonably well-constrained by radiocarbon dates, and is believed to correspond to the stage 2 lowstand. Although Knebel et al. (1979) considered the Paleo-Hudson Valley to be part of the R-reflection, Harris

(1983) showed the R-reflection to be stratigraphically higher, a finding which is consistent with this study.

5.5. Discussion

5.5.1. Depositional Sequence Interpretation

5.5.1.1. Sequence I. It is difficult to assign systems tracts to sequence I, because of the lack of prominent internal reflections. However, the general pattern of a thin, discontinuous sheet over much of the inner to mid-shelf, a zone where the sequence is absent on the outer shelf, and maximum thicknesses near the shelf edge (Fig. 19b), is similar to the distribution of Sequence IV. Therefore, it seems probable that the thick shelf edge sediments are analogous to the "Outer Shelf Wedge" and represent a lowstand to early transgressive wedge, whereas the discontinuous sheet on the inner to mid-shelf consists predominantly of transgressive deposits. Hence, the lowstand shoreline would have been at or near the shelf edge (Fig. 22a).

Steckler et al. (1993) predicted that on shelves where rates of sea-level change greatly exceeds sediment supply and tectonic subsidence, the TST and LST should predominate, and the HST should be poorly developed. One possible exception to this is seen on line 2-20, where a prominent prograding wedge can be seen that could represent a late stage highstand deposit. The similarity of this sequence I to sequence IV suggests a sea-level fluctuation of comparable amplitude. The SPECMAP curve suggests similar ice volumes during stage 6 and stage 2. If the base of sequence I was cut during stage 6, it would indicate a stage 5 age for the shell material found in AMCOR site 6020, which is consistent with the amino acid racemization data (Groot

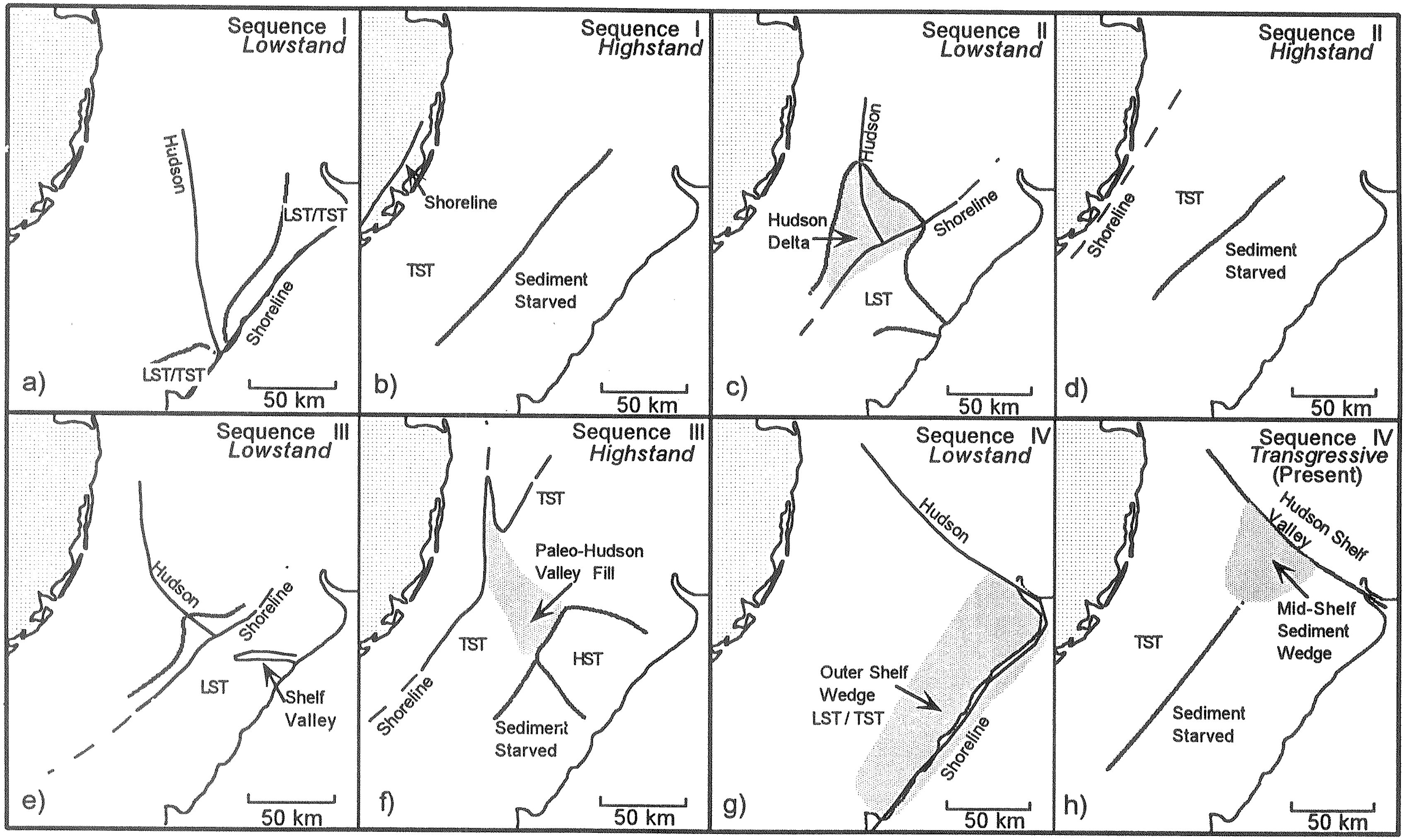


Figure 22. **Shoreline configurations, Hudson positions and deposits during deposition of shelf sequences.** Shoreline positions shown are the approximate maximum lowstand and highstand. Deposits shown in "highstand" intervals include transgressive deposits, those shown in "lowstand" may include early transgressive shelf edge deposits.

et al.,1995).

The most logical explanation for the deep mid-shelf swale seen in this unit (Fig. 19b) is that the Hudson River flowed through this region during at least part of stage 6, carving a broad valley (Fig. 22a). This valley was re-incised during the lowstand associated with sequence II, preventing preservation of a thick sediment package. The deeper, thicker sequence I just south of the Hudson Shelf Valley suggests that the landward end of the Hudson, was flowing in that vicinity during stage 6. The highstand shoreline (stage 5e) is represented by the Cape May Escarpment (Fig. 22b).

5.5.1.2. Sequence II. The two packages of clinoforms in the mid-shelf swale in this sequence, combined with the prominent thickening in this region (Fig. 20b), could represent a lowstand deltaic deposit. This implies that during sequence II, the shoreline prograded to a mid-shelf position, 75 km seaward of its current location (Fig. 22c). One possible explanation for the prominent, nearly planar surface between the two packages within sequence II is a marine flooding surface, which would make these two lowstand parasequences. An alternative explanation is that this surface actually represents a sequence boundary. The absence of transgressive and highstand deposits associated with the lower package could result from subsequent erosion of the thin TST and HST units developing during a short-lived sea-level cycle.

The most widespread unit in this sequence lies above this transgressive surface. The nearly horizontal onlapping reflections within in this unit, and absence of prograding clinoforms suggest that this unit belongs to the TST. This TST is downlapped by the Ashley et al. (1991) highstand shoreline that implies a highstand of approximately -20 m (Fig. 22d).

The Hudson River apparently continued to flow southward across the mid-shelf during deposition of this sequence. The outer shelf appears to have been largely sediment starved, except south of the Hudson Delta and within the Spencer Canyon area (Figs. 1, 20b, 22c). The concentration of sediment south of the primary input is consistent with the dominant southward transport of sediment by longshore drift in the Mid-Atlantic Bight (Swift et al., 1981).

5.5.1.3. Sequence III. The deepest, and presumably oldest, sediments in this sequence are preserved within the shelf-edge valley extending westward from South Toms Canyon (Fig. 21a). The fill of this deposit varies from chaotic on the Fay 021 Minisparker line to prograding on line 2-14, about 10 km landward (Fig. 12). The chaotic fill near the shelf edge is suggestive of mass movement or slumping. A possible explanation for this is that the combination of high sediment input from the nearby Paleo-Hudson, storm waves impinging on the bottom during the lowstand, and the steep gradient of the canyon wall at the shelf edge, triggered headward erosion of the canyon, producing a shelf valley. The valley filled through collapse of adjacent walls and seaward progradation of a lowstand wedge.

The seaward end of the Paleo-Hudson Valley is characterized by two incisions and filled by apparently landward inclined reflections (Fig. 11). These reflections could result from progradation of a flood-tidal delta into an estuary. Alternatively, because the seismic section crosses the valley obliquely, they could represent southward progradation of a baymouth spit. Similar features are seen on records from modern Delaware Bay (Ashley and Sheridan, 1994). In either case, the valley fill occurred during the transgression, and the sharp overlying reflection likely stems from a wave-cut ravinement surface. The position of this baymouth indicates that the shoreline prograded slightly farther offshore during deposition of this sequence than in the underlying sequence II (Figs. 22c, 22e).

The complex Paleo-Hudson valley fill has a striking tripartite division that may correlate to that predicted by the Dalrymple et al. (1992) model for estuarine fills, and the Ashley and Sheridan (1994) model for large incised valley fills. The lowest part consists of strong, gently dipping parallel reflections that could represent fluvial point bars or progradation of a bayhead delta into an estuary. The central, nearly reflection-free facies is resembles the muddy valley fills seen in late Pleistocene-Holocene deposits from the inner shelf of New Jersey (Ashley and Sheridan, 1994). Predominantly muddy sediments typically fill the central basin of wave-dominated estuaries (Dalrymple et al., 1992). The more chaotic, multiply-incised, upper fill is suggestive of migrating tidal channels as the baymouth approached the area and tidal influence increased. Thus, this fill also appears essentially transgressive in nature. An approximate position for the shoreline during the highstand of this unit is indicated in Fig. 22f.

The acoustically transparent sediments in the smaller valley fills indicate mud deposition, and are similar to the small- to medium- sized incised valley fills in the Late Pleistocene / Early Holocene section of New Jersey (Ashley and Sheridan, 1994). Thus, they also represent valley filling during transgression. Transgressive valley fill deposits constitute a substantial proportion of the deposits in Sequence III on the middle and inner shelf.

The largely parallel, concordant reflections above are probably also largely transgressive. On line 2-14, the downlapping reflections on the outer shelf may be the only preserved part of the highstand from this unit. Their presence reflects either a high enough sediment supply to allow preservation of deposits during falling sea level, or a prolonged still-stand at an elevation of approximately -75 m.

In summary, sequence III records a fall in sea level, resulting in extensive fluvial incision. During this time, the Hudson flowed southward across the shelf, but slightly to the northwest of its position in the underlying sequence II, and debouched to an outer shelf shoreline approximately 85 km southeast of Barnegat Inlet (Fig. 22e). A large shelf valley formed by headward erosion from South Tom's River Canyon toward the mouth of the Paleo-Hudson. Subsequently, sea level rose, filling the valleys and overtopping them with a relatively thick transgressive unit. On the outer shelf, a regressive highstand shoreline is recorded on the outer shelf, implying high sediment load during a period of stable or falling sea level (Fig. 22f).

5.5.1.4. Sequence IV. Following the terminology of Milliman et al. (1990), the deposits of sequence IV can be divided into an Outer Shelf Sediment Wedge, a Mid-Shelf Sediment Wedge, and a Transgressive Shelf Blanket (Swift et al., 1972). The oldest of these deposits, the Outer Shelf Wedge was described in detail by Davies et al. (1992), who found a secondary channelled horizon in the middle of the deposit. They gave two possible age models for its deposition, one in which the R-reflector represented a mid-Wisconsin sea-level fall, and the secondary channelled horizon represented the late Wisconsin maximum (~ 20 ka). The other model followed the interpretation of Milliman et al. (1990), placing the R-reflector at the late Wisconsin maximum, and attributing the channelled horizon to a later perturbation in sea-level.

The latter interpretation appears more likely, because the channels seen at this horizon could have been cut into a marine flooding surface by tidal creeks, rather than by fluvial incision. Davies et al. (1992) regarded the absence of terrestrial or littoral sediments as suggestive of a subtidal origin, but suggested that the stiffness of the underlying mud and meandering of the channels was indicative of a fluvial origin. However, tidal channels meander, and in the absence of extensive, deep incision on this surface, the cohesiveness of the material cannot be considered sufficient evidence of subaerial exposure. The absence of incisions of the depth expected for paleovalleys of the Hudson or even the Mullica River (Fig. 1) further south suggests that this surface is unlikely to represent the maximum lowstand regression of the late Wisconsin.

The outer-shelf wedge is truncated by a transgressive surface. Marine erosion during the transgression stripped much of the outer shelf of sediment, leaving older Pleistocene units exposed at the seafloor. The mid-shelf sediment wedge is a constructional feature (Milliman et al., 1990) formed by progradation, that downlaps this transgressive surface (Fig. 12). Davies et al. (1992) and Milliman et al. (1990) are likely correct that it must have formed during a time of high sediment input, and perhaps a slowing of the rate of sea-level rise during the late Pleistocene.

Subsequent transgressive deposits are discontinuous, concentrated in valley fills and sand ridges. The absence of further parasequence development during the transgression implies that sediment supply to the shelf has not been sufficient to offset the effect of relative sea-level rise at any time since the formation of the mid-shelf sediment wedge.

similar effect may have occurred during stage 6, as well. The effects of the post-glacial collapse of the peripheral bulge can be seen not only on ancient shorelines (Dillon and Oldale, 1978), but on the elevation of underlying reflections. The prominent horizontal reflection horizon within the outer shelf wedge, and the outcrop of sequence boundary I on the continental slope, are 20 to 40 m deeper on line 2-14 than on more southerly lines (Fig. 4).

The absence of re-incision of the Paleo-Hudson Valley during this sequence indicates that the Hudson must have switched to the modern shelf valley position by the time of the sequence IV lowstand (Fig. 22g). The change in course of the Hudson could be related to the advance of ice into the region. The weight of the ice sheet may have depressed the crust immediately in front of it, creating a peripheral bulge to the south, and enticing the Hudson into a more northerly course. A possible explanation for the tendency of the Hudson to occupy either a position near the modern shelf valley or one significantly farther south is local uplift on the outer New Jersey shelf resulting from salt tectonics. The only known salt diapirs in the study area lie between these two Hudson positions (Fig. 1).

5.5.3. Impact and Origin of Sediment Supply Changes on Stratigraphy

Changes in the amount and location of sediment supply to the shelf have influenced the stratigraphy in several ways. The thickest sediments in all sequences are typically located within the Hudson Valley , and to the south of the contemporaneous Hudson outflow (Figs. 19b, 20b, 21b). The mid-shelf sediment wedge in sequence IV is also located to the south of the Hudson Shelf Valley and are typically located within the Hudson Valley, and to the south of the contemporaneous Hudson outflow (Figs. 19b, 20b, 21b). The mid-shelf sediment wedge in sequence IV is also located to the south of the Hudson Shelf Valley and likely reflects sediment transported southward by longshore drift. The position of the Hudson River also appears to have influenced the activity of submarine canyons. The buried valley extending from South Toms canyon suggests that it may have resembled the modern Hudson Canyon during the deposition of sequence III, when the Hudson outflow was south of its current position.

Several characteristics of sequence III suggest a substantially enhanced sediment supply compared to the underlying sequence II. Sequence III is substantially thicker, particularly on the outer shelf, is less restricted in its distribution, and contains a well-developed highstand wedge. While a longer period of deposition, or less extensive erosion is possible as a partial explanation, increased sediment supply derived from glacial activity in the Hudson drainage basin is possible.

Glacio-lacustrine varves from the St. Lawrence Valley in southern Quebec (Fig. 5) imply that the St. Lawrence River was blocked by ice during at least part of stage 5, and throughout stages 4 through 2 (Clet and Occhietti, 1994). The Sunnybrook Drift, a diamicton unit in the Toronto area (Fig. 5), has been variously interpreted as a till from an early Wisconsin glacial advance (Karrow, 1967; Hicock and Dreimanis, 1989), or a glacial-lacustrine unit (Eyles and Eyles, 1983). Either interpretation implies that the St. Lawrence Valley was blocked by ice and there was ice in or near the Toronto area at this time. At least part of the St. Lawrence drainage basin must have been captured by the Hudson River during this time, as it was during the late

Pleistocene (Teller, 1987), adding substantially to the flow of the Hudson River. However, given that much of the drainage of the Laurentide must have been through proglacial lakes, where much of the sediment would have settled out, it may not be a viable source of sediment.

Additional sediment may have been directly provided by waxing and waning of small ice caps in the northern Appalachians (Fig. 5). The Chaudhière Till, possibly of stage 4 age, in southern Quebec was formed by ice moving out of the northern Appalachians that was confluent with the Laurentide ice in the St. Lawrence Lowlands (Karrow and Occhietti, 1990). The presence of ice in this region suggests that alpine ice may have been present in areas such as the White Mountains, the Adirondacks, and perhaps even the Catskills (Fig. 5). Currently, little evidence from New England and upstate New York exists to confirm or reject this hypothesis, because of few exposures that predate the late Wisconsin and little material suitable for dating. Koteff and Pessl (1985) suggested that the widespread "Lower Till" in New England was deposited during an early Wisconsin glacial advance, but Oldale and Colman (1992) regard it as an older unit, at least in southern New England. There is also evidence of possible early and middle Wisconsin glaciation in western New York, but the precise timing of these events is equivocal (Muller and Calkin, 1993).

An alternative source of sediment is reworking of pre-existing Quaternary units in the drainage basin. Proxy climatic data for the Hudson drainage basin is scarce, but inferred early to middle Wisconsin lacustrine deposits on the north shore of Lake Erie and western New York (Fig. 5) yield pollen assemblages suggestive of a moist, forest-tundra environment (Calkin et al., 1982; Dreimanis, 1992). Increased precipitation around the margins of the ice sheet is possible, as climatic models suggest increased storminess resulting from the thermal contrast between the ice sheet and the ocean (Kutzbach, 1987). Heavy precipitation combined with a change in vegetation from forest to tundra in uplands could result in increased erosion. Church (1972) notes that tundra climates are characterized by flashy, surficial drainage, which also favors increased rates of erosion.

A significant change in sediment input may also have produced the mid-shelf sediment wedge, which is associated with the Fortune shoreline (Fig. 7), which has been dated as approximately 11.5 ka (Dillon and Oldale, 1978; Milliman et al., 1990). Thus, it could relate to the slowing of sea-level rise between 12 ka and 10 ka recognized from reefs in the Bahamas and New Guinea (Fairbanks, 1989; Edwards et al., 1993). High sediment supply must also have played a role in order to create a sediment wedge of this size, however, with an approximate area of 2500 km^2 and average thickness of 10 m. The modern Hudson River supplies only 870,000 metric tons of sediment per year to the lower Hudson (Ellsworth, 1986). Given that rate of sediment supply, roughly 50,000 years would be required to create the deposit, clearly indicating the need for greater sediment supply during this late Pleistocene event.

5.5.4. Recognition of Paraglacial Effects on Stratigraphy

If the glacial record from the adjacent continent is not preserved, the paraglacial origin of the New Jersey Quaternary would not be obvious. It was beyond the limit of direct glacial action, so no tills were deposited in the region. Furthermore, the ice was land-based in this area, and the nearest calving ice margins were well to the north in New England (Dyke and Prest, 1986). As a result, ice-rafted debris is not reported from Pleistocene sections on the New Jersey slope.

In the absence of clear evidence of glaciation, a persuasive case based on the stratigraphy alone, might be made for some form of tectonic activity. During the Quaternary, there was a sharp increase in sediment accumulation on the slope and rise, and an increase in sediments derived from the northern Appalachians (Poag and Sevon, 1989). The glacial-isostatic effects seen in the northern part of the study area could be attributed to isostatic loading resulting from an orogenic event, rather than a glaciation. Thus, a case could be made for tectonic uplift of the source areas, especially those to the north of the study area, on stratigraphic evidence. How could the Quaternary record of the New Jersey margin be distinguished from one resulting from tectonic uplift, such as has been postulated for the Miocene in the same region? Detailed analysis of the deposit would yield three possible clues that might allow one to infer the presence of glaciers. These are evidence of high-frequency, high-amplitude eustasy, the concentration of sediment supply at certain points in the sea-level cycle, and evidence of coeval climatic fluctuations.

The presence of numerous subaerial erosion surfaces, the prominent difference in elevation between lowstand and highstand deposits, and the dominance of transgressive and lowstand systems tracts on the New Jersey margin all are suggestive of high rates of relative sea-level change (Steckler et al., 1993; Chapter 4). At least four sequences, recording substantial sea-level falls were deposited within a period of no more than 275 ka. The exceptional rapidity of these sea-level changes, especially if coeval deposits in other regions showed similar effects, might lead one to suspect a glacial-eustatic origin. However, this would be inconclusive, as tectonics can also produce rapid relative sea-level change, and even if glacio-eustasy accounts for the sea-level changes, the sediment supply need not be related to the glaciation.

More unusual is the timing of sediment delivery to the basin. Through most of a Pleistocene sea-level cycle, the New Jersey margin behaves as if it were sediment starved. Unlike the middle Miocene, which was characterized by the progradation of thick highstand deltas onto the shelf, the Quaternary section is marked by erosionally truncated highstands, lowstand shorelines that are detached from the underlying highstands, and sequences that are relatively thin and discontinuous across most of the shelf. These characteristics are suggestive of a forced regression, which is driven by relative sea-level fall rather than sediment supply (Posamentier et al., 1992). Shoreline progradation is recorded only in lowstand and early transgressive deposits, suggesting that the sediment supply was insufficient during other times. Even this would not conclusively demonstrate a glacial origin, because lowstands induced by tectonic uplift would also be associated with pulses of sediment.

A final piece of evidence is the proxy climatic record from New Jersey. Pollen samples from the margin show Quaternary vegetation changes from the modern vegetation to subarctic boreal forest, implying temperature shifts of up to 10°C (Groot et al., 1995). The extreme climatic shifts would give additional evidence of a glacial-eustatic origin for the sea-level changes, and the subarctic climate during lowstands would suggest that a glacier could be nearby. These data, combined with the concentration of sediment in lowstands and early transgressions, would strongly suggest a glacial origin. It should be remembered, however, that glacial drainage may extend far from the climatic region of the glacier, as the Mississippi drained much of the Laurentide ice sheet. Thus, a glacial interpretation should be considered even in subtropical basins if there is evidence of pronounced cooling associated with the lowstands.

5.6 Conclusions

Four late Quaternary depositional sequences can be recognized from the shelf stratigraphy of New Jersey. Sequences I and IV likely record the major glacial / interglacial changes during isotope stages 6/5e and 2/1 respectively, while sequences II and III resulted from smaller-scale glacio-eustatic fluctuations between stages 5e and 2. Sequences I and IV are marked by thick shelf-edge lowstand deposits, while sequences II and III tend to be thickest near lowstand shorelines developed on the mid-shelf.

Glaciation has influenced the region through glacial-isostasy and sediment supply changes. The northward deepening of reflection horizons and paleo-shorelines in sequences I and IV reflects the collapse of a peripheral bulge that developed south of the ice sheet. The peripheral bulge may also account for a shift in the course of the Hudson River after deposition of sequence III. The thickest accumulations of sediment are associated with the lowstand and early transgressions in sequences I and IV because of sediment shed from the Laurentide Ice Sheet directly by meltwater, or indirectly by rivers flowing over a newly deglaciated landscape. Sediment supply was also substantial during sequence III, perhaps because of glacial or paraglacial activity in the Hudson drainage basin during stages 4 to 3.

Because direct evidence of glaciation is absent from the study area, the sediment supply, relative sea-level and peripheral bulge effects recorded in the New Jersey shelf stratigraphy could be mistaken for tectonic effects. Only the exceptionally high frequency of the sea-level cycles, the extreme concentration of sediment supply during lowstands and early transgressions, and the correlation of the relative sea-level cycles with climatic fluctuations, would suggest a possible glacial effect. Geologists studying the stratigraphic records formed during global "icehouse" conditions, such as the Permo-Carboniferous and late Cenozoic, should be aware of the possibility of paraglacial influences on their sections, even in the absence of direct evidence of local glaciation.

CHAPTER SIX:
TIMING OF LATE QUATERNARY DEPOSITIONAL EVENTS ON THE NEW JERSEY MARGIN: IMPLICATIONS FOR REGIONAL GLACIAL AND SEA-LEVEL HISTORY.

6.1 Introduction

Oxygen-isotope records from deep-sea cores (Fig. 2) are generally regarded as proxy records of global ice volume, and therefore, Pleistocene sea level (Shackleton, 1967; 1987), but the precise timing and elevations of late Pleistocene sea levels remain controversial. The elevation of highstands during isotope stage 3 (or the "middle Wisconsin") is particularly contentious; some authors have argued for maximum elevations near or even above modern levels (e.g. Curray, 1965; Milliman and Emery, 1968). Others (e.g. Thomas, 1992) have contended that sea levels never exceeded -40 m during stage 3, based on correlation to the Gulf Coast and isotopic evidence. However, estimates of sea level amplitudes from oxygen isotopes are strongly dependent on temperature assumptions.

Due to the poor exposure and discontinuous preservation of deposits predating the late Wisconsin (isotope stage 2), the extent of glaciation in the northeastern United States during stages 5-3 (the eowisconsin, early, and middle Wisconsin"; Fig. 2) is also uncertain (Muller and Calkin, 1993). Some workers (e.g. Koteff and Pessl, 1985) interpreted the "Lower Till" in New England as resulting from an early Wisconsin glacial advance. Similar ages have been suggested for tills beyond the late Wisconsin limit in northern New Jersey and Pennsylvania (Crowl and Sevon, 1980) (Fig. 5). However, other authors have regarded these tills as stage 6 or older (e.g. Ridge et al. 1990; Oldale and Colman, 1992; Stanford, 1993).

Both eustasy and glaciation in the northeastern United States are reflected in the Quaternary stratigraphy of the New Jersey continental shelf. Continental glaciation impacted on the shelf through sediment supply changes and isostatic adjustment (Chapters 4 and 5). Thus, analysis of the late Quaternary stratigraphy of the region on the periphery of the Laurentide Ice Sheet will shed light on both late Quaternary sea level and the regional glacial history.

The objectives of this study are to use the sequence stratigraphic framework for the late Quaternary of New Jersey continental shelf developed in Chapter 5 to examine the record of glaciation and sea-level. Amino acid geochronology of mollusk shells and biostratigraphy of calcareous nannofossils from the Atlantic Margin Coring Project (AMCOR) drill holes are used to constrain the ages of these sequences. Comparison of the approximate ages and the inferred sea-level elevations during deposition of the sequences to $\delta^{18}O$ from deep-sea cores yields an approximate sea-level history for the last 150,000 years. The sea-level history and stratigraphic events recorded in the stratigraphy of the New Jersey shelf have significant implications for the glacial history of the adjacent continent.

6.2 Geologic Setting

The study area is the NE-SW trending segment of the U.S. Atlantic Coast continental shelf between the Hudson Shelf Valley and the Delaware Shelf Valley from 38°40' to 40° 30'N and 72°30' to 74°40'W (Fig. 1), an area of approximately 25000 km^2. The continental shelf of New Jersey is broad (120-150 km) and gently sloping (< 0.001). The steepening associated with the shelf break begins between 120 and

160 m water depth. The shelf is generally classified as storm dominated, and has a mixed energy shoreline, with tidal range of 1-2 m and mean significant wave heights of roughly 1 m.

The region is part of the Baltimore Canyon Trough, a passive margin with a hinge line running essentially parallel to the shoreline approximately 20 km offshore south of Barnegat Inlet (Grow et al., 1988). Farther north, the hingeline trends parallel to Long Island. Long-term regional subsidence is low, diminishing from near zero landward of the hinge line to about 0.015 mm/a on the upper slope (Greenlee et al., 1988). The Hudson River has been the primary sediment source for the area throughout the Quaternary (Poag and Sevon, 1989).

During isotope stage 2, the Laurentide ice sheet advanced as far south as Long Island and northern New Jersey (Fig. 2; Dyke and Prest, 1986; Stanford, 1993). The weight of the ice sheets caused substantial isostatic adjustment of the crust in the region. Estimates of the extent of isostatic adjustment from geophysical models appear to conflict with observed relative sea-level data from the mid-Atlantic. Existing isostatic models have generally predicted significant isostatic adjustment as far south as North Carolina (Clark et al., 1978; Tushingham and Peltier, 1992), but evidence from the Atlantic Coast suggests a more geographically limited effect. Dillon and Oldale (1978) demonstrated that late Wisconsinan shorelines preserved on the shelf were essentially horizontal from North Carolina to a "hinge zone" running southeastward from Barnegat Inlet, New Jersey. They suggested that the areas north and south of that hinge zone were on different crustal blocks that responded largely independently to ice loading. Ice loading resulted in relative subsidence of the northern half of the study area since 15 ka (Dillon and Oldale, 1978). Peltier (1994) unveiled a new model in which the Laurentide Ice Sheet is 35% thinner than in his previous models, which may address this problem, but a new relative sea-level curve for sites in the region has not yet been published.

6.3 Sequence Stratigraphic Framework

Chapter 4 describes the stratigraphy of the New Jersey shelf above a regionally extensive unconformity identified as the base of sequence I. Three other regional unconformities are interpreted as subaerial erosion surfaces, defining four sequences. All four sequences lie above the lowest occurrence of the calcareous nannofossil *Emiliania huxleyi*, and are therefore younger than about 275 ka (stage 8) (Thierstein et al., 1977; Berggren et al., 1980, Fig. 2). The regional geologic history inferred from this stratigraphy is summarized in Figure 22.

The lowermost sequence, I, is a thin, discontinuous sheet on the middle to outer shelf, and thickest (up to ~30 m) on the outermost shelf and upper slope. This is interpreted the thick outer shelf sediments as a lowstand to early transgressive wedge developed when the shoreline was near the shelf edge (Chapter 5, Fig. 22a). A broad swale in the sequence boundary that extends southward from the Hudson Shelf Valley implies that the Hudson River may have flowed southward across the shelf to the Spender Canyon area (Fig. 1). Although the unit cannot be traced onto the inner shelf because of interference by the water-bottom multiple, parts of the Cape May Formation in southern New Jersey may represent the highstand deposits of sequence I (Fig. 22b).

Sequence II is largely confined to the broad swale in the underlying sequence I; it is thin or absent elsewhere on the shelf. Within this swale on the mid-shelf are two packages of clinoforms that make up the lower part of a thick (up to 30 m) triangular wedge of sediment that are interpreted as a Paleo-Hudson delta (Chapter 5, Fig. 22c). Transgressive deposits of this unit can be traced onto the inner shelf to a buried shoreline 3 km offshore of Barnegat Inlet (Fig. 22d) found by Ashley et al. (1991). The base of sequence III is marked by two prominent valleys: a paleo-Hudson valley identified by Knebel et al. (1979) extending southward from the Hudson Shelf valley to the mid-shelf, and a second valley extending westward from South Toms Canyon (Fig. 3) toward the seaward end of the paleo-Hudson (Fig. 22e). The second valley was interpreted as a headward eroding extension of the submarine canyon on to the shelf resulting from a combination of high sediment input in the vicinity of the Hudson, and lower sea-level. The transgressive deposits of this sequence do not extend as far landward as those in the underlying sequence II (Fig. 22f). The Hudson River did not re-incise this valley, indicating that it switched to the modern shelf valley position prior to the sequence IV lowstand.

The uppermost sequence, sequence IV, is similar to sequence I in that it is generally thin on the inner to middle shelf, and thickens to as much as 50 m on the outer shelf and upper slope. The thick shelf-edge deposits are interpreted as lowstand to early transgressive deposits. Subsequent transgression reworked the shelf sediments into a thin modern transgressive shelf blanket described by Swift et al. (1972), except for a 10-15 m thick wedge of sediment on the mid-shelf deposited during the last transgression (Milliman et al., 1990).

6.4 Data Sources and Methods

The seismic data used were collected on several USGS cruises between 1974 and 1982 (Fig. 4). The principal cruise data used were the Atlantis II Leg 2 lines 12-14, 16-18, and 20-22 that comprise about 750 km of Uniboom™ data collected in 1975. An additional 300 km of Uniboom data from the 1976 Mount Mitchell cruise, and about 120 km of 3.5 kHz and 8 kJ Mini-sparker data collected on the Atlantis II cruise, provided tie-lines in the mid-shelf region. On the outer shelf, 150 km of the Fay 021 8 kJ Mini-sparker data collected in 1974 and the seismic reflection stratigraphy developed from sparker data by Harris (1983) were examined. The Atlantis II data was tied to the Barnegat Inlet stratigraphy established by Ashley et al. (1991) through use of about 250 km of Uniboom™ and 3.5 kHz data collected on the Whitefoot cruise in 1981.

Although there are large numbers of Vibracores penetrating the upper 6-9 m of the sediment, particularly on the inner shelf, and several wells drilled by oil companies have cored deeper deposits, groundtruth for the upper Pleistocene section below the surficial sediments is limited to the five holes drilled by the Atlantic Margin Coring Project (AMCOR) (Fig. 4.). Poor recovery rates (rarely more than 30% in the upper Pleistocene) limit the usefulness of the cores for lithostratigraphy (Hathaway et al., 1976), but the cores have provided material for determining the approximate age of the deposits through foraminiferal and calcareous nannofossil biostratigraphy (Poag in Hathaway et al., 1976; Valentine in Harris, 1983), palynostratigraphy and aminostratigraphy (Groot et al., 1995). Additional amino acid results were obtained for two samples from core 6010, and three samples each from core 6009 and 6020.

Shell material was extracted from the sediment, rinsed with distilled deionized water, and placed in heat-treated glass vials. The samples were then sent to the University of Delaware, where the D-alloisoleucine/A-isoleucine ratios were obtained through the liquid chromatography method described in Wehmiller (1984).

6.5 Results

Figures 23, 24, and 25 are line diagrams showing the principal reflection horizons seen on the shore-perpendicular Atlantis II lines 2-14, 2-16, and 2-20, respectively (Fig. 4). Line 2-14 is the longest continuous seismic line and extends from the inner shelf to Toms Canyon, lying north of the Dillon and Oldale (1978) "inflection zone" that separates areas affected by the peripheral bulge of the Laurentide Ice Sheet from areas experiencing little glacial-isostasy. Line 2-16 lies 15 to 20 km south of Line 2-14 and provides a tie-line for AMCOR sites 6010 and 6020. It is also along the same roughly shore-perpendicular transect as holes 6011 and 6021 (Fig. 4). Line 2-20 provides good data for the middle to outer shelf in southern New Jersey, running offshore from Great Egg Harbor and tying in AMCOR site 6009. The reflection horizons from which sea-level estimates may be made are discussed below. Note that no corrections have been applied for subsidence to these sea-level elevations.

6.5.1 Sea Level Indicators

6.5.1.1 Sequence I. If the shelf edge sediments seen along 2-14 (Fig. 23, location A) represent coastal onlap of a lowstand wedge they imply that relative sea-level could have been as deep as -190 m during the lowstand of this sequence. On line 2-20 part

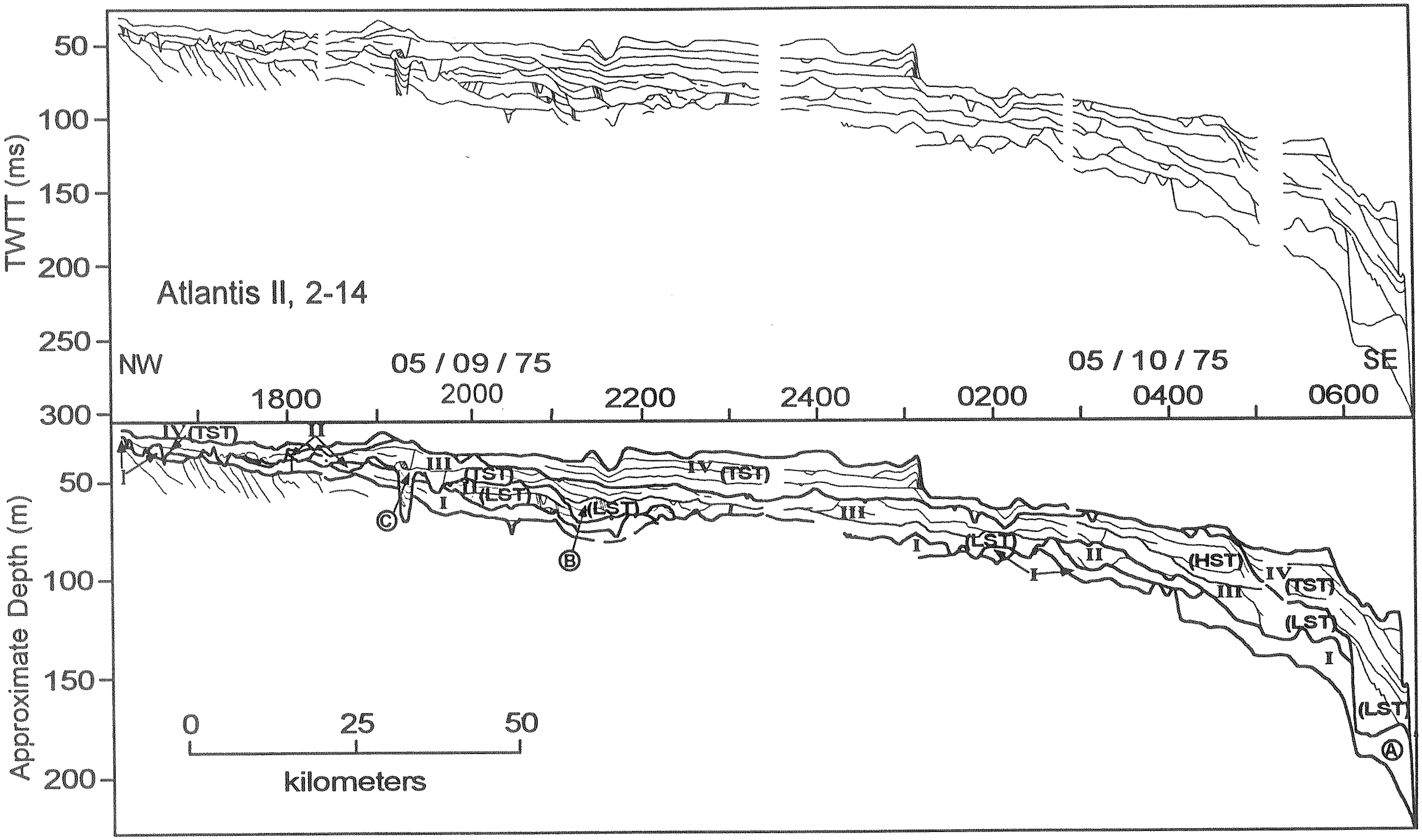


Figure 23. Seismic-stratigraphic Section Along Line 2-14. An uninterpreted line diagram of prominent reflectors seen in the data (above), and the sequence stratigraphic interpretation (below). See Fig. 4 for track line. Depth scale based on velocity of sound in water (1500 m/s). Date and times of data collection are indicated at top. Letters indicate locations mentioned in text.

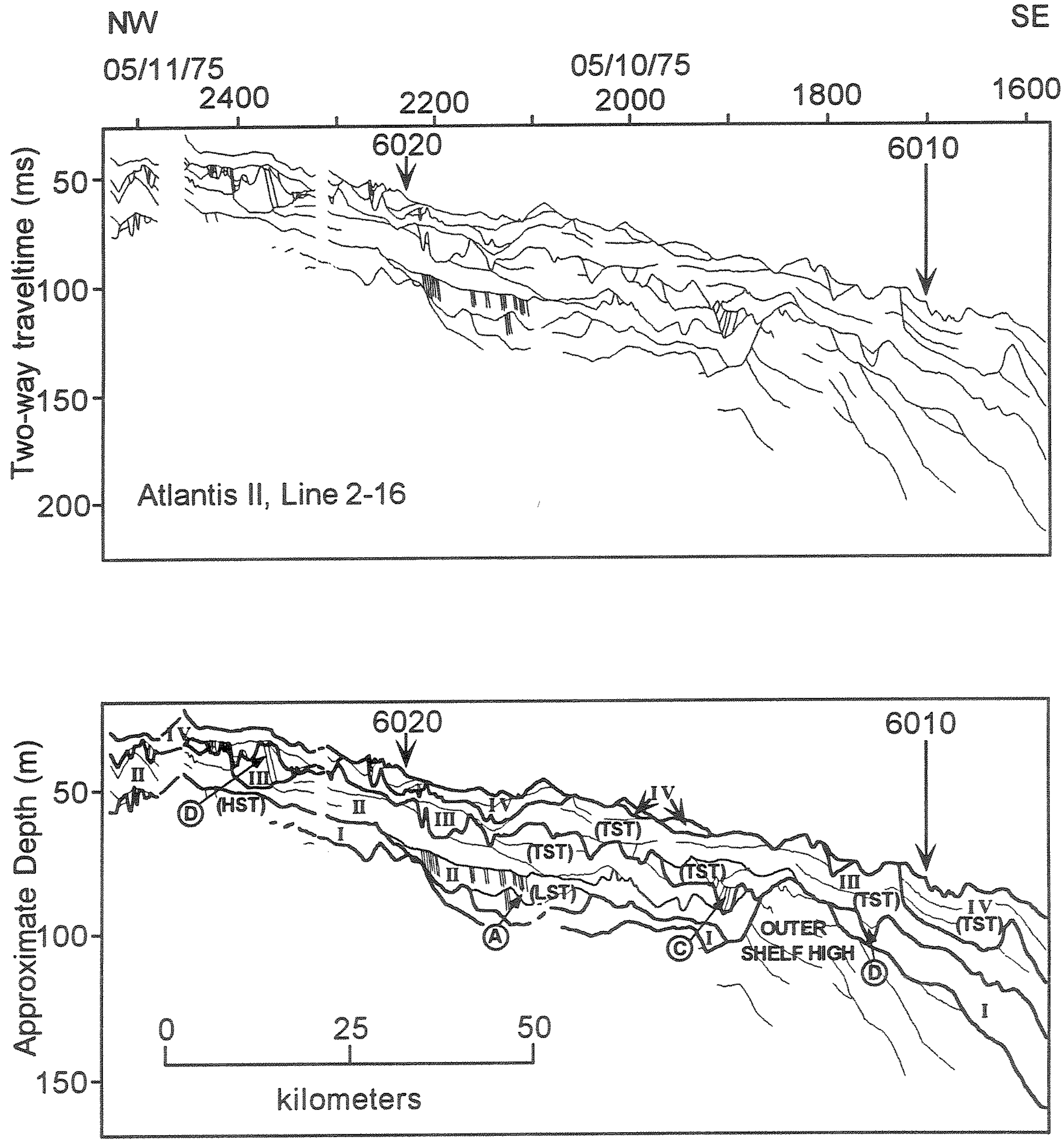


Figure 24. Seismic-stratigraphic Section Along Line 2-16. An uninterpreted line diagram of prominent reflectors seen in the data (above), and the sequence stratigraphic interpretation (below). See Fig. 4 for track line. Depth scale based on velocity of sound in water (1500 m/s). Date and times of data collection are indicated at top. Letters indicate locations mentioned in text.

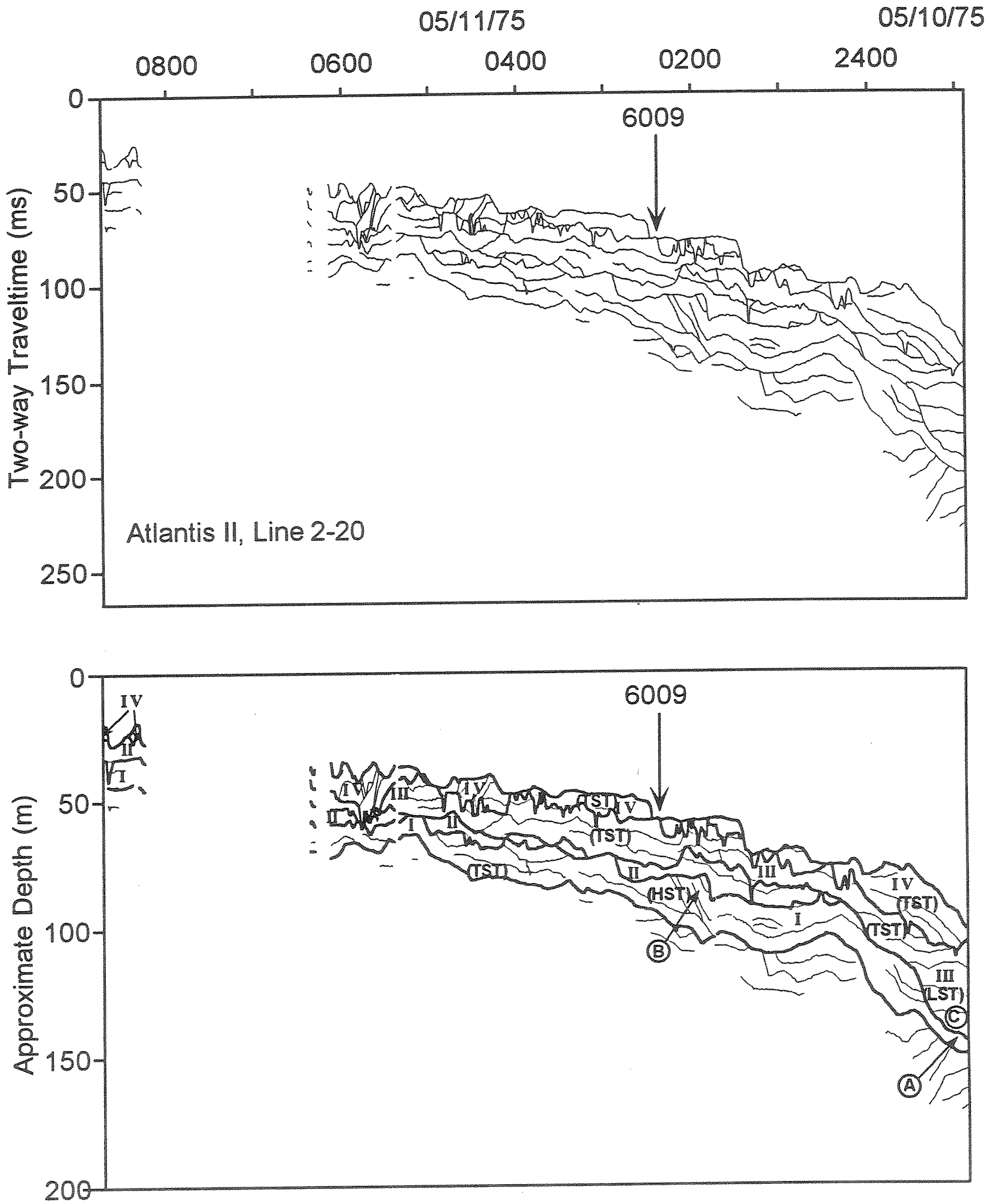


Figure 25. Seismic-stratigraphic Section Along Line 2-20. An uninterpreted line diagram of prominent reflectors seen in the data (above), and the sequence stratigraphic interpretation (below). See Fig. 4 for track line. Depth scale based on velocity of sound in water (1500 m/s). Date and times of data collection are indicated at top. Letters indicate locations mentioned in text.

of this onlapping wedge can be seen at a depth of approximately 103 m (Fig. 25, location A). This is probably not the lowest lowstand position (as the line does not extend to the shelf edge), but indicates that sea-level must have dropped to at least this elevation.

The highstand elevation cannot be determined from the shelf data, but must have been shallower than -30 m, where the unit becomes lost within the multiple on the inner shelf. However, if this sequence is stage 5e, as proposed in Chapter 4, it could include part of the Cape May Formation in southern New Jersey, which was deposited when sea-level rose as much as 5 to 10 m above modern levels (Mixon et al., 1974). The upper surface of a possible highstand wedge is seen on the outer shelf along line 2-20 (Fig. 8, location B) that is truncated by the overlying sequence boundary at an elevation of ~-84 m, suggesting sea-level dropped to that elevation during the ensuing lowstand.

6.5.1.2 Sequence II. The lowstand deltaic lobe on the mid-shelf in this sequence (Fig. 22c) contains two packages of clinoforms (Fig. 24, location A). The prominent, nearly planar surface between the two packages could be a marine flooding surface, making these two lowstand parasequences. An alternative explanation is that this surface actually represents a sequence boundary. The absence of transgressive and highstand deposits associated with the lower package could simply be a result of the erosion of these deposits. This would be particularly likely if the sea-level fluctuation was of short duration, as the resulting deposits would then likely be thin.

The most seaward of the clinoforms is truncated by the transgressive surface approximately 78 m below modern sea-level on line 2-16 (Fig. 24, location A).

Assuming that some sediment (10 m?) was eroded during the transgression, as estimated for the Delaware shoreline (Kraft et al., 1987), and that the delta front was near sea-level, this suggests that the lowstand was at least 68 m below modern levels.

The most widespread unit in this sequence is above this transgressive surface. The nearly horizontal onlapping reflections within in this unit, and absence of prograding clinoforms, suggest that this unit belongs to the TST. This TST is downlapped by the Ashley et al. (1991) highstand shoreline that implies a highstand of approximately -20 m.

6.5.1.3 Sequence III. The lowest elevation of onlap of this lowstand wedge to the sequence boundary is approximately -90 m along both lines 2-16 (Fig. 24, location B) and 2-20 (Fig. 25, location C), giving a deeper limit for the lowstand of this sequence. There is also a package of prograding clinoforms, similar to those in the underlying sequence II, on line 2-14, that could represent a lowstand shoreline (Fig. 23, location B) at approximately -68 m. However, this shoreline is approximately 25 km landward of the seaward end of the Paleo-Hudson Valley (Fig. 24, location C), so it seems unlikely to represent the maximum lowstand shoreline unless the shoreline configuration was very unusual.

The seaward end of the Paleo-Hudson Valley is characterized by apparently landward-inclined reflections that could result from southward progradation of a baymouth spit or progradation of a flood-tidal delta into an estuary (Chapter 4), in prep.). In either case, the valley fill occurred during the transgression, and the sharp overlying reflection likely stems from a wave-cut ravinement surface at approximately -84 m below modern sea-level. Assuming that the wave-cut surface eroded to a depth of 5-10 m below sea-level, sea-level must have dropped below -74 m.

Where line 2-14 crosses the valley (Fig. 23, location C), no well-developed baymouth spit or flood tidal delta complex is present, suggesting that the baymouth did not reach this region during deposition of sequence III. However, the presence of marine shell deposits in the upper section (Knebel et al., 1979) implies that sea-level reached at least -29 m. A series of prograding reflections near the landward limit of preservation of this sequence on line 2-16 may represent a highstand shoreline (Fig. 24, location D) at an elevation of at least -34 m.

On line 2-14, the downlapping reflections on the outer shelf are may be the only preserved part of the highstand from this unit. Their presence reflects either a high enough sediment supply to allow preservation of deposits during fall sea-level, or a prolonged still-stand at an elevation of approximately -75 m.

6.5.2 Age Constraints on the New Jersey Shelf Sequences

All four sequences are known to lie within the calcareous nannofossil *Emiliania huxleyi* in AMCOR 6010 (Harris, 1983), which first appeared during stage 8, ca 275 ka (Thierstein et al., 1977; Berggren et al., 1980). Additionally, there are radiocarbon dates for surficial and near-surficial material in many areas of the shelf. Knebel and Spiker (1977) reported several radiocarbon dates of > 40 ka from surficial sediments on the outer shelf just seaward of the Mid-Shelf Sediment Wedge that probably belong to Sequence II. Other radiocarbon ages from this unit ranging from 28 ka to > 40 ka were reported by Knebel et al. (1979) and Davies et al. (1992).

The base of Sequence IV corresponds to the R-reflection of McClennen (1973). The lower boundary of this sequence is reasonably well-constrained by radiocarbon dates, and is believed to correspond to the stage 2 lowstand. Coaly organics in the "strandplain" deposits described by Snedden et al. (1994), associated with this sequence boundary, produced radiocarbon ages ranging from 18,770 to 25,010 yrs. BP. Although Knebel et al. (1979) considered the Paleo-Hudson Valley to be part of the R-reflection, Harris (1983) showed the R-reflection to be stratigraphically higher, a finding which is consistent with our study, and with the ~30 ka radiocarbon dates on material within the valley fill.

The only other chronological data that can distinguish between material older than 40,000 years (ie.beyond the limit of radiocarbon), and younger than stage 8 (FAD of *Emiliania huxleyi)* available from these sequences are amino acid racemization ratios. D-alloisoleucine / L-isoleucine ratios from mollusk shells, which are summarized in Table IV. The amino acid ratios from sites 6009, 6010 and 6020 all show consistent upward decreases in D-alloisoleucine / L-Isoleucine ratio, with the exception of the shell fragment from core 2-3 in 6020. The exceptionally high value for that fragment could result from reworking of older deposits, or it may be from a species that racemizes particularly quickly. The general upward decrease suggests that the amino acid data may be useful for relative age dating within the area, and that absolute age estimates based on the *Mercenaria* aminozones for the Atlantic Coastal Plain (Wehmiller et al., 1992) may yield reasonable results (Fig. 2).

Fragments of shells of unknown origin from approximately 20 m below the

Table IV: Amino Acid Results from Upper 50 m of AMCOR Sites 6009, 6010 and 6020

AMCOR Site	Core No.	Sequence	Genus	Lab No.	Sample No.	A/I Area	A/I Height
6009	1-1	IV	unknown	950106	jw95-005-1	0.02	0.04
6009	1-1	IV	unknown	950107	jw95-005-2	0.05	0.06
6009	2-1	II	*Astarte?*	950103	jw95-002-1	0.20	0.26
6009	2-1	II	unknown	950104	jw95-002-2	0.32	0.36
6009	6-1	below I	unknown	950100	jw95-001-1	0.66	0.79
6009	6-1	below I	unknown	950100	jw95-001-1	0.67	0.75
6009	6-1	below I	unknown	950101	jw95-001-2	0.62	0.68
6010	5-1	below I	unknown	900324	89-187-1	0.29	0.32
6010	5-1	below I	*Cyclocardia?*	900325	89-187-2	0.27	0.33
6010	5-1	below I	*Cyclocardia?*	900326	89-187-3	0.35	0.40
6010	7-4	below I	*Crassostrea?*	950105	jw95-003	0.59	0.63
6020	2-3	III	*Mercenaria* or *Spisula*?	950109	jw95-006	0.27	0.35
6020	3-2	II	*Astarte*	930271	JW93-171	0.18	
6020	5-1	I	unknown	910089	jw89-189-1	0.21	0.25

Data for lab numbers less than 920000 are from Groot et al. (1995); all others are courtesy of John Wehmiller (pers. comm., 1995). Refer to Section 8.1 for sample elevations.

boundary in hole 6009 gave ratios ranging from 0.62 to 0.67, comparable to those from a deeper unit (core 7-4) in 6010. If these shells have similar racemization rates to *Mercenaria*, they would be placed in aminozone IId', suggesting ages of at least 500 ka, possibly even early Pleistocene (Wehmiller, pers. comm., 1995). Higher in core 6010, above the *Pseudoemiliania lacunosa* last appearance datum (LAD ~ 474 ka, Berggren et al., 1980), amino acid ratios in possible *Cyclocardia* samples from just below Sequence I in hole 6010, gave D-Alloisoleucine / L-Isoleucine ratios of 0.27 to 0.35 (Groot et al., 1995). These could fall within aminozones IIc (stage 7-9?) or IId (stages 11 to 15?) (Groot et al., 1995); these results are consistent with the biostratigraphy in core 6010. An unknown shell from just above the boundary in core 6020 gave an amino acid ratio of 0.21, which could represent a stage 5 age (aminozone IIb, Fig. 2), if the shell racemizes at a rate comparable to *Mercenaria* (Groot et al., 1995).

Amino acid ratios have been obtained for three samples from Sequence II. A whole *Astarte* shell in good condition at AMCOR site 6020 gave a D/L ratio of 0.18, while a possible *Astarte* fragment from Sequence II at site 6009 produced a ratio of 0.20. A shell fragment of unknown genera from the same core gave a ratio of 0.32. The racemization kinetics of *Astarte* are relatively poorly constrained. An *Astarte* from the Delaware shelf that produced a ratio of 0.14 was radiocarbon dated at > 49 ka; using parabolic kinetics, this suggests an age > 85 ka for ratios of 0.18 to 0.20 (Wehmiller, pers.comm., 1995). The higher ratio recorded from the other sample in 6020 may reflect the presence of a faster racemizing species, or a shell fragment reworked from older material.

The sole amino acid result from Sequence III comes from site 6020, where a shell fragment, possibly a *Spisula* or *Mercenaria* gave a ratio of 0.27, higher than any of the results from deeper units at the same site. The shell fragment may be reworked.

6.6 Discussion

6.6.1 Age Models for Sequences

Two age models have been proposed for the sequences observed in the study area. Both models interpret sequence I as stages 6/5e, and sequence IV as stages 2/1 (Fig. 2), but differ in their inferred timing of sequences II and III. Sheridan et al.(1993) and Carey et al. (1995a) suggested that sequence II represented isotope stages 4 to early 3, and sequence III was deposited during a sea-level fluctuation within stage 3; this model will be called the "short" chronology. Carey et al. (1995b; 1996) raised the possibility of a "medium" chronology, in which sequence II was deposited during a stage 5 sea-level fluctuation, and that the stage 4 lowstand was represented by sequence III.

A third, "long" chronology, based on the observation that neither the biostratigraphic data nor radiometric dates preclude a stage 8 age for the base of sequence I, could also be advanced. This hypothesis suggests that sequence III represents the stage 6/5e transition and sequence II a fluctuation within stage 7 (Fig. 2). Assuming that sequence boundaries are cut at approximately the time of maximum rate of sea-level fall, correlation to the SPECMAP oxygen isotope curve (Imbrie et al., 1984), gives the inferred timing for sequences and systems tracts for each of the three models summarized in Table V:

Table V: Age Models for Sequences

Sequence	Stratigraphic Feature	Short Chronology	Medium Chronology	Long Chronology
IV	Base	30 ka	30 ka	30 ka
III	Highstand	40 ka	55 ka	125 ka
III	Lowstand	45 ka	65 ka	150 ka
III	Base	50 ka	70 ka	160 ka
II	Highstand	55 ka	80 / 100 ka	170 ka
II	Lowstand	65 ka	90 / 110 ka	185 ka
II	Base	70 ka	95 / 115 ka	190 ka
I	Highstand	125 ka	125 ka	200 ka
I	Lowstand	150 ka	150 ka	270 ka
I	Base	160 ka	160 ka	280 ka

The short chronology corroborates the findings of Ashley et al. (1991) and Wellner et al. (1993) that the -20 m shoreline near Barnegat Inlet correlates to the early stage 3 New Guinea coral terrace IV (Bloom et al., 1974), and the subsequent filling of the Knebel Paleo-Hudson valley during sequence III occurred later in stage 3, perhaps in stage 3.1. This is consistent with the biostratigraphy and radiocarbon ages, although it would require placing the prograding wedge seen on the outer shelf on line 2-14 within the lowstand. The amino acid racemization data suggest an age of at least stage 5 for sequence III, but in view of the uncertainties of the racemization kinetics of *Astarte*, and the variable quality of the specimens analyzed, this is inconclusive.

The medium chronology correlates the lowstand deposits in sequence III to stage 5.4 (5d) or 5.2 (5b) . A possible interpretation is that the lower "parasequence" is stage 5d, and the upper one 5b, and stage 5c deposits are absent. This places the Barnegat Inlet shoreline in stage 5a or 5c as proposed by Thomas (1992), and contemporaneous with the Delaware shelf shorelines recognized by Toscano and York (1992). The early stage 3 peak represented in New Guinea (Bloom et al., 1974) then correlates with the Knebel Paleo-Hudson valley fill. This interpretation fits all available age data, but suggests that no record of a subsequent 38-40 ka highstand is preserved on the New Jersey shelf.

The long chronology, shown in Table V, also fits all the available age data; if the fragment above sequence I from AMCOR site 6020 is from a slower racemizing species than *Mercenaria*, a 0.21 ratio could represent a stage 7 deposit. However, there are stratigraphic objections to this interpretation. The stage 6 lowstand was apparently of similar magnitude and significantly more prolonged than the stage 2 lowstand (Fig. 2). The more extended period of exposure to subaerial erosion would make it more difficult to preserve underlying deposits, making it less likely that the record of stage 7 sea-level would be preserved. Furthermore, the Paleo-Hudson valley in the sequence III surface is much narrower than the broad swale in the sequence I surface. This suggests that the Hudson River was fixed throughout the lowstand, which seems more likely if it was of short duration like those in stages 5 and 4. Finally, while it is conceivable that the valley extending westward from Toms Canyon was actually carved by the Hudson rather than shelf edge processes, no clear valley could be seen joining them on Atlantis II leg 3 lines 17-19. This, combined with the thick middle-outer shelf deposit, so different from sequence IV, suggests that during deposition of sequence III, sea level probably did not reach the shelf edge.

In view of these objections to the long chronology, it appears less likely than the other two, and will not be discussed further. The other two age models will be used to develop subsidence-corrected New Jersey sea-level histories for comparison to other records of late Quaternary sea level.

6.6.2 Estimated Rates of Subsidence

The simplest approach to estimating late Pleistocene subsidence is to assume that the topography of the continental shelf was similar at times of similar sea-level elevation. If sea level during stage 6 was approximately the same as during stage 2, as implied by the SPECMAP $\delta^{18}O$ curve (Fig. 2), and the elevation of the shelf during the cutting of sequence boundary I was similar to that during the incision of sequence boundary IV, approximately 130,000 years later, the thickness of sediment between the two sequence boundaries approximates the amount of subsidence that took place over those 130,000 years.

Along the two most southerly lines, which are least influenced by Hudson River migration and isostatic adjustment to the weight of the ice, the thickness of sequences I through III increased in an approximately linear fashion by 35 m between the hinge line to the shelf edge, implying a rate of subsidence of ~0.25 mm/yr at the shelf edge. This is extremely fast compared to the long-term subsidence rates reported by Greenlee et al. (1988) for the outer shelf at COST-B2, where subsidence has averaged only 0.015 mm/yr since the late Cretaceous. However, the upper slope was sediment-starved during much of the Tertiary. In contrast, sedimentation rates have been exceptionally high during the Pleistocene (Poag and Sevon, 1987). Christensen et al.

(in press) indicate that at ODP sites 902 and 903, on the continental slope, 110 m of sediment were deposited during stages 12 through 6, a mean sedimentation rate of approximately 0.3 mm/a. This is even greater than sedimentation rates during the mid-Miocene (Mountain et al., 1994), when subsidence rates were approximately 0.05 mm/a, largely due to isostatic loading by sediment (Greenlee et al., 1988). The rather high rates of subsidence suggested by the stratigraphy may partly reflect aggradation of the shelf to higher elevations in stage 2 than in stage 6. However, isostatic loading by sediment should have resulted in subsidence rates substantially higher than the Tertiary average for the region. For the purposes of this study, I will assume that subsidence increases linearly from 0 at the hinge line to 0.25 mm/a at the shelf edge.

6.6.3 Sea-Level History

Assuming that the rate of subsidence increases linearly from 0 at the hinge line to 0.25 mm/yr at the shelf edge, and using the short and medium chronologies (Table V), two possible sea-level histories accounting for the sequence stratigraphy observed were constructed (Table VI). Sea-level indicators fall into three categories: onlapping lowstand wedges, prograding shorelines, and valley fill deposits. The deepest onlapping reflection horizon in a lowstand wedge was used to estimate an approximate maximum depth for the lowstand.

The shallowest reflector, where the wedge is truncated by the transgressive surface, provides an estimate of sea-level at the close of the lowstand. Kraft et al. (1987) found that the erosional part of the shoreface on the Delaware shoreline is approximately 10 m high. Therefore, assuming that erosion during transgression is primarily a result of shoreface retreat, and shoreface conditions were similar to those

Table VI: Sea-Level Estimates from the New Jersey Margin

Sequence / Systems Tract	Feature	Sea-Level Estimate	Rate of Subsidence	Model 1 Sea-Level and Age	Model 2 Sea-Level and Age
IV - TST	Mid-Shelf Wedge, Line 2-16	> - 40 m	0.14 mm/a	11.5 ka ; > - 38 m	11.5 ka ; > - 38 m
IV - TST	Onlapping Wedge, Line 2-20	< - 67 m	0.20 mm/a	14 ka ; < - 64 m	14 ka ; < - 64 m
IV - TST	Onlapping Wedge, Line 2-14	< - 78 m	0.21 mm/a	14 ka ; < - 75 m	14 ka ; < - 75 m
IV - LST	Onlapping Wedge, Line 2-14	> -129 m	0.24 mm/a	20 ka ; > -124 m	20 ka ; > - 124 m
III - HST	Prograding Shoreline, Line 2-14	> - 80 m	0.21 mm/a	45 ka?; > - 70 m	45 ka ; > - 70 m
III - HST	Prograding Shoreline, Line 2-16	> - 34 m	0.06 mm/a	38 ka ; > - 32 m	55 ka ; > - 31 m
III - TST	Knebel Valley Fill, Line 2-14	> - 29 m	0.02 mm/a	38 ka ; > - 28 m	55 ka ; > - 28 m
III - TST	Baymouth Complex, Line 2-16	> - 77 m	0.18 mm/a	40 ka ; > - 70 m	60 ka ; > - 66 m
III - LST	Onlapping Wedge, Line 2-20	< - 88 m	0.21 mm/a	45 ka ; > - 78 m	70 ka ; > - 73 m
III - LST	Onlapping Wedge, Line 2-16	< - 90 m	0.19 mm/a	45 ka ; > - 74 m	70 ka ; > - 78 m
III - LST	Prograding Shorleine, Line 2-14	< - 58 m	0.06 mm/a	45 ka ;< - 55 m	70 ka ; < - 54 m
II - HST	Barnegat Inlet Shoreline	> - 20 m	0.00 mm/a	55 ka ; > - 20 m	80 ka ; > - 20 m
II - LST	Prograding Shoreline, Line 2-16	< - 68 m	0.11 mm/a	70 ka ; < - 60 m	90 ka ; < - 58 m
II - base	Truncates clinoforms, Line 2-20	> - 84 m	0.17 mm/a	75 ka ; < - 71 m	115 ka ; < - 64 m
I - TST	Onlapping wedge, Line 2-20	> - 103 m	0.21 mm/a	135 ka ; > - 75 m	135 ka ; > - 75 m
I - LST	Onlapping wedge, Line 2-14	> - 184 m	0.24 mm/a	150 ka ; > -148 m	150 ka ; > - 128 m

experienced in the late Holocene,the elevation of the most seaward clinoform in lowstand prograding shorelines indicates a position ~10 m below the deepest sea-level position during the lowstand. The most landward clinoform in highstand shorelines gives a shallowest sea-level elevation for the highstand sea level. Where valley fill deposits are truncated by wave-cut ravinement surfaces, they imply sea levels 10 m above that elevation at the time of ravinement, assuming no tidal ravinement subsequent to shoreface transgression. Where valley fill deposits are truncated by younger sequence boundaries, they still give the deepest possible elevation of sea-level at the time of valley filling.

A partial list of the sea-level indicators used is given in Table VI. According to both age models, during stage 6, sea level fell below -75 m, and may have fallen to as low as -148 m on line 2-14. However, the great depth of the later figure could reflect collapse of a peripheral bulge developing in response to a glaciation of similar extent to the late Wisconsin. Sea-level then rose to 5-10 m above sea-level during stage 5e, as recorded by the Cape May shoreline (Mixon et al., 1974).

According to the medium chronology, sea-level dropped to at least -64 m at least once during stage 5, then rose to the -20 m Barnegat Inlet shoreline found by Ashley et al. (1991) during stage 5a or 5c. During stage 4 sea level fell to -66 to -78 m, then rose above -28 m during stage 3, filling the Paleo-Hudson valley around 55-60 ka. This medium chronology implies that any subsequent fluctuations in stage 3 sea-level left no preserved deposits on the New Jersey shelf, except for a possible prograding shoreline associated with a stillstand at approximately -70 m.

The short chronology suggests that little record from stage 5 is preserved. Sea level fell below -71 m during stage 4, and the subsequent Barnegat Inlet shoreline preserves the record of the -55 to -60 m highstand, at approximately -20 m. A further sea-level fluctuation is recorded during stage 3, with a lowstand of ~ -70 m, and a highstand that filled the Paleo-Hudson valley to above -28 m, around 38-40 ka. Both models imply that stage 2 sea level likely fell to -120 to -130 m. For some time, sediment supply kept up with sea-level rise, until transgression occurred when sea-level reached -70 to -75 m, at around 13-14 ka (Milliman et al., 1990). A second period of progradation resulted in deposition of the mid-shelf wedge, which is estimated to have formed around 11.5 ka by Milliman et al. (1990). This progradation may partly reflect a slowing of sea-level rise at that time; there is evidence of a reduced rate of sea-level rise between 12 ka and 10 ka from coral reefs in the Barbados (Fairbanks, 1989) and New Guinea (Edwards et al., 1993).

6.6.4 Comparison to Other Records of Sea-Level Change

The principal difference between the two age models is that the medium chronology indicates that high-amplitude sea-level fluctuations occurred within stage 5, with lowstands deeper than -65 m, while the short chronology suggests slightly higher sea levels during stage 3 highstands. The 55-60 ka highstand lies at or above -20 m, according to hypothesis 1, and a later highstand reaches at least -28 m. One way of evaluating the validity of these different sea-level histories is to compare them to other records of sea-level change. The two types of proxy sea-level record that have the best chronological control are oxygen-isotope records from deep sea cores and $^{234}U/^{230}Th$-dated coral reef terraces.

The SPECMAP curve (Fig. 2) is perhaps the most global proxy of ice volume and sea level available, because it is a composite of the oxygen isotopes from benthic foraminifera from deep-sea cores throughout the world. If the relationship between $\delta^{18}O$ and sea level were linear, and sea-level during the stage 2 lowstand was about -120 m, sea level never fell below -45 m during stage 5, or rose above -70 m during stage 3. However, the SPECMAP curve is stacked and smoothed, thereby suffering some loss of resolution and preventing recognition of high-frequency sea-level changes. Furthermore, the $\delta^{18}O$ records from benthic foraminifera on which it is based are influenced by changes in deep water temperature as well as ice volume. These difficulties probably account for the apparent inconsistencies between the SPECMAP curve and the sea-level records from coral terraces (Chappell and Shackleton, 1986).

Perhaps the most widely known of these are the tectonically-uplifted terraces of the Huon Peninsula New Guinea, where a series of terraces ranging in age from stage 5e to 3 has been dated (Bloom et al., 1974; Aharon and Chappell, 1986). The amount of tectonic uplift was estimated by assuming an initial elevation of +6 m for the stage 5e highstand, based on elevations of corals in tectonically stable areas, such as the Bahamas, Hawaii, and Australia (Broecker and Thurber, 1965; Ku et al., 1974; Veeh et al., 1979). This resulted in sea-level elevations during substages 5c and 5a of approximately ($\pm$ 5 to 6 m) -12 m and -19 m, respectively (Fig. 26a). These results are similar to those obtained from studies of reef terraces in Barbados, where sea-level is estimated to have reached -13 to -18 m during 5c, and -12 to -17 m during 5a (Dodge et al., 1983; Gallup et al., 1994), and Haiti, where the elevations were -10 m and -13 m for 5a and 5c, respectively (Dodge et al., 1983). Despite the consistency of these results, there remains some controversy concerning the elevation of stage 5 sea levels, as recent speleothem data from stable platforms such as Bermuda and Florida, and raised shoreline data from California imply sea-level elevations near, or even slightly above, modern sea level at around 80 ka (Mylroie and Carew, 1988; Muhs et al., 1994; Ludwig et al., 1996). These high sea-level elevations are in apparent conflict with the SPECMAP curve, as well as with the Barbados and New Guinea data. Richards et al. (1994) contended that continuous speleothem growth at -18 m during stage 5c in the blue holes in the Bahamas precludes such high sea-level elevations.

The elevations of lowstands during stage 5 are less well-constrained, but appear incompatible with a linear interpretation of the relationship between benthic $\delta^{18}O$ and sea level, particularly during 5d. Steinen et al. (1973) inferred a sea level of -71 m $\pm$ 11 m from the Barbados subsurface, and workers in New Guinea suggest that sea level dropped to at least -60 m at this time, based on the elevation of deltaic gravels (Chappell, 1974; Chappell and Shackleton, 1986; Fig. 26a). The 5b lowstand was apparently not as deep, but still deeper than the SPECMAP curve would suggest, reaching approximately -50 m (Chappell and Shackleton, 1986).

Reef terraces associated with stage 3 highstands are also reported from Barbados and New Guinea. The early stage 3 highstand (reef IV) reached roughly -28 m, and subsequent stage 3 highstands reached -37 to -42 m (Bloom et al., 1974; Aharon and Chappell, 1986; Fig. 9a). The ~ 60 ka and ~35 ka terraces in Barbados (James et al., 1971) have been estimated as representing sea levels near -20 m and -40 m

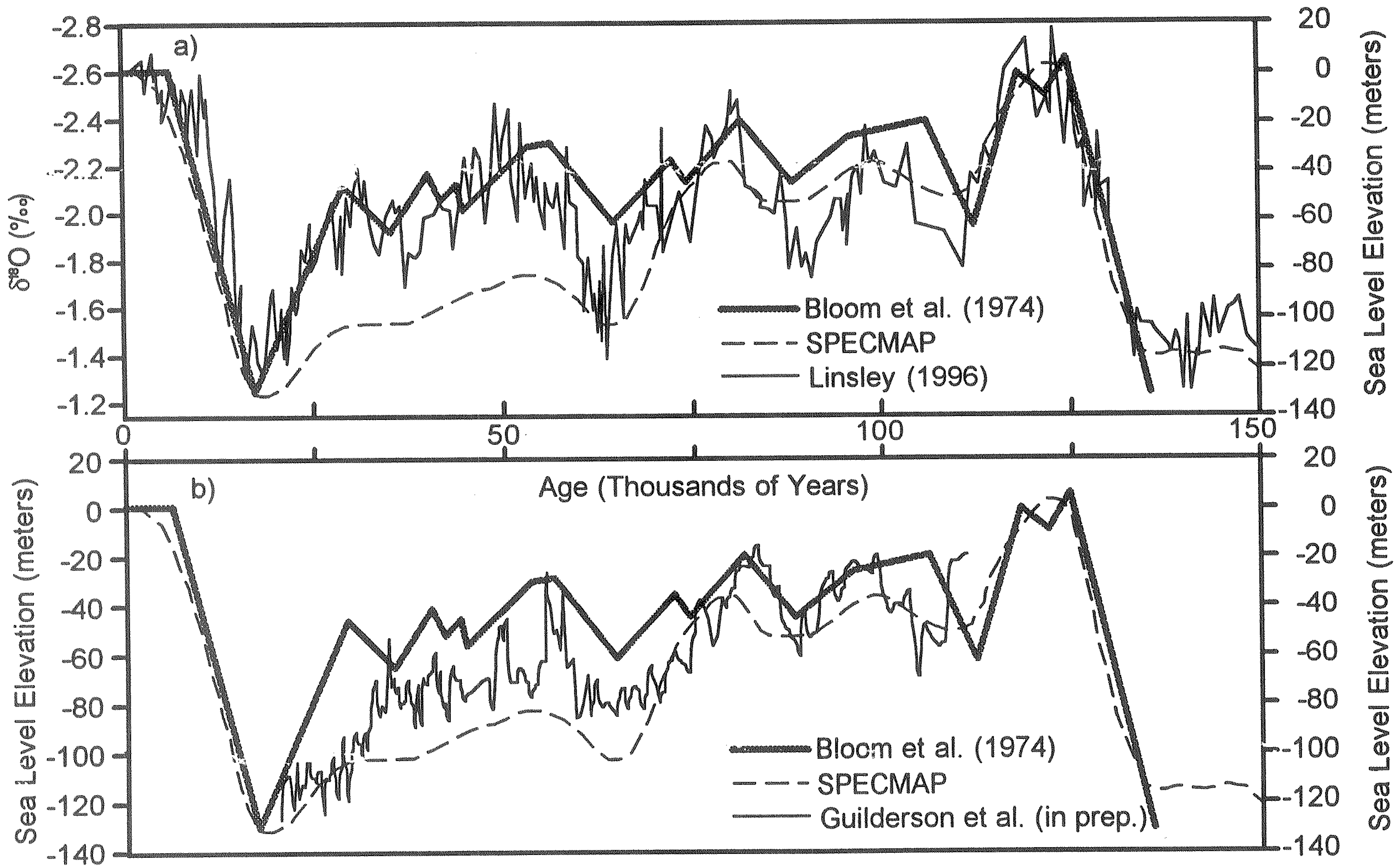


Figure 26. Sea-Level Curves for the Past 150 ka. The Bloom et al. (1974) New Guinea reef data and a SPECMAP curve (Imbrie et al., 1984) scaled to the reef data is compared to a) the Linsley (1996) sea level estimated from Sulu Sea planktonics and b) the Guilderson et al. (in prep.) sea level estimated from Pacific benthics.

respectively. These results are consistent with blue hole speleothem data from the Bahamas that indicate sea level could not have risen above -20 m at 55-60 ka or -25 m at 40 ka (Richards et al., 1994). Some workers have argued for sea levels of -20 m or even exceeding modern levels at 30-40 ka, but most arguments for sea level above -35 m during this age range are based on radiocarbon dates, (e.g. Curray, 1965; Milliman and Emery, 1968; Finkelstein and Kearny, 1988) which should be viewed with caution during this age range (Thom, 1973). Two exceptions to this rule are the Bahamas cave speleothem data of Mylroie and Carew (1988), and the shell carbonate data from southern Australia reported by Murray-Wallace et al. (1992). Mylroie and Carew (1988) report marine serpulid encrustations at approximately modern sea-level between stalagmite layers dated at 49 ka and 37 ka by the ^{234}U / ^{230}Th technique. Murray-Wallace et al. (1992) reported a -22 m sea-level stand from South Australia hat gave radiocarbon ages of 30-45 ka, bolstered by an amino acid age estimate of 46.4$\pm$9.1 ka. However, the amino acid age estimate would not preclude correlation of this highstand to the earlier stage 3 peak ca. 55 to 60 ka, which would be more consistent with the Barbados and New Guinea data. Still, even using relatively conservative estimates for these highstand elevations, such as -28 m and -40 m, the SPECMAP curve severely underestimates sea level during stage 3.

Two approaches have been used to reconcile the disparity between sea-level estimates derived from stratigraphic and isotopic methods. The first is to adjust benthic oxygen isotope records for cooling of deep ocean water during glaciation (e.g. Chappell and Shackleton, 1986; Shackleton, 1987; Guilderson (pers. comm, 1996). Guilderson et al. (pers. comm., 1996) have constructed a sea-level curve from a high-resolution benthic foraminiferal $\delta^{18}O$ record from the Caribbean, assuming that deep water temperatures cooled in a stepwise fashion from 115 to 15 ka (after Chappell and Shackleton, 1986), and a correction of 10 m per 0.11‰ change in $\delta^{18}O$. The resulting curve is a close match for the New Guinea sea-level record, although the later stage 3 highstand is weakly developed (Fig. 26b).

The second approach is to use $\delta^{18}O$ from planktonics in a region that is believed to have maintained essentially constant surface temperatures throughout the late Quaternary (e.g. Linsley, 1996). One such area is the Sulu Sea, located within the western equatorial Pacific "warm pool", where foraminiferal transfer functions imply temperatures within 2°C of modern conditions throughout the late Quaternary (Thunnell et al., 1994). The Sulu Sea record produces a very close agreement with the New Guinea highstands (Linsley, 1996), but suggests excessively deep lowstands in stages 5 and 4 (-80 m during stage 5, in excess of -100 m during stage 4) (Fig. 26a). One possible explanation for this is that the circulation between the surrounding oceans and the Sulu Sea is significantly restricted during lowstands of sea level (Linsley, 1996), which may increase the influence of salinity changes on the $\delta^{18}O$ ratios. However, Linsley (1996) argues that the closeness of the difference in $\delta^{18}O$ between stage 2 and stage 1 to the presumed global ice-volume signal suggests this is not a problem. The timing of the stage 3 highstands also appears to be slightly offset from the reef peaks, with the early stage 3 highstand appearing to peak at about 49 ka on the Linsley (1996) curve (Fig. 26a). Linsley assumes a constant sedimentation rate between 59 ka and 23 ka to arrive at his curve. Perhaps fluctuations in sedimentation rates result in the apparent offsets.

A comparison of the Linsley (1996) and Guilderson (pers. comm., 1996) curves shows generally similar shapes, but during stage 3, the peaks are consistently ~ 5 ka later on the Linsley (1996) curve (Fig. 27). The amplitude of the two curves also differs, with the Linsley (1996) curves showing much deeper lowstands during stages 5-4 and higher highstands in stage 3.

New Jersey sea-level estimates using the short and medium chronologies are shown in Figure 27, along with the results of Guilderson (pers. comm., 1996), and Linsley (1996) the SPECMAP curve, and the Bloom et al. (1974) New Guinea reef data. The short chronology generally fits the Linsley (1996) curve from stage 4 on, but tends to be approximately 10 m too high, except early in stage 3. Compared to the Guilderson (pers. comm., 1996) curve, the results are similar, except that the later stage 3 highstand is much too high. This interpretation suggests that no record of the stage 5 fluctuations shown is preserved.

The medium chronology shows very close agreement with the Guilderson (pers. comm., 1996) throughout the late Quaternary, and the ~-75 m stillstand suggested from 30-40 ka could account for the highstand wedge seen on the outer shelf on line 2-14. Compared to the Linsley (1996) record, it underestimates the depth of the stage 4 lowstand, but otherwise is in close agreement, except that no record of a later stage 3 highstand is preserved.

The medium chronology is consistent with the amino acid racemization data, and does not conflict significantly with data from Barbados, New Guinea or with the Guilderson et al. (in prep.) and Linsley (1996) curves. The absence of late stage 3 highstand deposits could be a result of erosion during the stage 2 lowstand. In view of

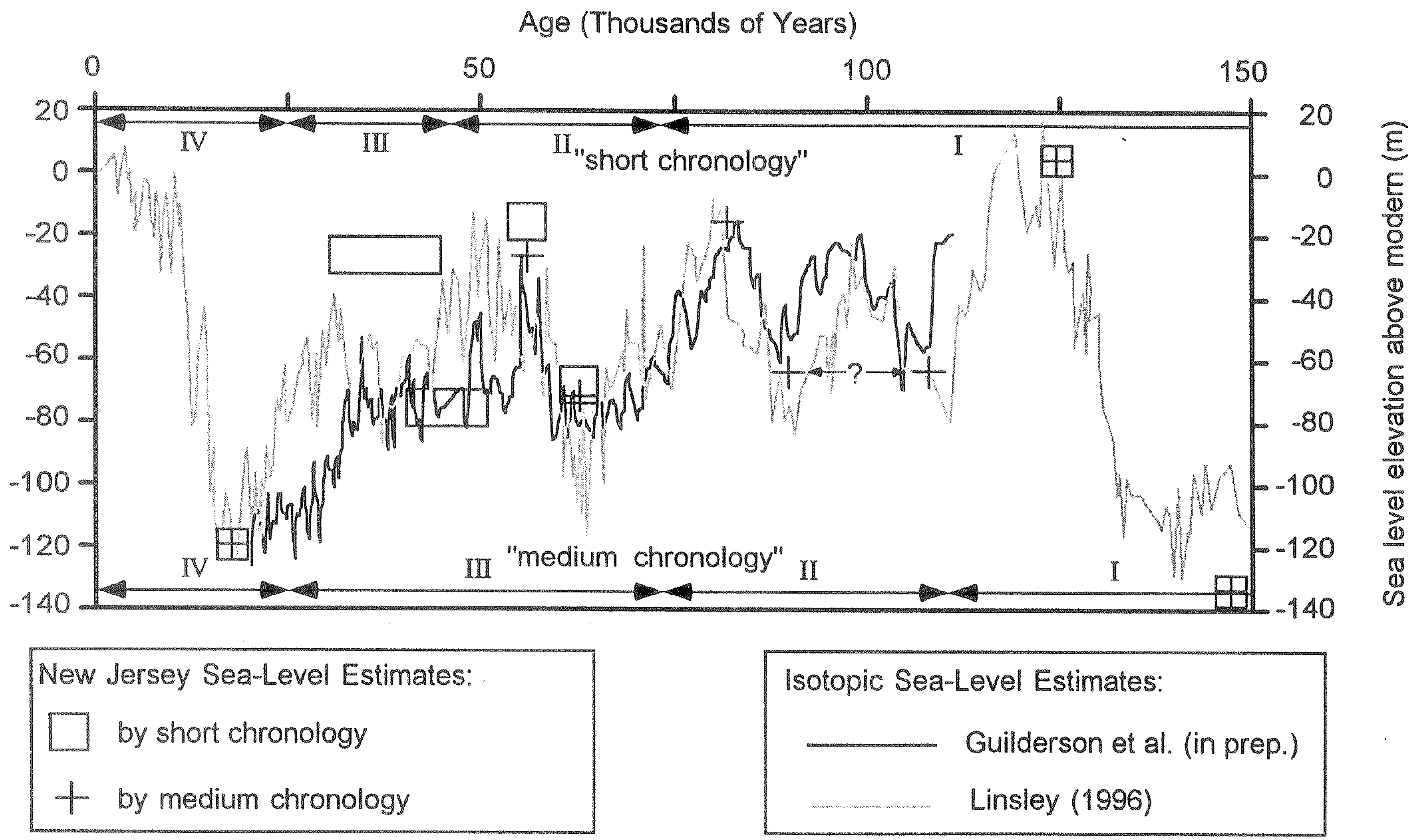


Figure 27. New Jersey Sea-Level Estimates. Shows approximate positions of sea-level, adjusted for subsidence, from the New Jersey shelf sequences using the short and medium chronologies, along with the Guilderson (in prep.) and Linsley (1996) sea-level estimates. Also shown are the approximate temporal ranges of sequences I through IV using both chronologies.

the paucity of isotopic data supporting high sea levels late in stage 3, it seems likely that the sea levels recorded in the reef terraces were short-lived. They may have left only thin deposits on the New Jersey shelf that were reworked beyond recognition. The short chronology can also generally fit the sea-level curves, though the stage 3 highstands appear about 10 m too high. This could reflect inaccuracy resulting from glacial-isostasy and tectonics, however. The amino acid racemization data tend to support the medium chronology, but given the uncertainties related to the racemization kinetics of *Astarte*, insufficient evidence exists to reject the short chronology.

Recently, Chappell (1996) dated corals from near the crest of the Huon Peninsula reef terraces and obtained significantly older dates for reef terraces II, III and IV. After correcting for tectonic uplift, Chappell (1996) presented a new sea-level record that yields similar results to oxygen-isotopic estimates. Based on these new data, the highest sea-level stand during stage 3 is approximately -40 m, and the stage 3.1 highstand is only about -60 m. If the revised ages of the terraces prove correct, the higher highstands of the short chronology are probably untenable; in fact, even the medium chronology indicates sea-level at approximately 10 m higher than that of Chappell (1996) for stage 3.3. However, Quaternary researchers have been correlating their sea-level records to the Huon Peninsula reef terraces for twenty years, and generally obtained consistent results (Murray-Wallace et al., 1992). In view of the wide acceptance of the prior curve, caution should be used in applying the results of the Chappell (1996) study, until corroborating results have been obtained.

6.6.5 Relationship to Glacial History of Eastern North America.

If the medium chronology is correct, substantial (60 m) falls in sea level occurred during stage 5, implying the presence of significant ice volume in North America (Boulton et al., 1985). Evidence of such an ice buildup at this time exists in southern Quebec, where the presence of varved lacustrine clay deposits implies that the St. Lawrence River (Fig. 2) was dammed by ice during at least part of stage 5 (Lamothe, 1989; Clet and Occhietti, 1994).

There is evidence of substantially increased sediment input during sequence III, compared to sequence II (Chapter 5), especially if the short chronology is correct. A possible explanation for the enhanced sediment supply is glacial activity in the Hudson drainage basin occurring in stage 4 / early stage 3 by the medium chronology, or later in stage 3, according to the short chronology. The maximum extent of glaciation during stage 4 in eastern North America is not firmly established. Sea-level during isotope stage 4 (the early Wisconsin) was apparently no lower than -65 to -80 m, substantially higher than during the Late Wisconsin maximum (stage 2). This is consistent with the findings of Ridge et al. (1990) that "Altonian" drift beyond the late Wisconsin maximum (Fig. 2) is pre-Wisconsin, rather than early Wisconsin. However, there is evidence of a significant glacial advances during the early to middle Wisconsin in Canada. The Chaudière Till in southern Quebec was formed by ice moving out of the Appalachians that was confluent with the Laurentide in the St. Lawrence Lowlands (Fig. 5), and may be of stage 4 age (Karrow and Occhietti, 1990). The Sunnybrook Drift, a diamicton unit in the Toronto area (Fig. 5), has been variously interpreted as a till from an early Wisconsin glacial advance (Karrow, 1967;

Hicock and Dreimanis, 1989) or a glacial-lacustrine unit (Eyles and Eyles, 1983). Either interpretation implies that the St. Lawrence Valley was blocked by ice, and there was ice in or near the Toronto area at this time.

The extent of ice farther south is unclear because of few exposures of pre-Late Wisconsin sediments and little material available for dating. Koteff and Pessl (1985) suggested that the widespread "Lower Till" in New England was deposited during an early Wisconsin glacial advance, but Oldale and Colman (1992) regard it as an older unit, at least in southern New England. Evidence for early and middle Wisconsin events in New York state is largely restricted to western New York (Muller and Calkin, 1993). Fullerton (1986) regards the "brown till" at Gowanda, New York (Fig. 5) as early Wisconsin, and Dreimanis (1992) tentatively correlates it to the Sunnybrook drift in the Toronto area. Lacustrine sediments in the Finger Lakes region (Fig. 5) suggest that ice dammed their drainage basins during the early or middle Wisconsin, but no till units from this time have been clearly identified (Muller and Calkin, 1993). In summary, the limit of the Laurentide appears to have been within the southern Great Lakes Lowlands during stage 4 and parts of stage 3. The Hudson must have drained much of this ice sheet, but the drainage was likely through a proglacial lakes, where much of the sediment would likely have settled out. Thus, the Laurentide may not be a viable sediment source for the region during stages 4 and 3. However, the presence of ice in the region, and presence of ice in the Appalachians of southern Quebec and Maine (Karrow and Occhietti, 1990) suggests that alpine ice could have been present in areas such as the Green Mountains, the Adirondacks, and perhaps even the Catskills (Fig. 5). Subsequent glaciation has destroyed any record of these small ice sheets, except perhaps in northern New England, but they could have played an important role in sediment delivery during stages 4 and 3.

Alternatively, sediment supply could increase through denudation of pre-existing Quaternary sediment in the drainage basin (Chapter 5). Kutzbach (1987) predicted increased storminess near the southeastern margin of the Laurentide ice sheet resulting from the thermal contrast between the glacier and the ocean. Thus, increased precipitation, particularly combined with reduced vegetation cover resulting from a colder climate, could result in increased erosion.

6.7 Conclusions

Four late Quaternary depositional sequences can be recognized from the shelf stratigraphy of New Jersey. The age of these units remains uncertain, with two age models that are reasonable in view of other sea-level records and with isotopic curves (Fig. 28). The aminostratigraphic data tend to prefer the "medium" chronology of Carey et al. (1995b, 1996) over the "short" chronology of Sheridan et al. (1993), but these data are insufficiently calibrated to other absolute age data at this time.

During stage 6, the lower boundary of the oldest sequence, sequence I, was cut. The Hudson flowed southward across the shelf to the vicinity of Spender Canyon for at least some of this time. The highstand associated with this unit includes parts of the Cape May Formation 5-10 m above modern sea level, which formed during stage 5e. Using the medium chronology, sea-level fluctuations during stage 5 are represented by sequence II. It is possible that both lowstands, 5b and 5d are preserved. Sea level

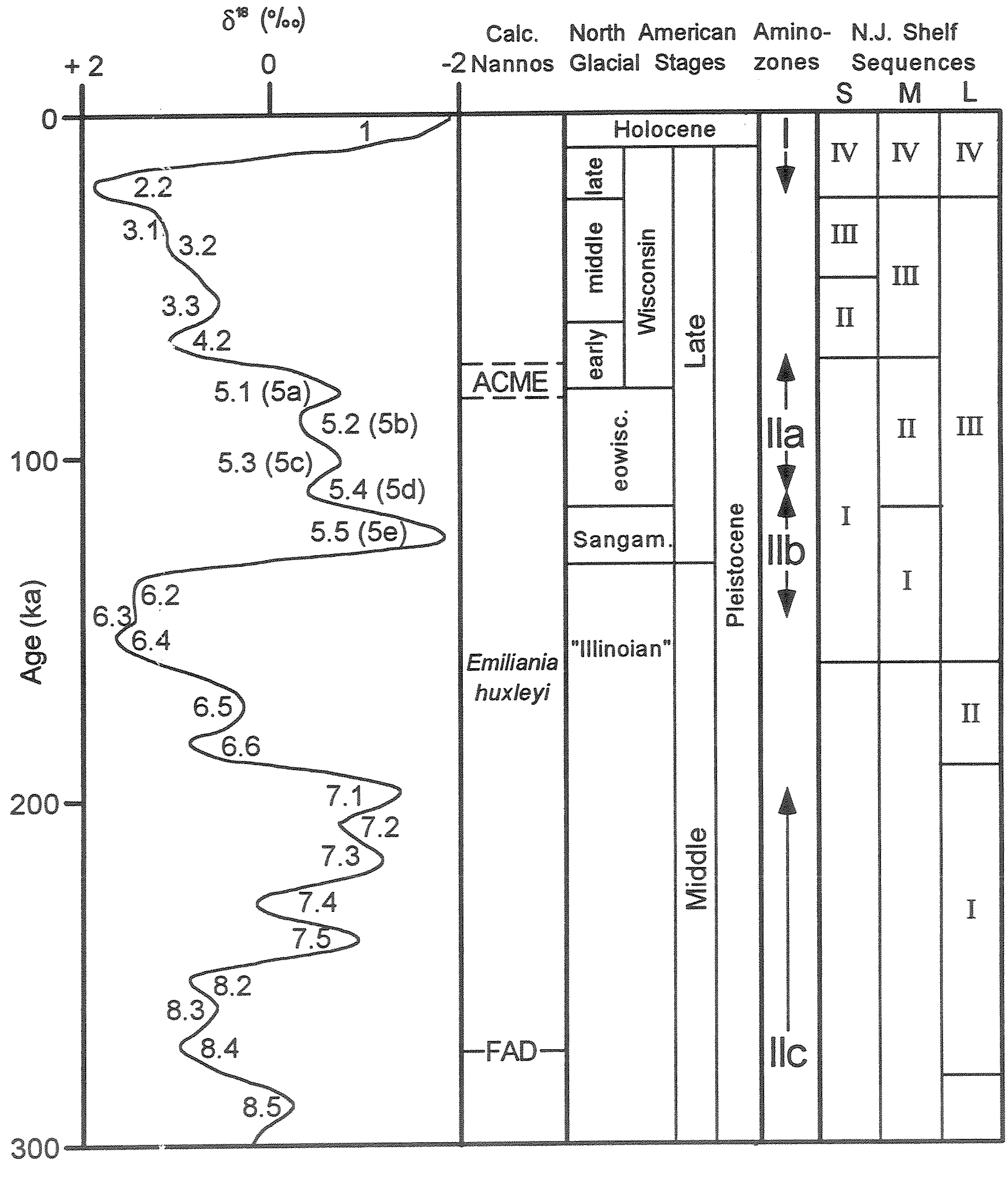


Figure 28. Age of the New Jersey shelf sequences. The age of the sequences using each of the chronologies (S - short, M - medium, and L - long) plotted with the other late Quaternary time scales from Figure 2.

during this time fell to at least -58 m, then rose to approximately -20 m. A prominent Paleohudson Delta formed on the mid-shelf southeast of Barnegat Inlet during this time. Elsewhere, this sequence is thin to absent.

During stage 4, sea level fell to between -64 and -80 m, cutting the lower boundary of sequence III. The prominent valley described by Knebel et al. (1979) was incised at this time, but largely filled during the subsequent highstand. Headward erosion of a valley landward of South Tom's River Canyon also occurred during stage 4. During the early part of stage 3, 55-60 ka, sea level rose above -28 m, filling numerous incised valleys. A subsequent stillstand in sea level at about -70 m, may be recorded as a prograding shoreline on the outer shelf. Sediment input to the shelf at this time appears to have increased substantially, which may be related to glacial activity in the Hudson drainage basin.

The youngest sequence, sequence IV, was deposited during the late Wisconsin and Holocene (stages 2-1). The "Outer Shelf Wedge", "Mid Shelf Wedge" and "Transgressive Shelf Blanket" all belong to this sequence. Sea level dropped to about -124 m during the maximum lowstand. Sediment supply was able to keep up with sea-level until sea-level rose to approximately -75 m; subsequent transgression truncated the Outer Shelf Wedge.

If the short chronology is correct, the interpretation of sequences I and IV does not change. However, no record of stage 5 sea-level fluctuations is preserved and the prominent delta in sequence II formed during stage 4. The -20 m shoreline near Barnegat Inlet would then have formed early in stage 3. This model suggests that the large shelf valley near South Tom's River Canyon formed during a stage 3 lowstand,

and that sediment input to the shelf was significantly greater during stage 3 than during stage 4.

CHAPTER SEVEN: CONCLUSIONS

1) Four regionally extensive unconformities were identified in the upper Quaternary section of the New Jersey shelf. These unconformities defined four sequences, numbered I through IV from oldest to youngest; sequence IV includes the modern transgressive systems tract. An unconformity within sequence II may exist, if so, it consists of two sequences.

2) Sequences I and IV formed during major glacial / interglacial sea-level fluctuations. Lowstand shorelines associated with these sequences are located near the shelf edge, while highstand shorelines were near the modern shoreline.

3) Sequences II and III formed during stadial / interstadial sea-level fluctuations. Lowstand shorelines associated with these sequences are located on the mid-shelf and record sea levels of ~-60 to -75 m. During the highstand intervals of these sequences, the shoreline was located on the inner shelf at elevations of ~-20 to -30 m.

4) The lower boundary of sequence I is no older than stage 8, and probably stage 6. The lower boundary of sequence IV was cut during the late Wisconsin (stage 2) lowstand.

5) The age of the stadial / interstadial sequences, II and III remains uncertain. Two alternative chronologies appear possible. One is that sequence II records sea-level fluctuations in stage 5 and sequence III the stage 4/3 fluctuation. The other possibility is that sequence II records the stages 4 and 3, while sequence III resulted from sea-level fluctuation during stage 3.

6) Sequences formed under conditions of high-amplitude, and high-frequency eustatic change, and relatively low rates of sediment supply and subsidence, tend to be thin and fragmentary due to intensive fluvial incision during lowstands, and marine erosion during transgressions and highstands. As predicted by Steckler et al. (1993), type I sequence boundaries predominate and highstand tracts are poorly preserved.
7) Although isotopic curves and coral reef terraces suggest that at least four stadial - interstadial fluctuations took place between 125 and 20 ka, the deposits of only two, or possibly three, of these are preserved on the New Jersey margin. Preservation of interstadial sequences does not necessarily improve seaward, but can be influenced by local factors such as fluvial valley positions and differential tectonism related to hinge lines, salt movement and glacial isostasy. On the New Jersey shelf, the interstadial sequences are best preserved within a broad swale in the sequence I boundary on the mid-shelf probably formed by a Hudson River flowing well south of the modern Hudson Shelf Valley.
8) The collapse of peripheral bulges formed during major glacial cycles resulted in a northward deepening by 20 to 40 m of reflection horizons within sequences I and IV. The late Wisconsin peripheral bulge may have caused the Hudson River to switch from the buried valley discovered by Knebel et al. (1979) to the modern shelf valley position.

9) The New Jersey shelf stratigraphy records substantial variations in sediment supply. The volume of the mid-shelf and outer shelf wedges in sequence IV implies that sediment input to the region was at least an order of magnitude greater during the last deglaciation than under modern conditions. Although sequences II and III record sea-level fluctuations of similar magnitude, sequence III is generally thicker, more continuous, and contains preserved prograding highstand deposits. This suggests that during deposition of sequence III, sediment supply was substantially greater.

10) Fluctuations in sediment supply recorded in the New Jersey shelf likely reflect changes in the Hudson drainage basin. Sediment may have been carried directly by glacial meltwater, or been remobilized from pre-existing glacial deposits because of changes in climate and vegetation associated with glaciation. Unfortunately, the specific relationships cannot be recognized because of the imprecise dating of both the shelf sequence and the terrestrial glacial events.

11) In the absence of direct evidence of glaciation, the sediment supply, glacial-isostatic and sea level changes recorded in the shelf stratigraphy could be mistaken for tectonic effects. The best evidence of glaciation in the area would be the high frequency of the sea-level events, the concentration of sediment supply in lowstands and early transgressions, and the correlation of the sea-level changes to dramatic climatic changes.

12) Despite the glacial-isostatic adjustments and the crude estimate of subsidence used in the study, sea-level estimates from the New Jersey margin fell within 10 to 15 m of "eustatic" estimates from coral terraces and isotopic records, using either of two chronologies. This suggests that the corrections required to account for glacio-isostasy in southern New Jersey are relatively small, and that sequence stratigraphic analysis of passive margins can yield useful estimates of global sea level change.

13) Current models of glacial isostasy use only sea-level data from modern coastal areas. Incorporation of the sea-level histories recorded on the world's continental shelves may help to improve these models.

14) Continental shelf sections are generally more complete, and more easily dated, than terrestrial Quaternary sections. Therefore, increased study of ice-marginal continental margins may yield a better understanding of the terrestrial glacial record.

APPENDIX

1. Data Summaries for AMCOR Drill Sites

Data for all New Jersey shelf sites are summarized in the form of two tables. The data for sites 6009B, 6010, 6011 and 6020 may be found on pages 146, 147, 148, and 149, respectively.

The first table lists physical properties of the core samples. The depth of origin below the sea bottom, the length recovered, and description of the sample in the core catcher are from Hathaway et al. (1976). The grain size data and moisture contents reported for samples taken are from Poppe et al.(1976).

The second table summarizes the biological and geophysical characteristics of the samples. The calcareous nannofossil data are from Valentine in Harris (1983), the benthic foraminifera from Poag (1979), the pollen climate index and *Quercus* species are from Groot et al. (1995), and the ostracod data from Cronin (1979). The D-alloisoleucine / L-isoleucine ratios are from data reported in this study and in Groot et al. (1995). The seismic stratigraphy indicates how these cores relate to the stratigraphy developed in this study.

AMCOR 6009B

Location 38°51.27'N, 73°35.47'W, Water Depth 58.5m
Near AII 2-20, 5/11/75, 0225 (uniboom line)
Also near AII, Line 2-3, 5/4/75, 1420 (3.5 KHz line)

#	Depth (m)	Length (m)	Description (core catcher sample)	Grain Size Analysis Loc.; Mean, Sorting; Gravel/Sand/Silt Clay	Moisture Content
1	0.0 - 7.6	1.1	dark gray, sandy clay	1-1, 81cm: 6.94, 1.69 0.0 / 3.0 / 69.7 / 27.3	1-1, 149cm: 25%
2	7.6 - 15.5	2.6	dark gray, coarse sand	2-1, 125cm:2.46, 0.81 0.0 / 98.4 / 0.9 / 0.7	
3	15.5 - 24.7	1.7	dark gray smooth plastic clay		cc: 38%
4	24.7 - 34.1	2.5	dark gray clay	4-1: 115cm:8.25,1.69 0.5 / 1.7 / 35.6 / 62.4	4-2, 139cm: 29%
5	34.1 - 43.6	1.6	dark gray clay, slightly silty		5-1, 149cm: 32% cc: 29%
6	43.6 - 53.0	0.09	olive-gray silty clay with shells		

#	Nannofossil Zonation	Foraminifera	Ostracods	Pollen Index/ *Quercus* sp.	D / L ratio	Seismic Stratigraphy
1		*Elphidium excavatum*		-0.4	0.02 0.05	Seq. III
2				no samples	*Astarte:* 0.20 Unknown: 0.32	Seq. III
3		*E. excavatum* *N pachyderma (dextral)*		-0.7 / 5 and 6, not 1		Seq. III/II
4		*E. excavatum* *N.pachyderma (dextral)*		0.4 / 0.6		Seq. II/I
5						Base of Seq. I
6					Unknown: 0.66, 0.67 0.62	

AMCOR 6010

Location: 39°03.70'N; 73°05.90'W
Water Depth: 75.9m
Near AII 2-16, 05/09/16, ~1700

#	Depth (m)	Length (m)	Description (core catcher sample)	Grain Size Analysis Loc.; Mean, Sorting; Gravel/Sand/Silt Clay	Moisture Content
1	0.0 - 8.2	3.4	clay, v. dk. gray, sl. silty, plastic		cc: 40%
2	8.2 - 16.8	2.4	clay, v. dk. gray, plastic	2-2, 119cm: 8.68,1.59 0.0 / 1.6 / 24.9 / 73.6	2-2, 149cm: 38% cc: 37%
3	6.8 - 26.2	2.1	clay, v. dk. gray, sl. sandy, plastic, fossil.	3-2, 105cm: 8.19,1.84 0.3 / 3.5 / 36.0 / 59.9	3-2, 149cm: 41% cc: 19%
4	26.2 - 35.7	1.8	clay, v. dk. gray, v. sl. sandy, v. plastic	4-2, 105 cm:8.24,1.64 0.0 / 0.7 / 40.1 / 59.5	4-2, 149cm: 33%
5	35.7 - 45.1	1.2	clay, v.dk. gray, v. sl. sandy, plastic	5-1, 85cm: 6.0 / 7.4 / 43.7 / 42.6	cc: 29%
6	45.1 - 54.9	3.7	clay, v. dk. gray, v. sl. sandy, plastic, sl. fossil.		

#	Nannofossil Zonation	Foraminifera	Ostracods	Pollen Index/ *Quercus* sp.	D / L ratio	Seismic Stratigraphy
1			*Leptocythere, Cytheridea*			Seq. IV
2						Seq. IV
3	*Emiliania huxleyi*		*Leptocythere, Cytheridea*			Seq. III
4						Seq. I
5	*Pseudoemiliania lacunosa*		*Munseyella*		unknown: 0.29 *Cyclocardia* 0.27, 0.35	Base of Seq. I
6						

AMCOR 6011

Location: 39°43.5'N, 73°58.6'W
Water Depth: 22.3m
Near Whitefoot Line 6/15/81, 0130
Also near Whitefoot Line 6/16/81, 1940

#	Depth (m)	Length (m)	Description (core catcher sample)	Grain Size Analysis Loc.; Mean, Sorting; Gravel/Sand/Silt Clay	Moisture Content
1	0.0 - 9.1	1.4	olive gray silty clay	1-1, 100cm: 7.21,1.94 0.1 / 3.2 / 59.0 / 37.5	1-1, 149cm: 49%
2	9.1 - 18.3	0.5	med.- coarse sand, w/ hvy. minerals		
3	18.3 - 27.4	0.6	med. to coarse sand		
4	27.4 - 32.6	0.3	fine to med. quartzose, sand w/ hvy. minerals		

#	Nannofossil Zonation	Foraminifera	Ostracods	Pollen Index/ *Quercus* sp.	D / L ratio	Seismic Stratigraphy
1						Seq. IV
2						Seq. IV/II/I
3						Base of Seq. I
4						

AMCOR 6020

Location: 39°25.41'N, 73°35.63'W
Water Depth: 39m
Near AII Line 2-16, 05/09/75, 2220

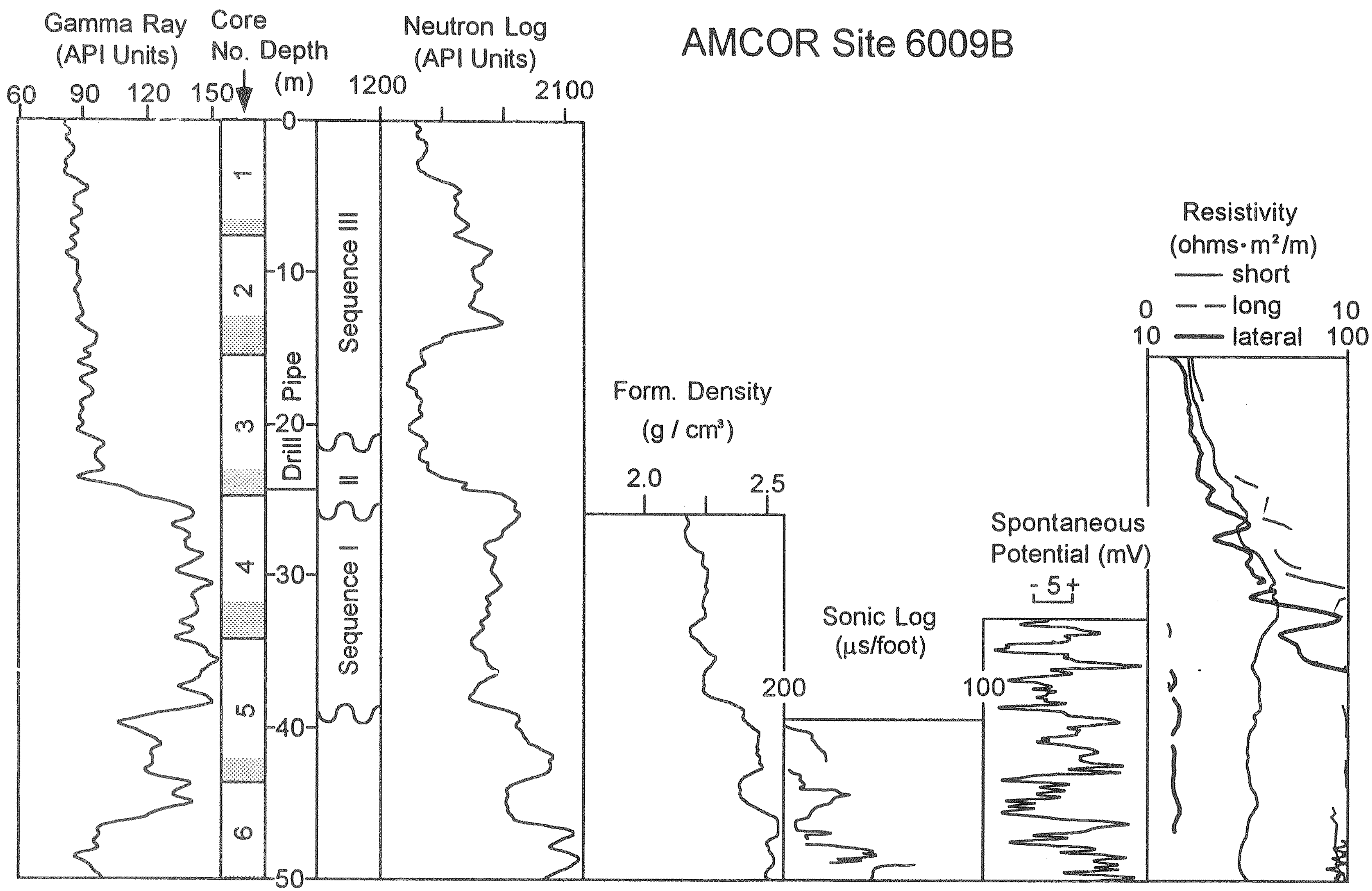
#	Depth (m)	Length (m)	Description (core catcher sample)	Grain Size Analysis Loc.; Mean, Sorting; Gravel/Sand/Silt Clay	Moisture Content
1	0.0 - 7.9	0.9	dk. gray, silty clay	1-1, 85cm: 7.70, 2.30 0 / 11.5 / 32.9 / 55.9	cc: 46%
2	7.9 - 16.5	4.1	dk. gray, silty clay	2-2, 25cm: 7.62, 2.45 0 / 12.7 / 31.5 / 55.8	2-2, 149cm: 30% 2-3, 0cm: 30% cc: 33%
3	16.5- 25.3	1.7	slightly silty, sticky gray clay	3-2, 118cm: 8.11,1.72 0.0 / 1.4 / 42.1 / 56.5	cc: 34%
4	25.3- 31.4	1.8	dk. gray, sandy,sticky clay	4-2, 15cm: 6.27, 3.28 0 / 28.8 / 31.2 / 40.1 4-2, 40cm: 5.33, 4.32 4.0 / 41.4 / 8.8 / 45.9	cc: 31%
5	31.4- 34.1	0.3	gray, sandy clay, shells		cc: 28%
6	34.1- 43.9	1.4	pebble gravel (rounded) nd gray clay	6-1, 72cm: 7.43, 2.31 0.8 / 9.2 / 42.5 / 47.8	6-1, 129cm: 19%

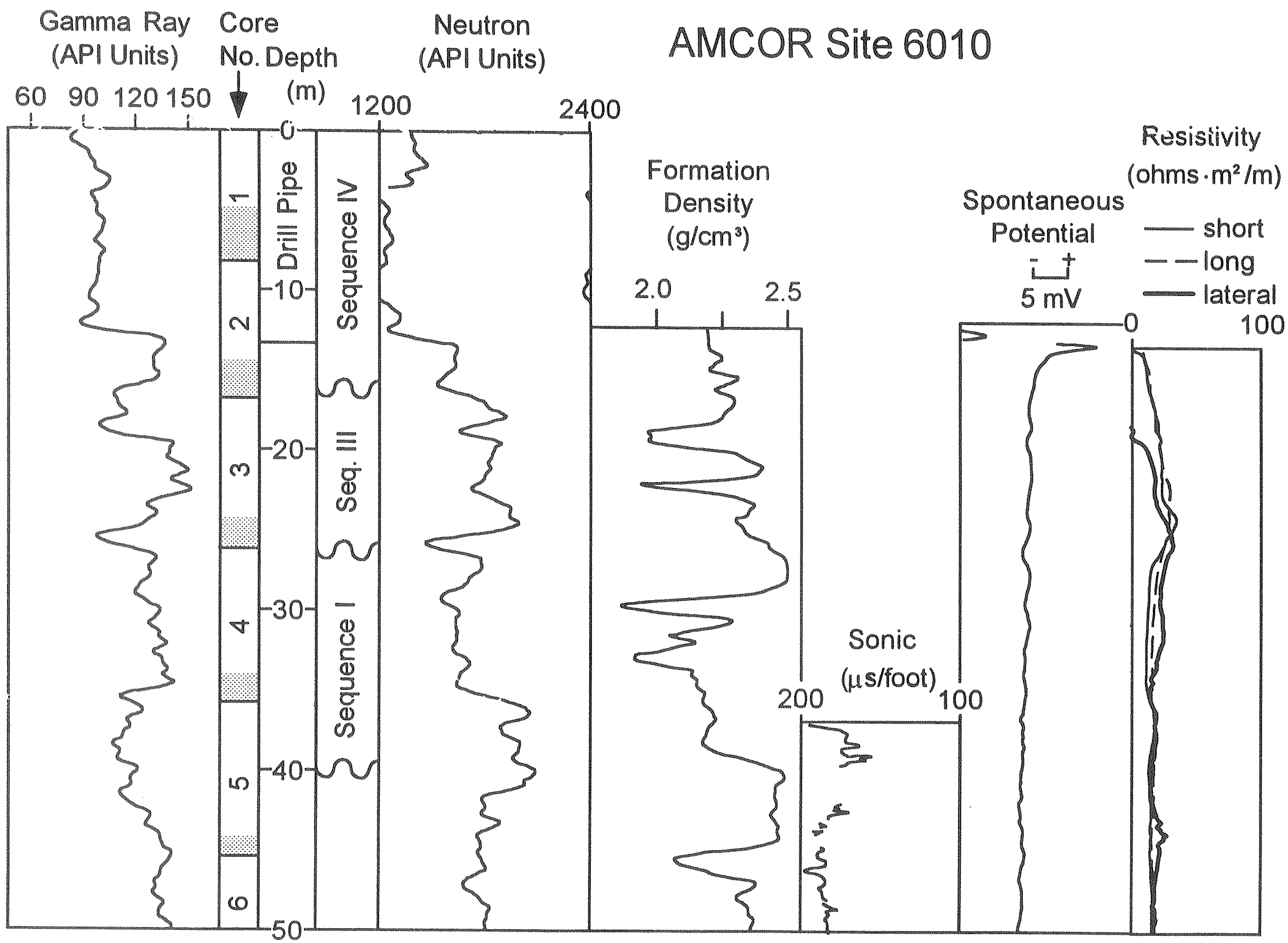
#	Nannofossil Zonation	Foraminifera	Ostracods	Pollen Index/ *Quercus* sp.	D / L ratio	Seismic Stratigraphy
1						Seq. IV
2					Unknown 0.27	Seq. IV / III / II
3					*Astarte* 0.18	Seq. II
4						Seq. II / I
5					unknown. 0.21	Seq. I
6						Base of Seq. I

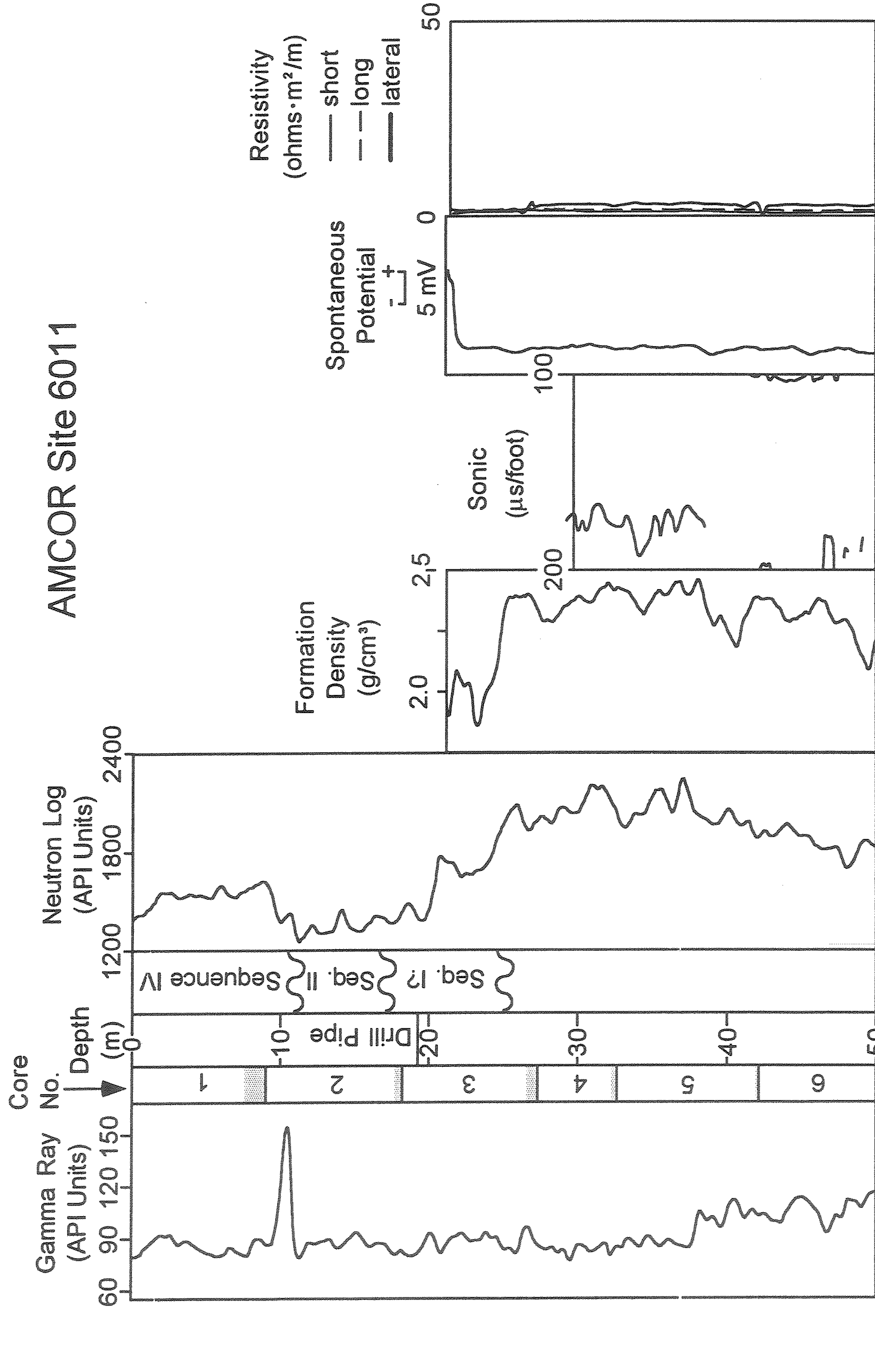
2. Geophysical Well Logs from AMCOR Drill Sites

Geophysical well logs from AMCOR sites 6009B, 6010 and 6011 are shown on pages 151, 152, and 153, respectively. No well logs were run at site 6020. Each page shows the gamma ray, neutron, formation density, sonic, spontaneous potential and resistivity logs. Also shown are the approximate elevations of the sequence boundaries, the limit of the drill pipe, and the positions of AMCOR cores. The shaded parts of the core number column indicate the amount of material recovered in each core.

AMCOR Site 6009B

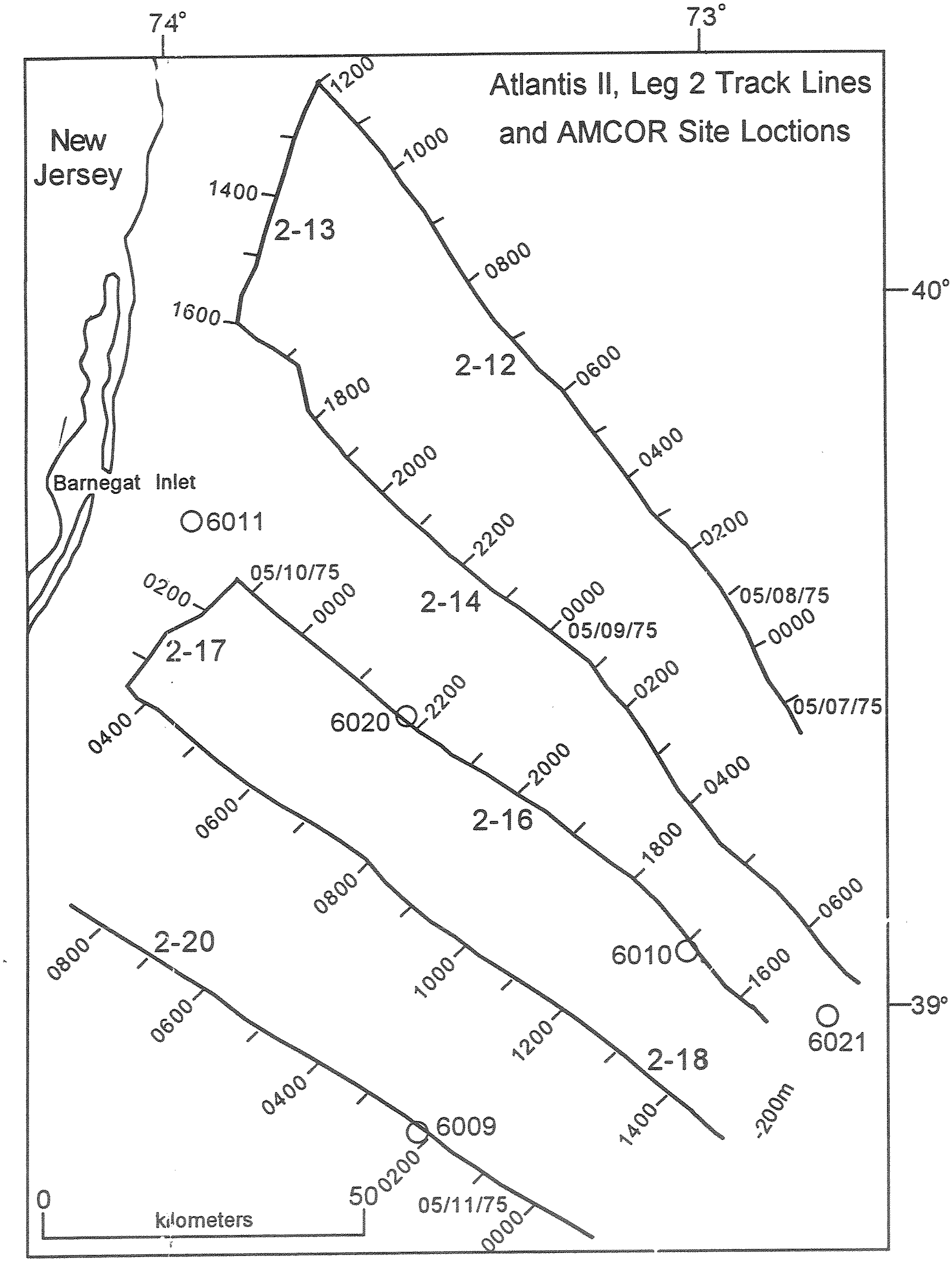


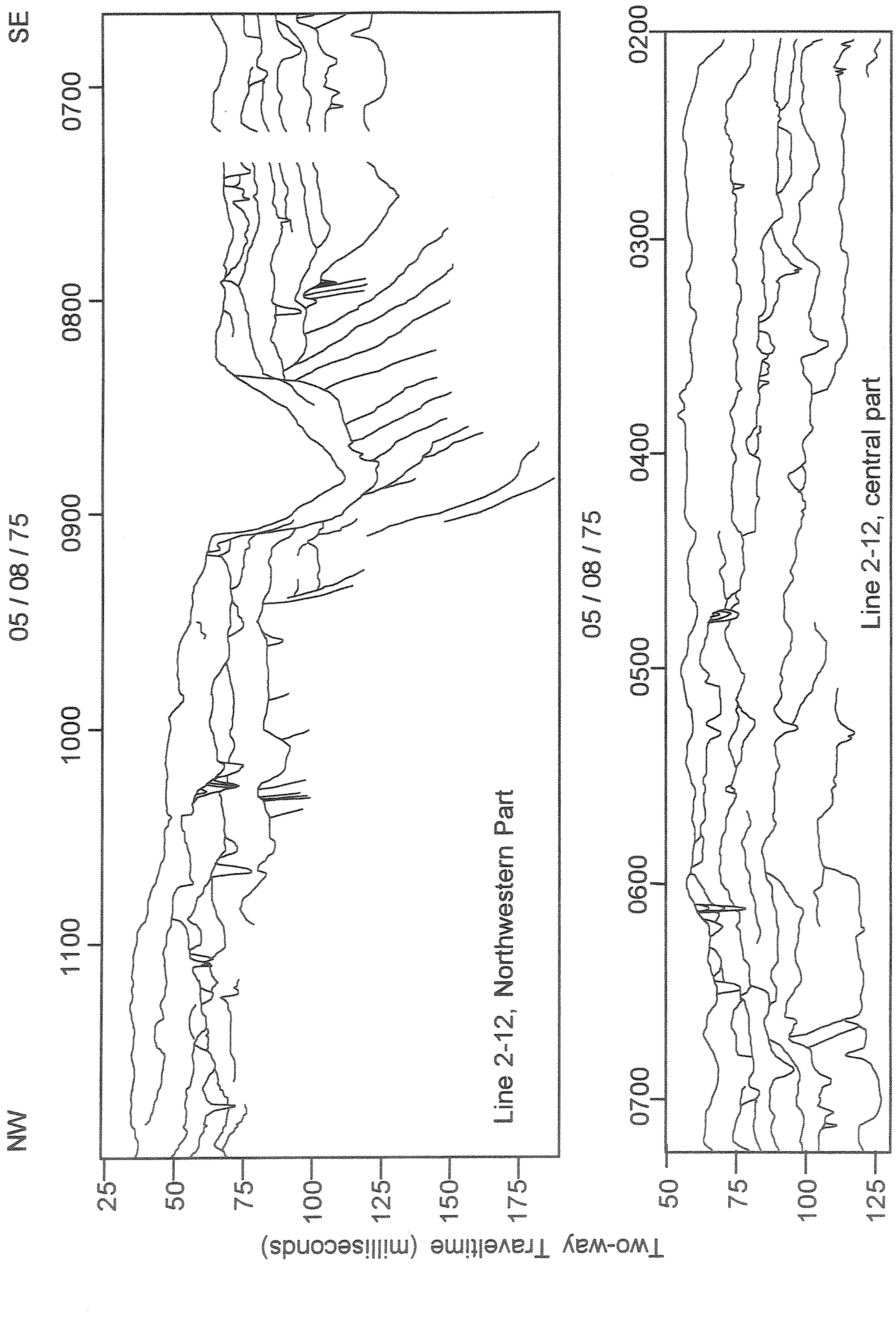

AMCOR Site 6010
Gamma Ray
(API Units)
60
90
120
150
Core
No.
Depth
(m)
0
10
20
30
40
50
1
2
3
4
5
6
Drill Pipe
Sequence IV
Seq. III
Sequence I
Neutron
(API Units)
1200
2400
Formation
Density
(g/cm³)
2.0
2.5
Sonic
(µs/foot)
200
100
Spontaneous
Potential
- +
5 mV
Resistivity
(ohms·m²/m)
short
long
lateral
0
100

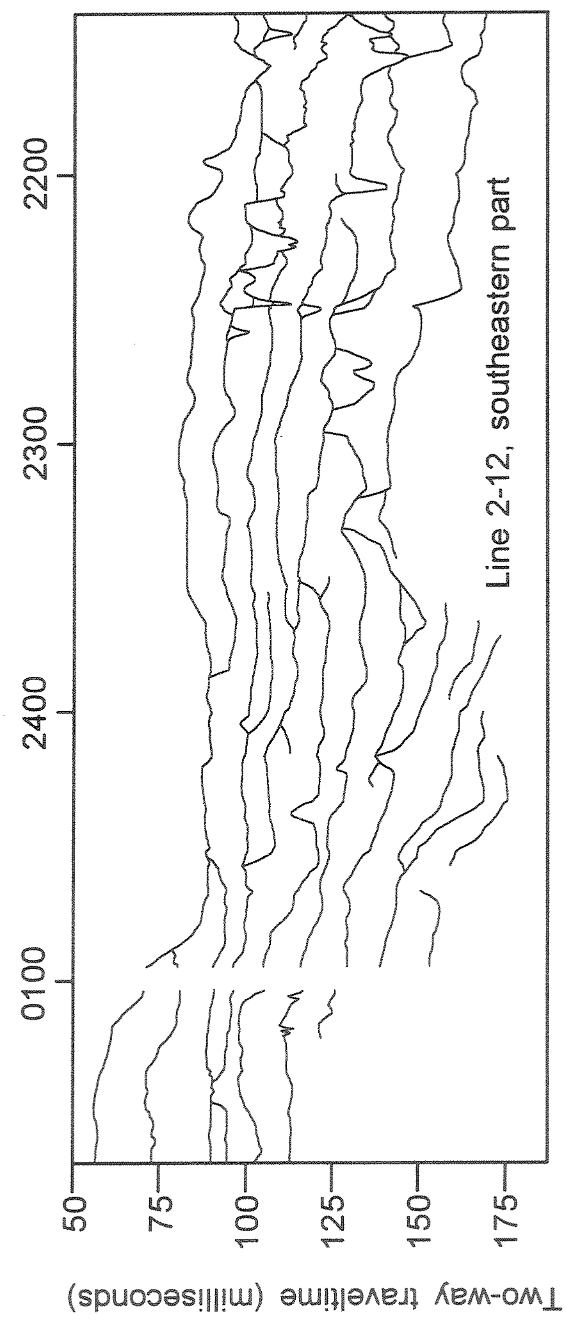

AMCOR Site 6011
Gamma Ray (API Units)
60 90 120 150
Core No.
1 2 3 4 5 6
Depth (m)
0 10 20 30 40 50
Drill Pipe
Sequence IV
Seq. II
Seq. I?
Neutron Log (API Units)
1200 1800 2400
Formation Density (g/cm³)
2.0 2.5
Sonic (µs/foot)
200 100
Spontaneous Potential
5 mV
Resistivity (ohms·m²/m)
short
long
lateral
0 50

3. Atlantis II, Leg 2 Seismic Data

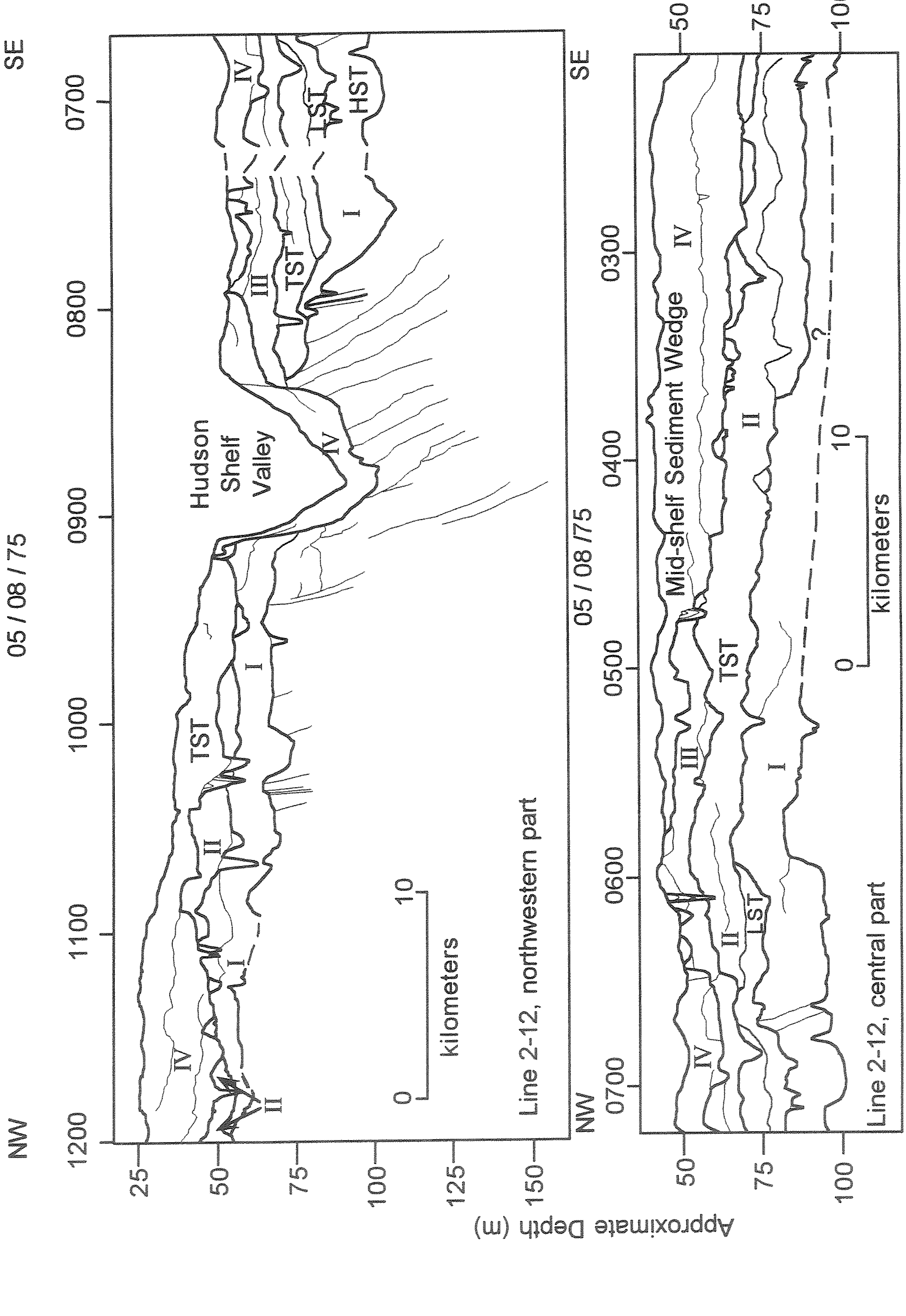
The following sections summarize Atlantis II, Leg 2, uniboom lines 12-14, 16-18 and 20. Complete tracklines with locations and times of data acquisition are shown on page 155. Following this are line drawings and examples of data from each line. Data shown were selected on the basis of coverage (they generally lie ~ 2 hours apart on the records), and to show salient features, and the relative quality of data at various times of collection. Line 2-12 is covered on pages 156-163, 2-13 on pages 164-166, 2-14 on pages 167-173, 2-16 on pages 174-178, 2-17 on pages 179-180, 2-18 on pages 181-187, and 2-20 on pages 188-194.

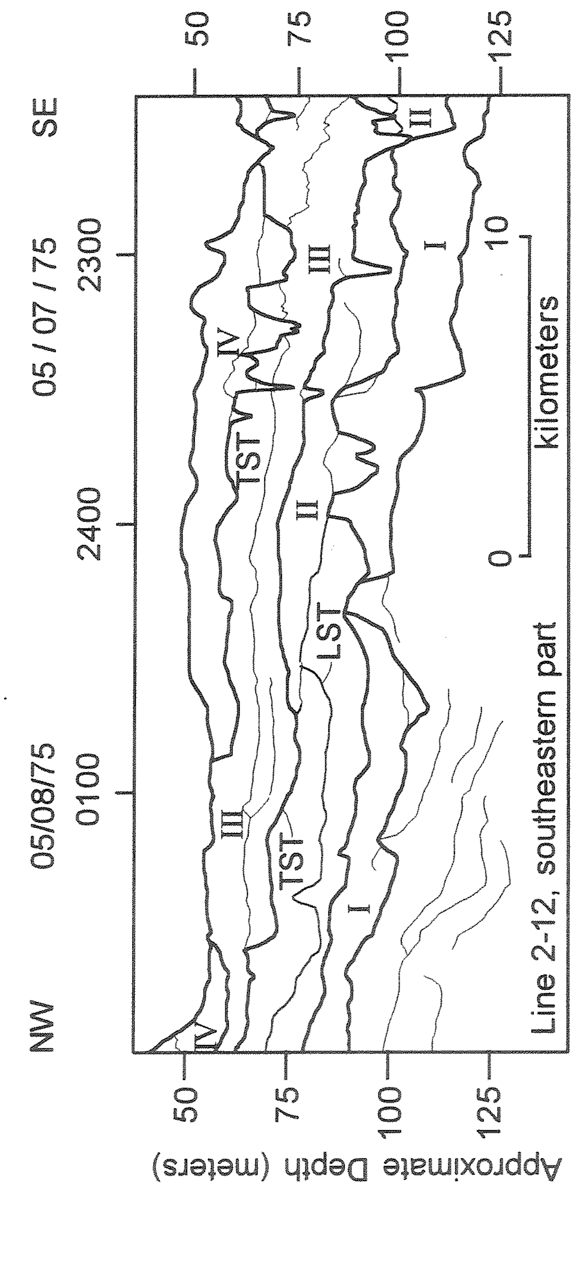

74°
73°
40°
39°
Atlantis II, Leg 2 Track Lines
and AMCOR Site Loctions
New
Jersey
Barnegat Inlet
2-12
2-13
2-14
2-16
2-17
2-18
2-20
6011
6020
6010
6021
6009
05/07/75
05/08/75
05/09/75
05/10/75
05/11/75
1200
1000
0800
0600
0400
0200
0000
1400
1600
1800
2000
2200
-200m
0
50
kilometers

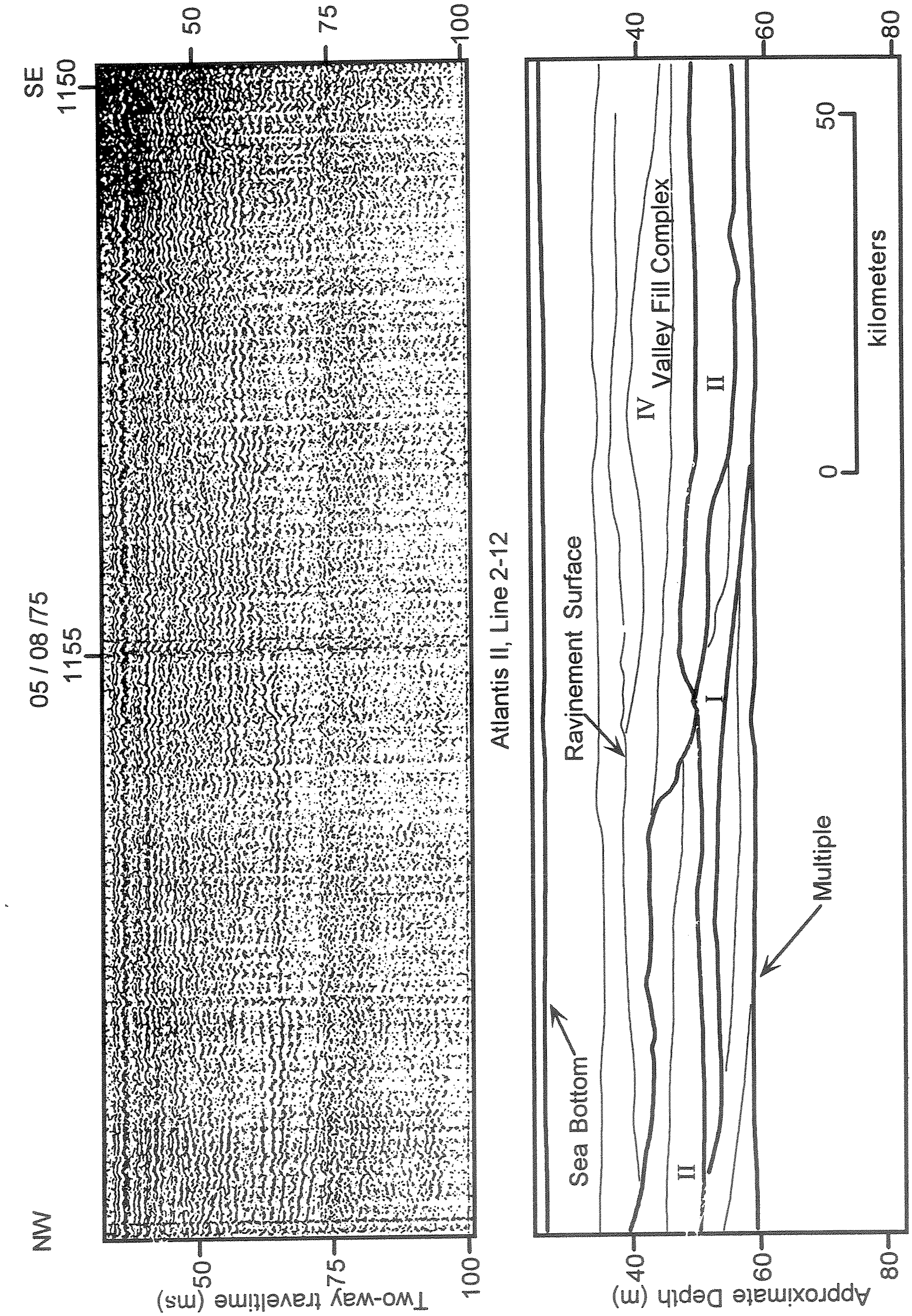

NW
SE
05 / 08 / 75
1100
1000
0900
0800
0700
Line 2-12, Northwestern Part
25
50
75
100
125
150
175
Two-way Traveltime (milliseconds)
05 / 08 / 75
0700
0600
0500
0400
0300
0200
Line 2-12, central part
50
75
100
125

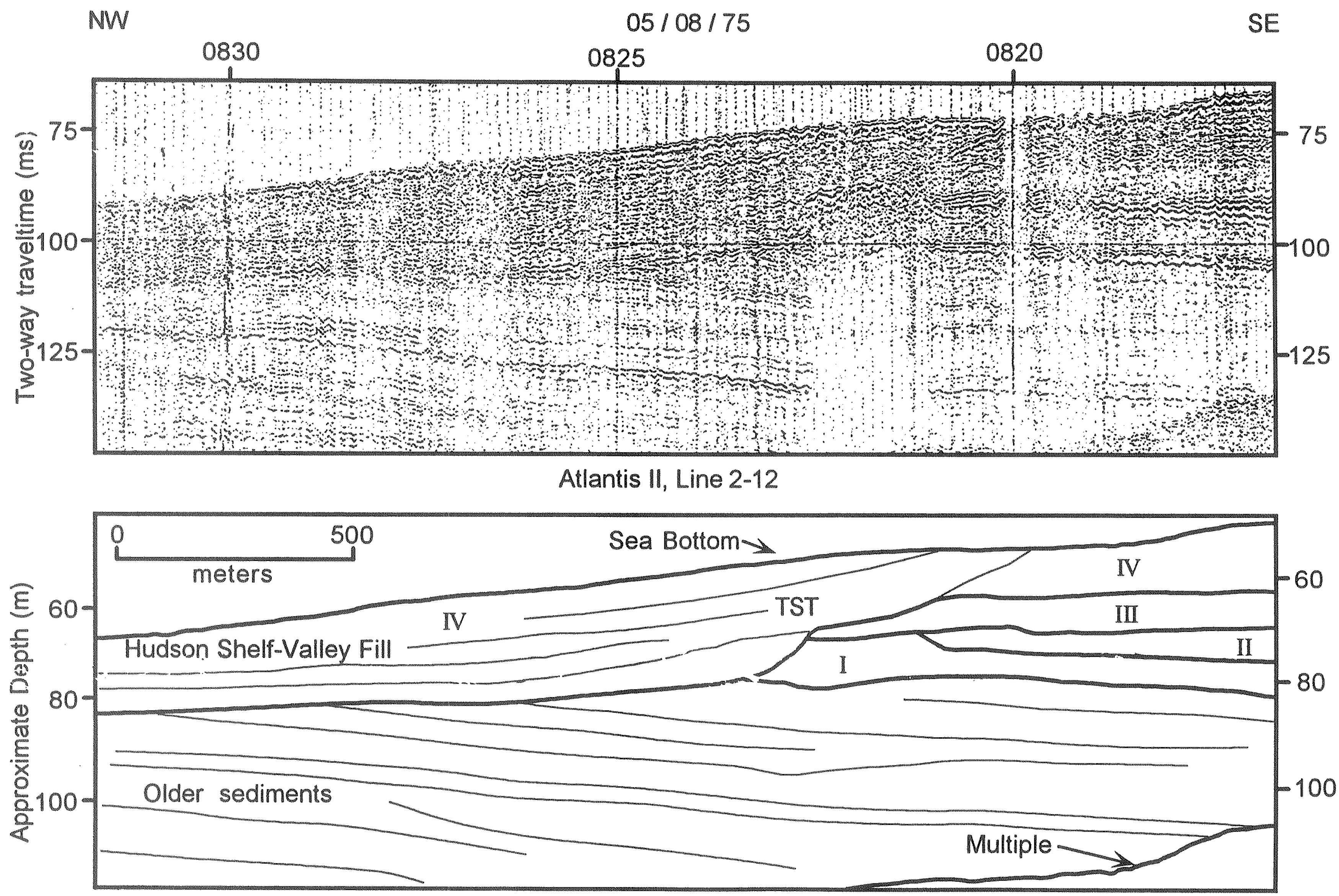


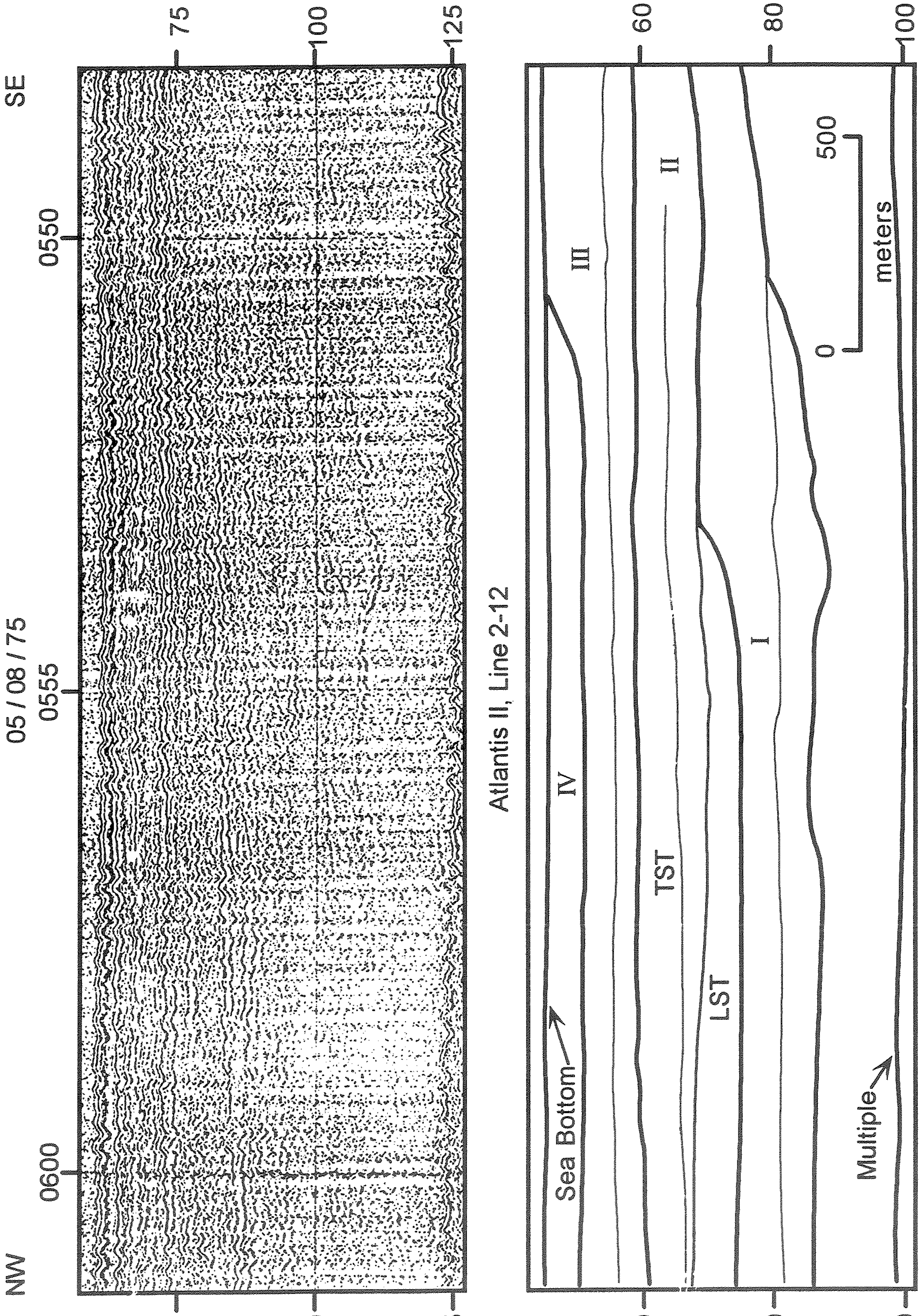
Two-way traveltime (milliseconds)
50
75
100
125
150
175
0100
2400
2300
2200
Line 2-12, southeastern part


05 / 08 / 75
NW
SE
1200
1100
1000
0900
0800
0700
Approximate Depth (m)
25
50
75
100
125
150
Hudson
Shelf
Valley
TST
LST
HST
I
II
III
IV
0
10
kilometers
Line 2-12, northwestern part
05 / 08 /75
0700
0600
0500
0400
0300
Mid-shelf Sediment Wedge
?
Line 2-12, central part

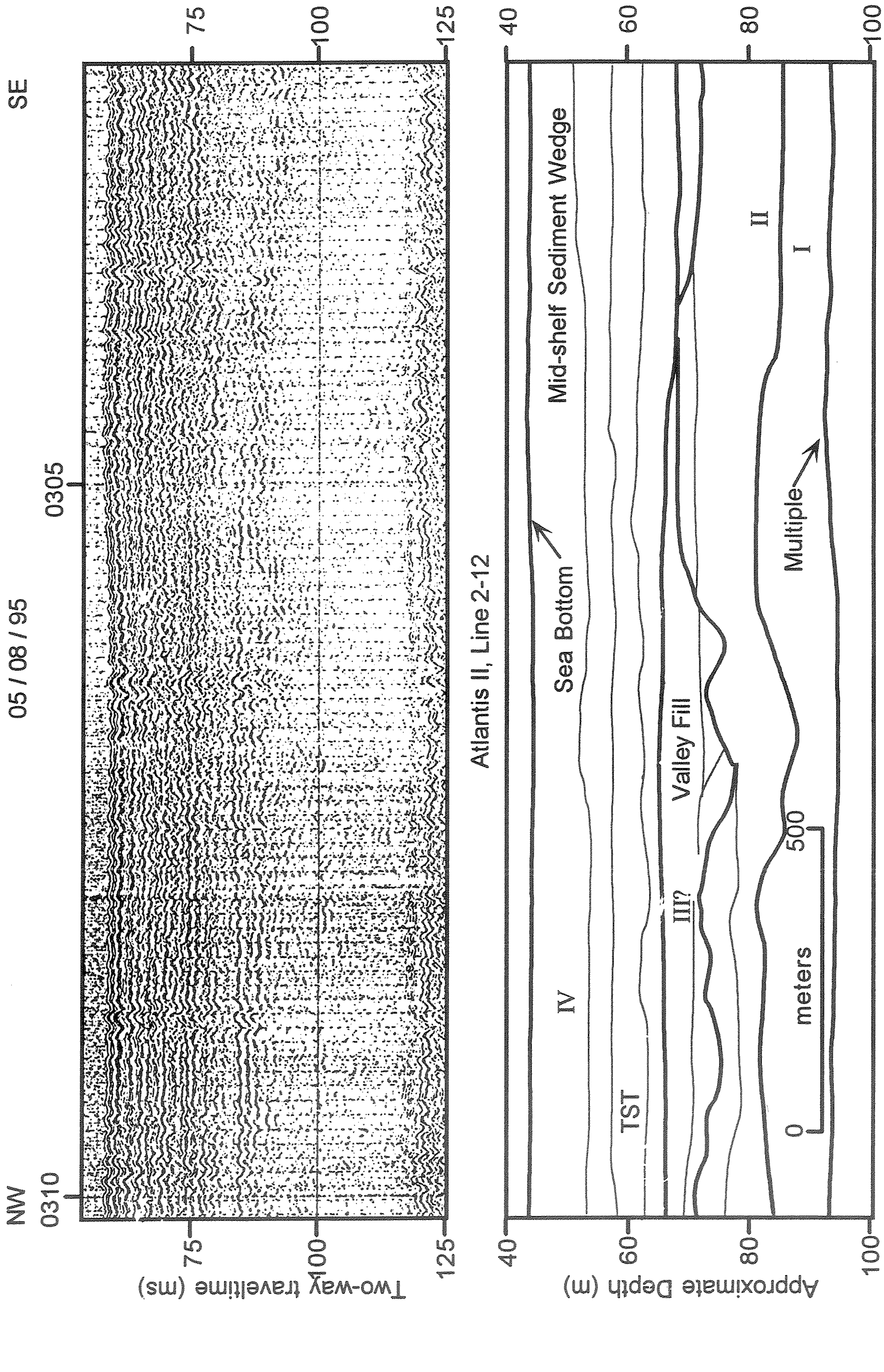

NW
SE
05/08/75
0100
2400
05 / 07 / 75
2300
Approximate Depth (meters)
50
75
100
125
IV
III
TST
LST
II
I
0
10
kilometers
Line 2-12, southeastern part


NW
SE
1150
05 / 08 /75
1155
50
75
100
Two-way traveltime (ms)
Atlantis II, Line 2-12
Sea Bottom
Ravinement Surface
Valley Fill Complex
IV
II
I
Multiple
0
50
kilometers
40
60
80
Approximate Depth (m)

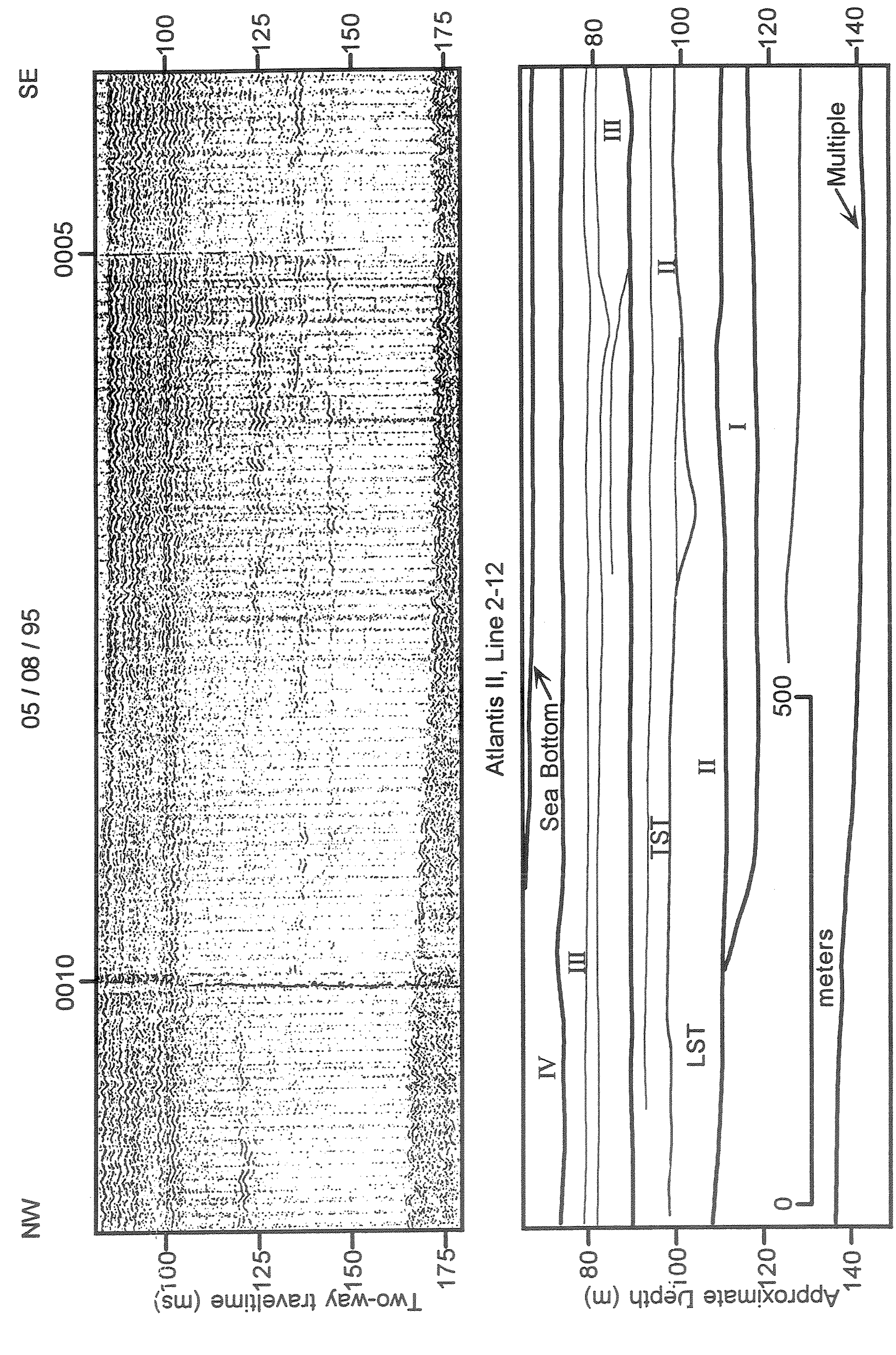

NW
05 / 08 / 75
SE
0830
0825
0820
Two-way traveltime (ms)
75
100
125
Atlantis II, Line 2-12
0
500
meters
Sea Bottom
IV
TST
Hudson Shelf-Valley Fill
III
II
I
Older sediments
Multiple
Approximate Depth (m)
60
80
100

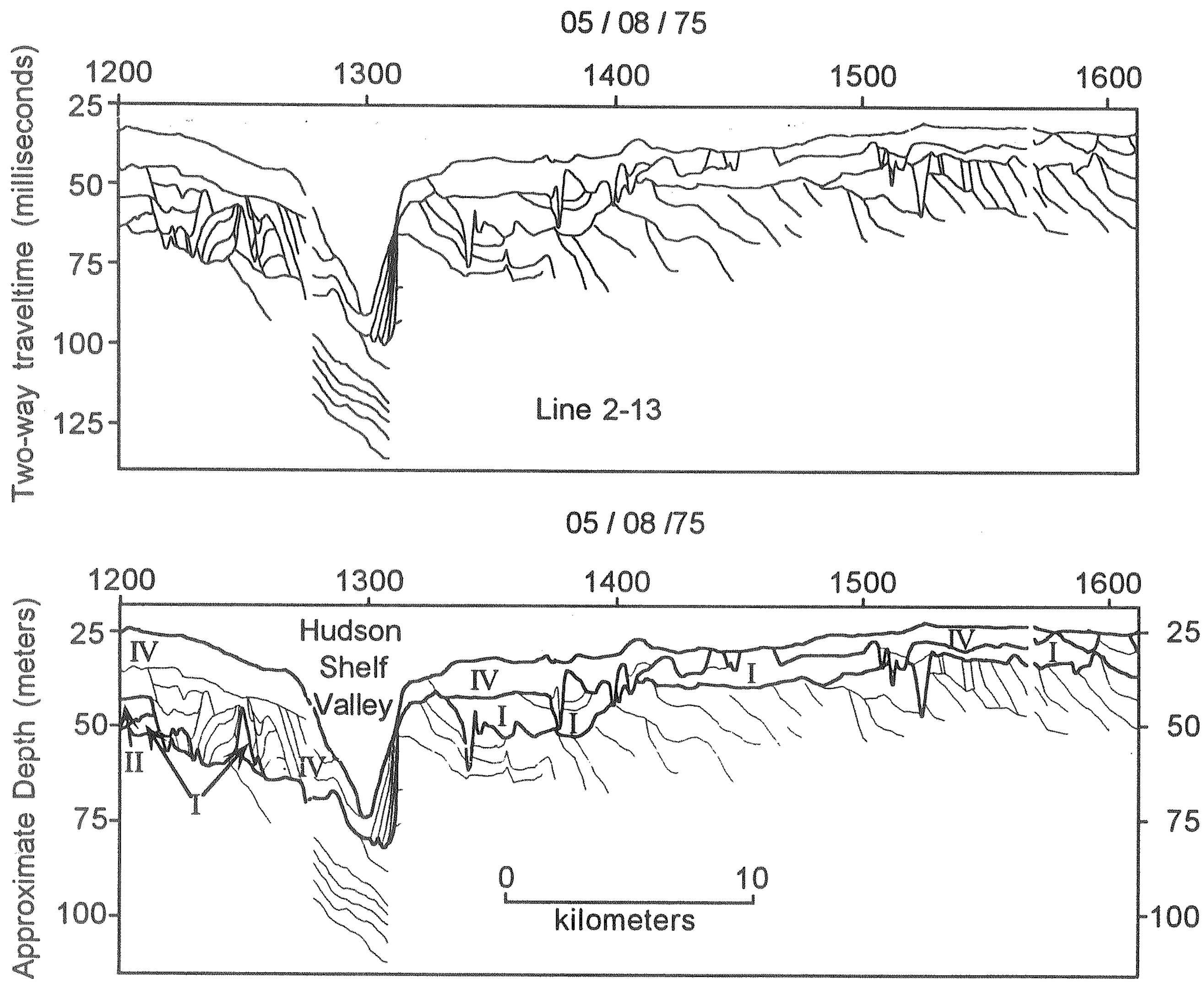


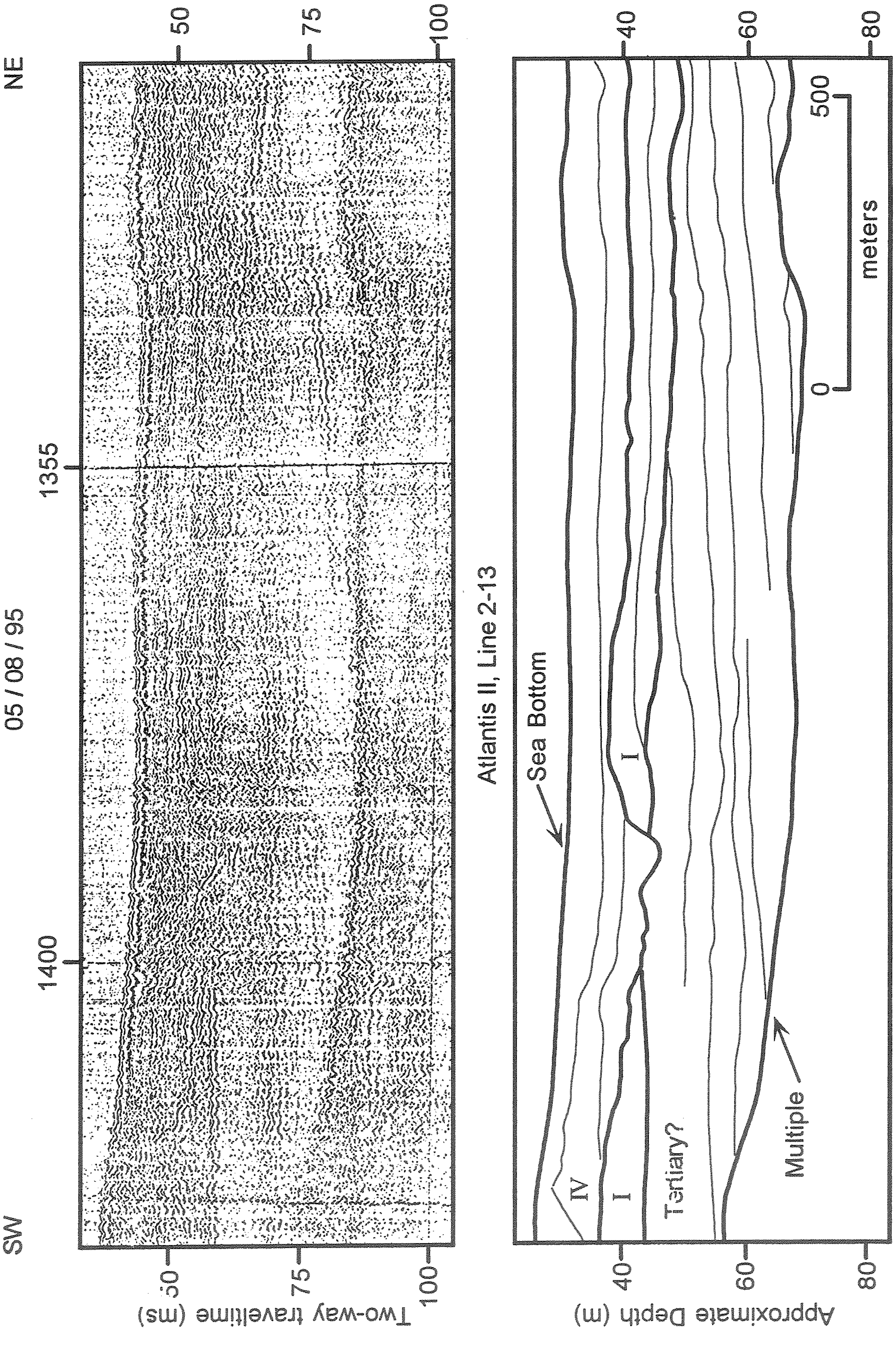
NW
SE
0600
0555
05 / 08 / 75
0550
75
100
125
Two-way traveltime (ms)
Atlantis II, Line 2-12
Approximate Depth (m)
60
80
100
Sea Bottom
Multiple
IV
III
II
I
TST
LST
0
500
meters

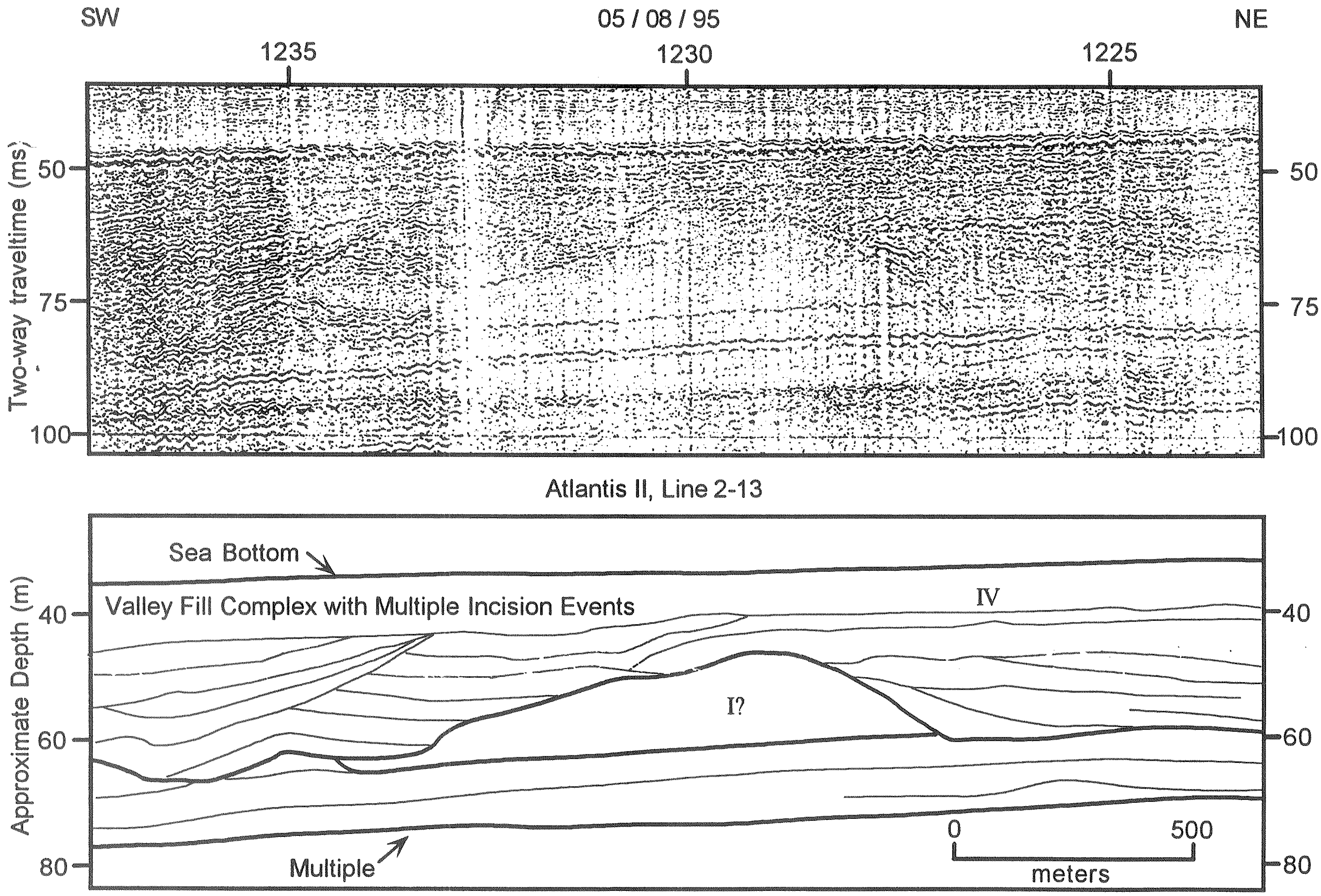


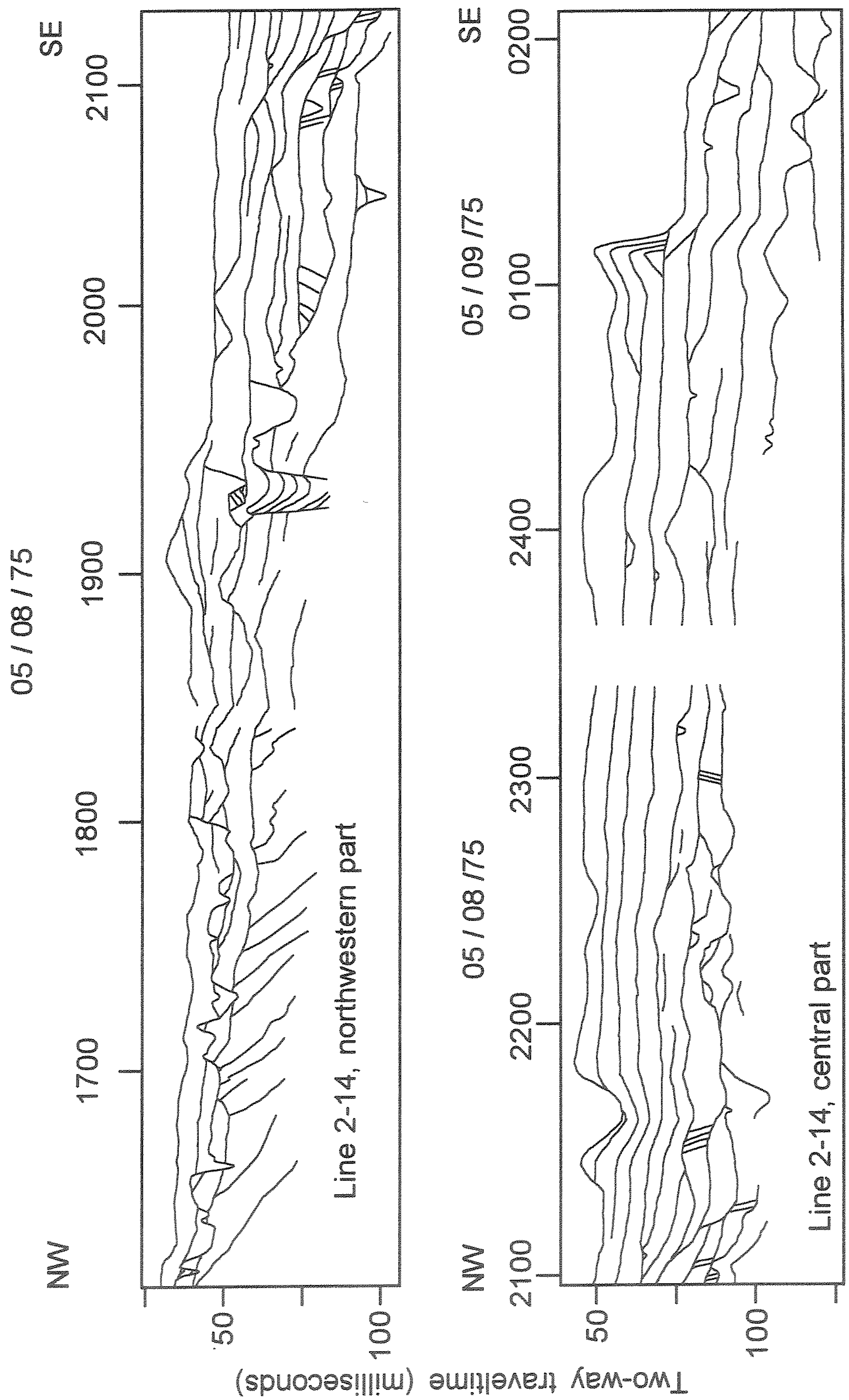
NW
SE
0310
0305
05 / 08 / 95
Two-way traveltime (ms)
75
100
125
Atlantis II, Line 2-12
Approximate Depth (m)
40
60
80
100
Mid-shelf Sediment Wedge
Sea Bottom
Valley Fill
TST
IV
III?
II
I
Multiple
0
500
meters

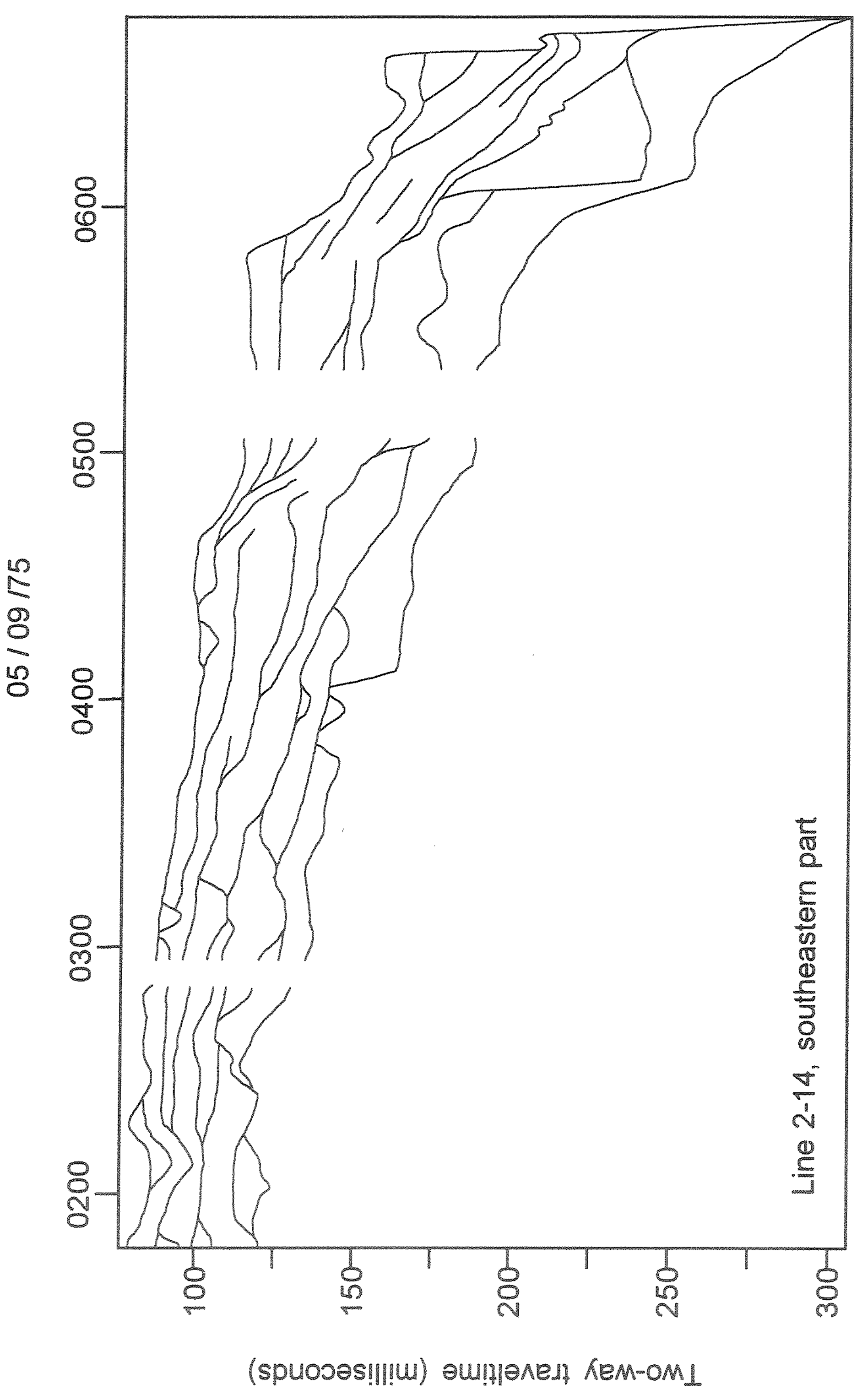

NW
05 / 08 / 95
SE
0010
0005
100
125
150
175
Two-way traveltime (ms)
Atlantis II, Line 2-12
Sea Bottom
IV
III
II
I
TST
LST
Multiple
0
500
meters
80
100
120
140
Approximate Depth (m)


05 / 08 / 75
1200
1300
1400
1500
1600
25
50
75
100
125
Two-way traveltime (milliseconds)
Line 2-13
05 / 08 /75
Hudson
Shelf
Valley
IV
II
I
Approximate Depth (meters)
0
10
kilometers

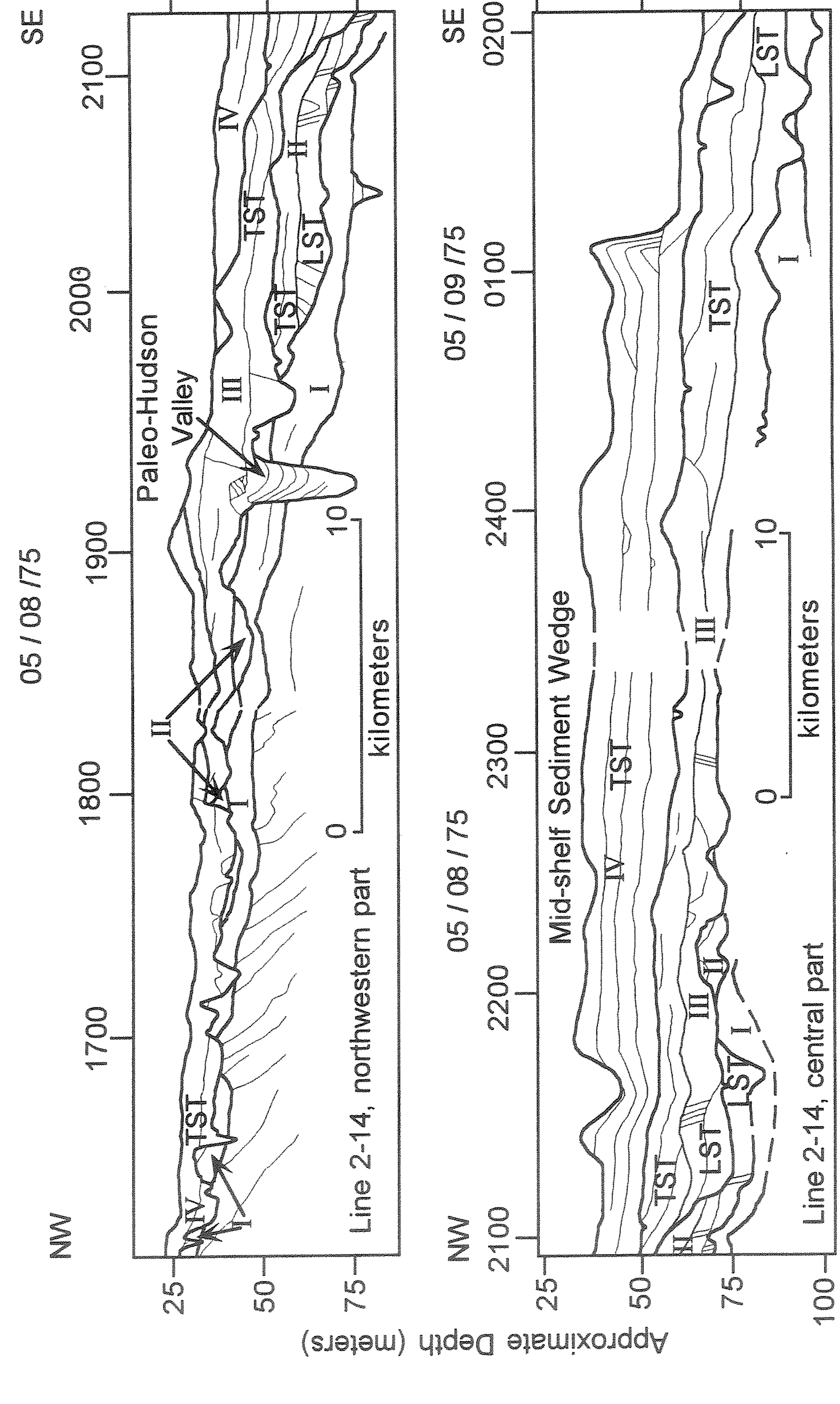

SW
NE
05 / 08 / 95
1400
1355
Two-way traveltime (ms)
50
75
100
Atlantis II, Line 2-13
Sea Bottom
IV
I
I
Tertiary?
Multiple
0
500
meters
Approximate Depth (m)
40
60
80


SW
05 / 08 / 95
NE
1235
1230
1225
Two-way traveltime (ms)
50
75
100
Atlantis II, Line 2-13
Sea Bottom
Valley Fill Complex with Multiple Incision Events
IV
I?
Approximate Depth (m)
40
60
80
Multiple
0
500
meters

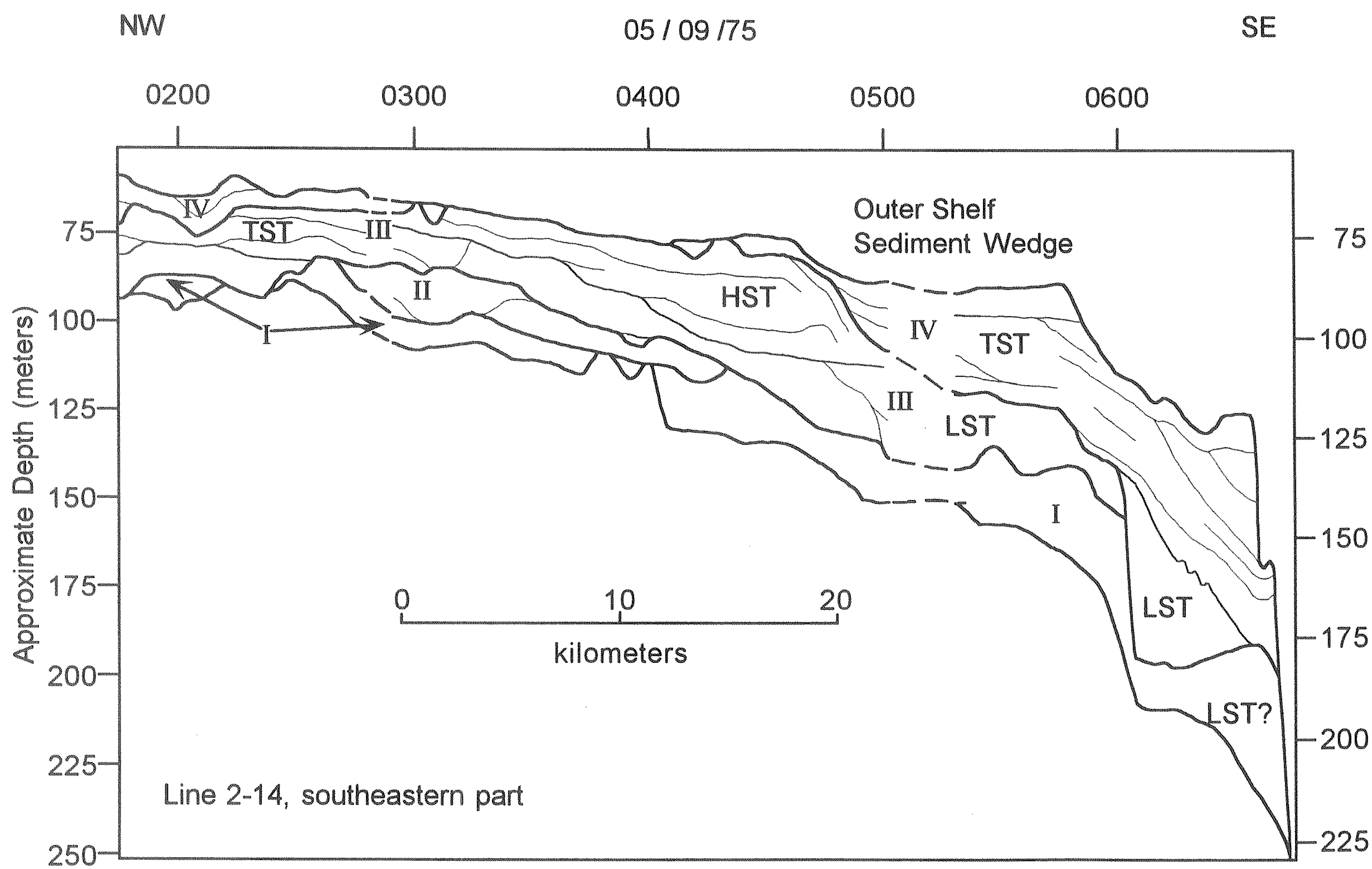

05 / 08 / 75
NW
1700
1800
1900
2000
2100
SE
Line 2-14, northwestern part
05 / 08 /75
05 / 09 /75
NW
2100
2200
2300
2400
0100
0200
SE
Line 2-14, central part
50
100
Two-way traveltime (milliseconds)



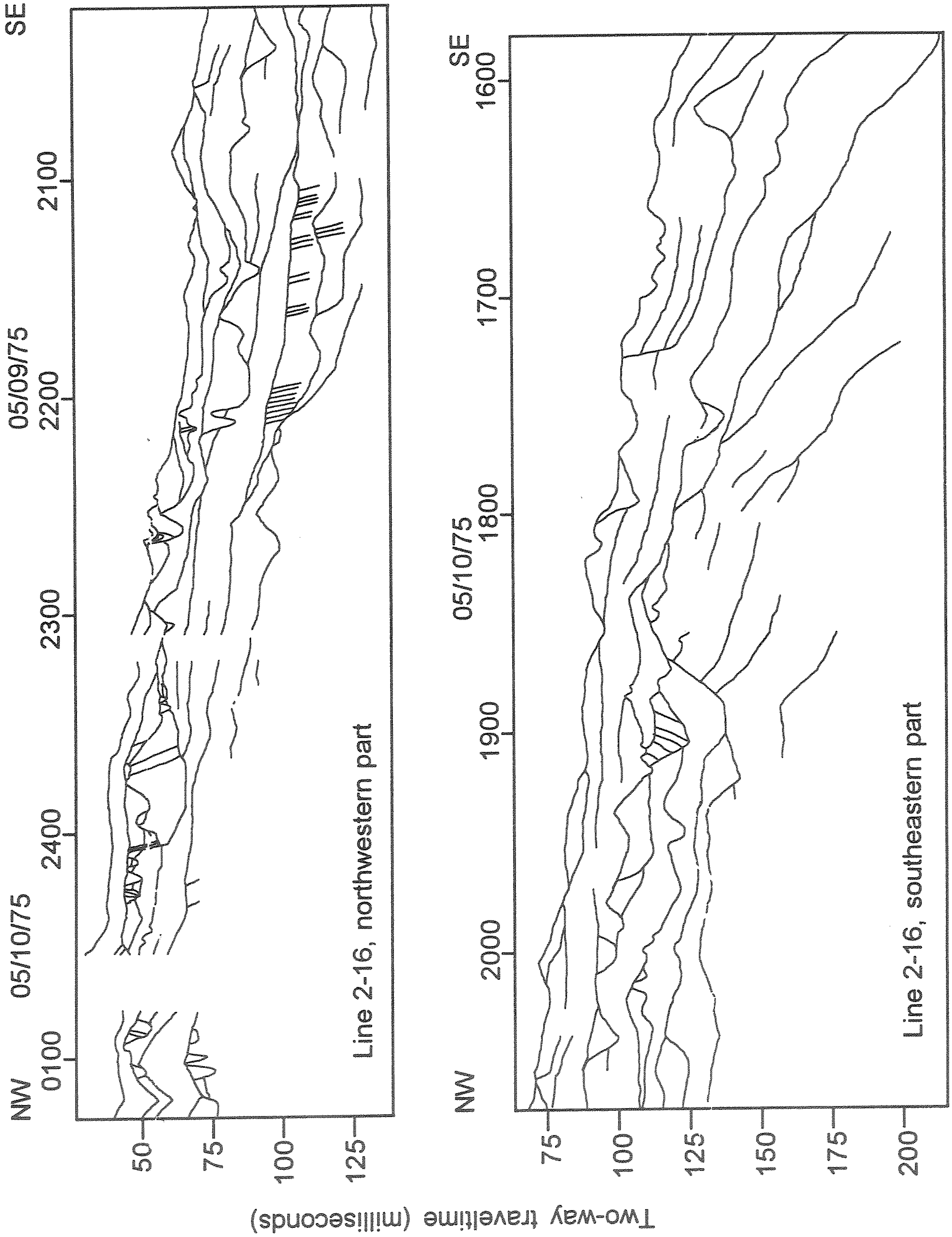
05 / 09 /75
Line 2-14, southeastern part
0200
0300
0400
0500
0600
100
150
200
250
300
Two-way traveltime (milliseconds)

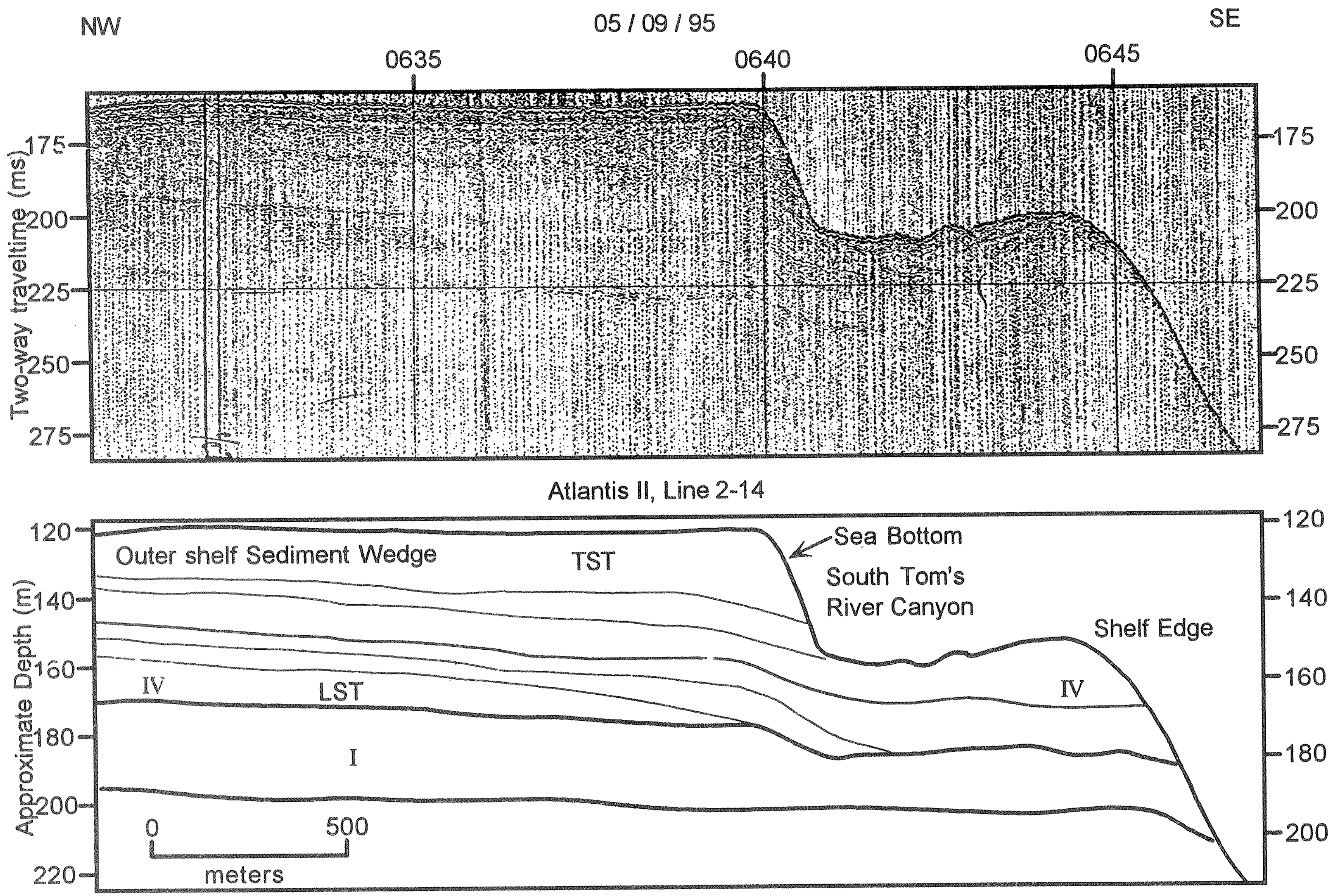


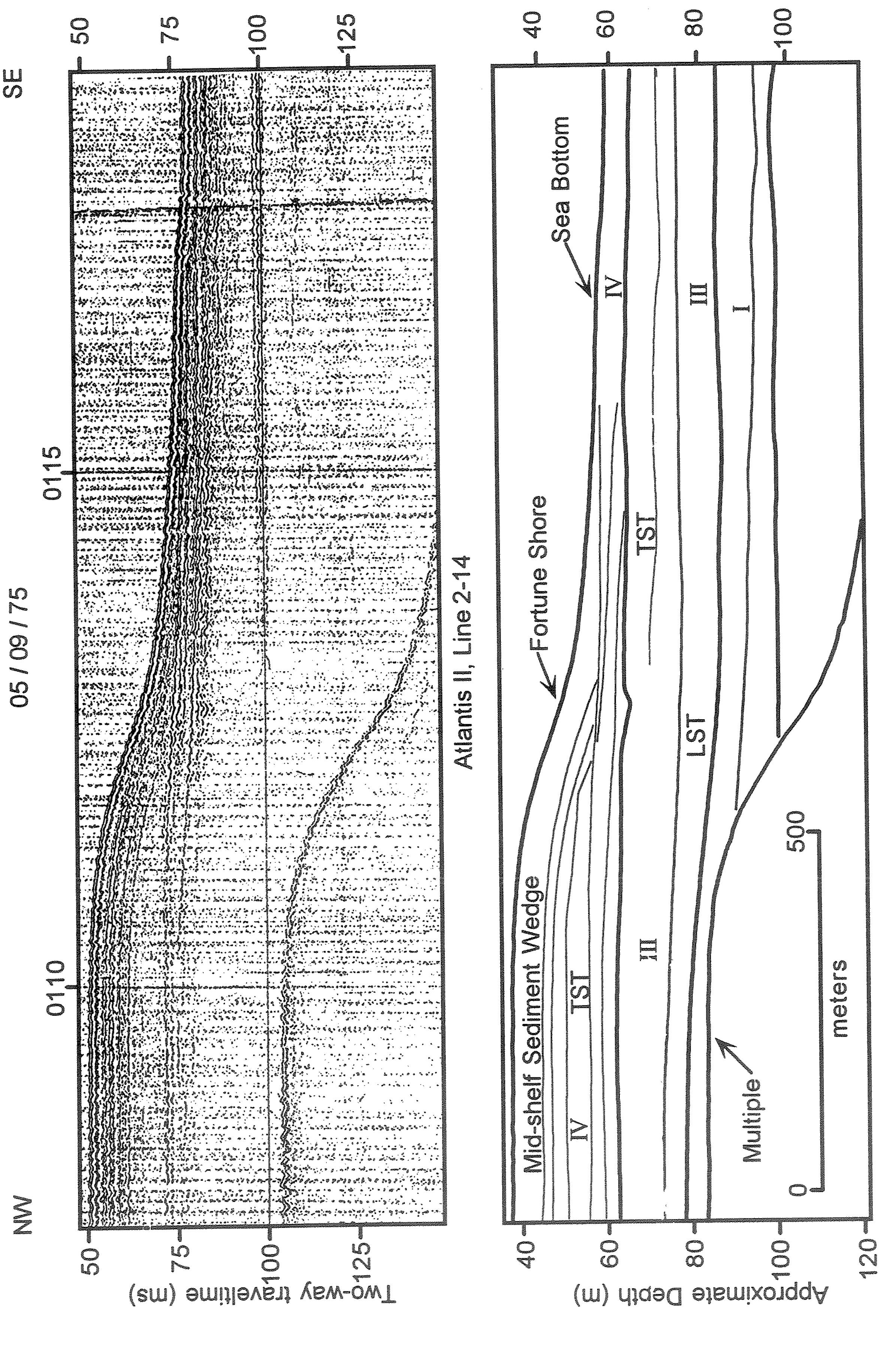
NW
SE
05 / 08 /75
Paleo-Hudson Valley
Line 2-14, northwestern part
kilometers
05 / 08 / 75
05 / 09 /75
Mid-shelf Sediment Wedge
Line 2-14, central part
kilometers
Approximate Depth (meters)

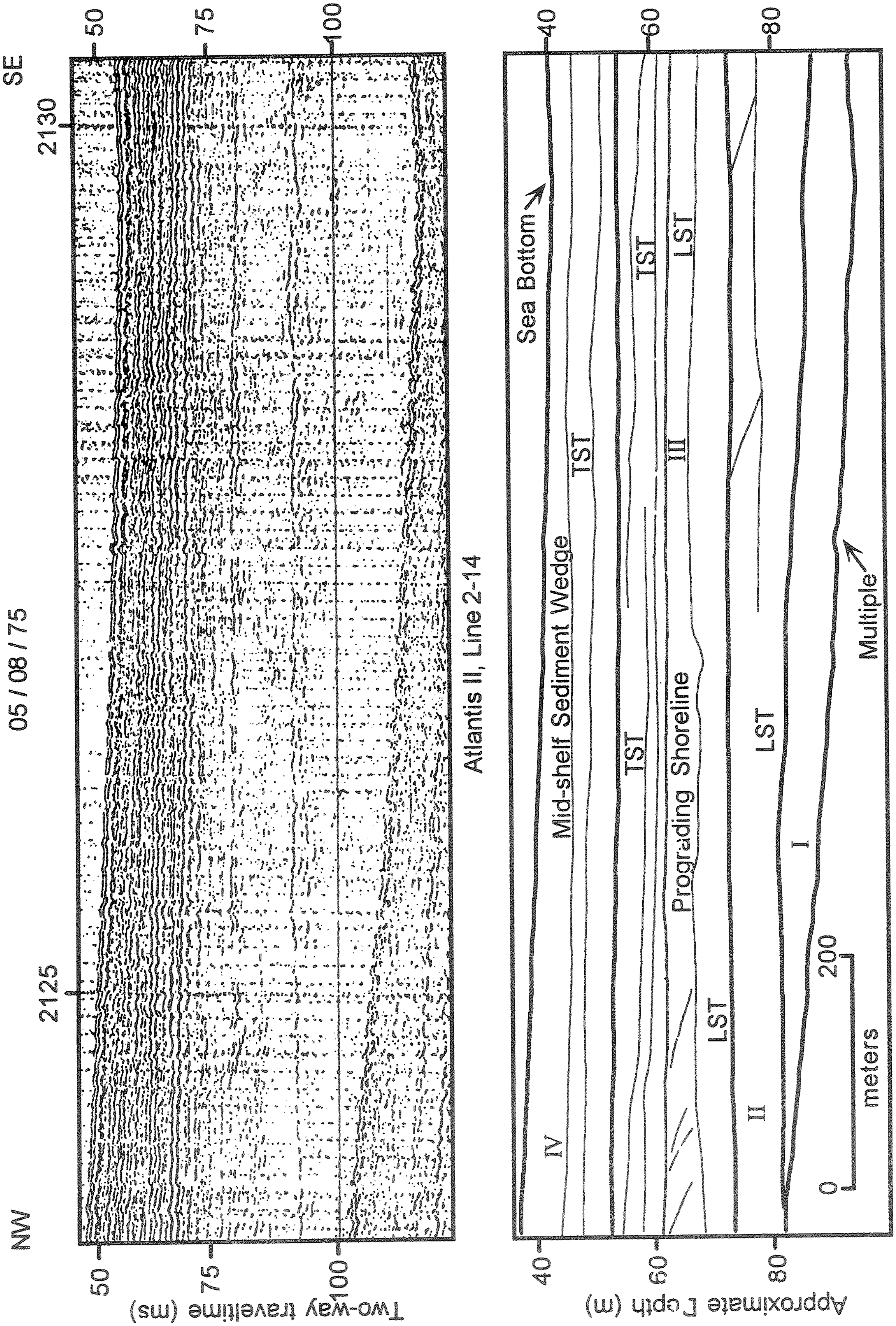


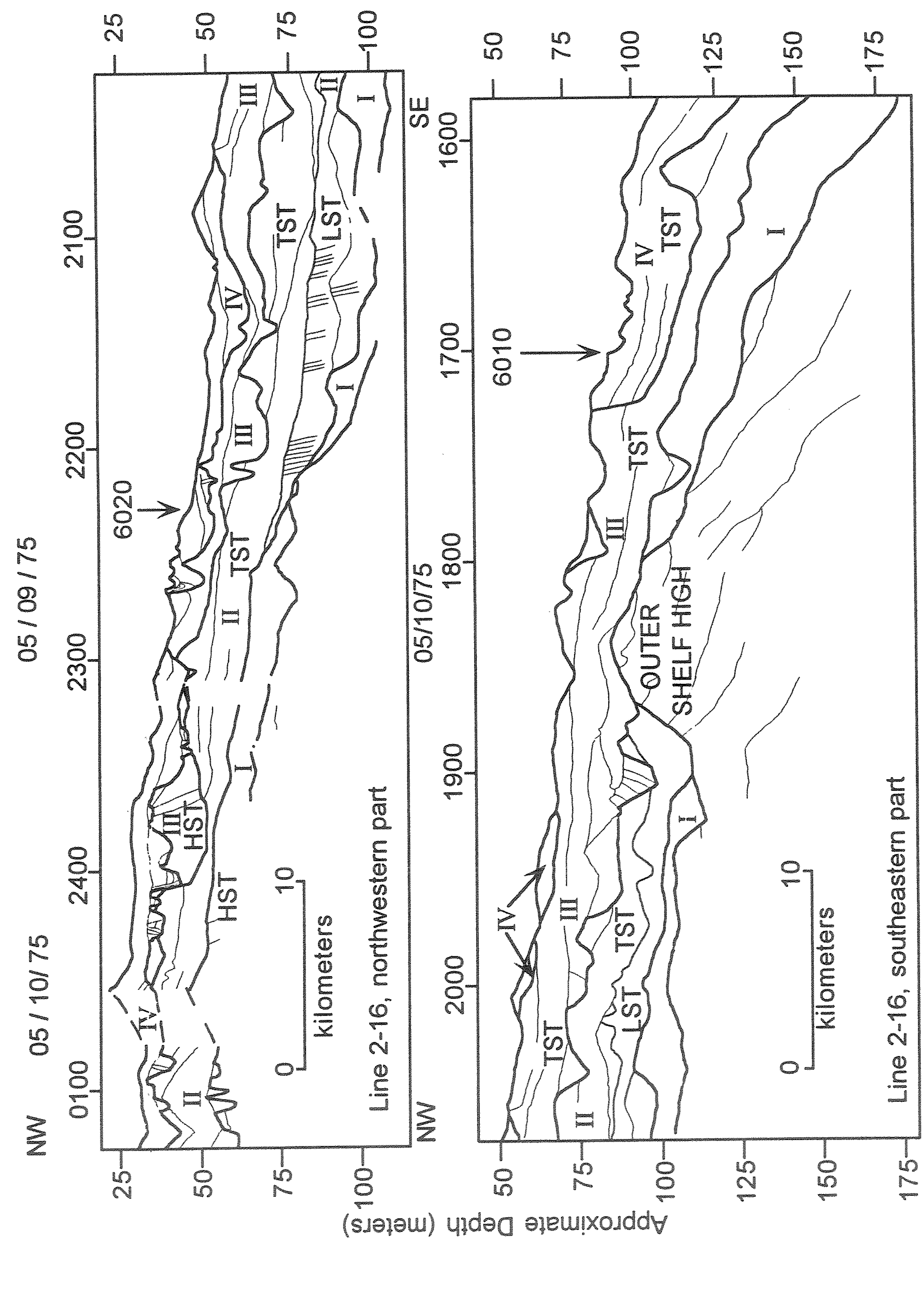
NW
05 / 09 /75
SE
0200
0300
0400
0500
0600
Outer Shelf
Sediment Wedge
IV
TST
III
II
I
HST
IV
TST
III
LST
I
LST
LST?
Approximate Depth (meters)
75
100
125
150
175
200
225
250
0
10
20
kilometers
Line 2-14, southeastern part


NW
0100
05/10/75
2400
2300
05/09/75
2200
2100
SE
Line 2-16, northwestern part
50
75
100
125
NW
2000
1900
05/10/75
1800
1700
1600
SE
Line 2-16, southeastern part
75
100
125
150
175
200
Two-way traveltime (milliseconds)

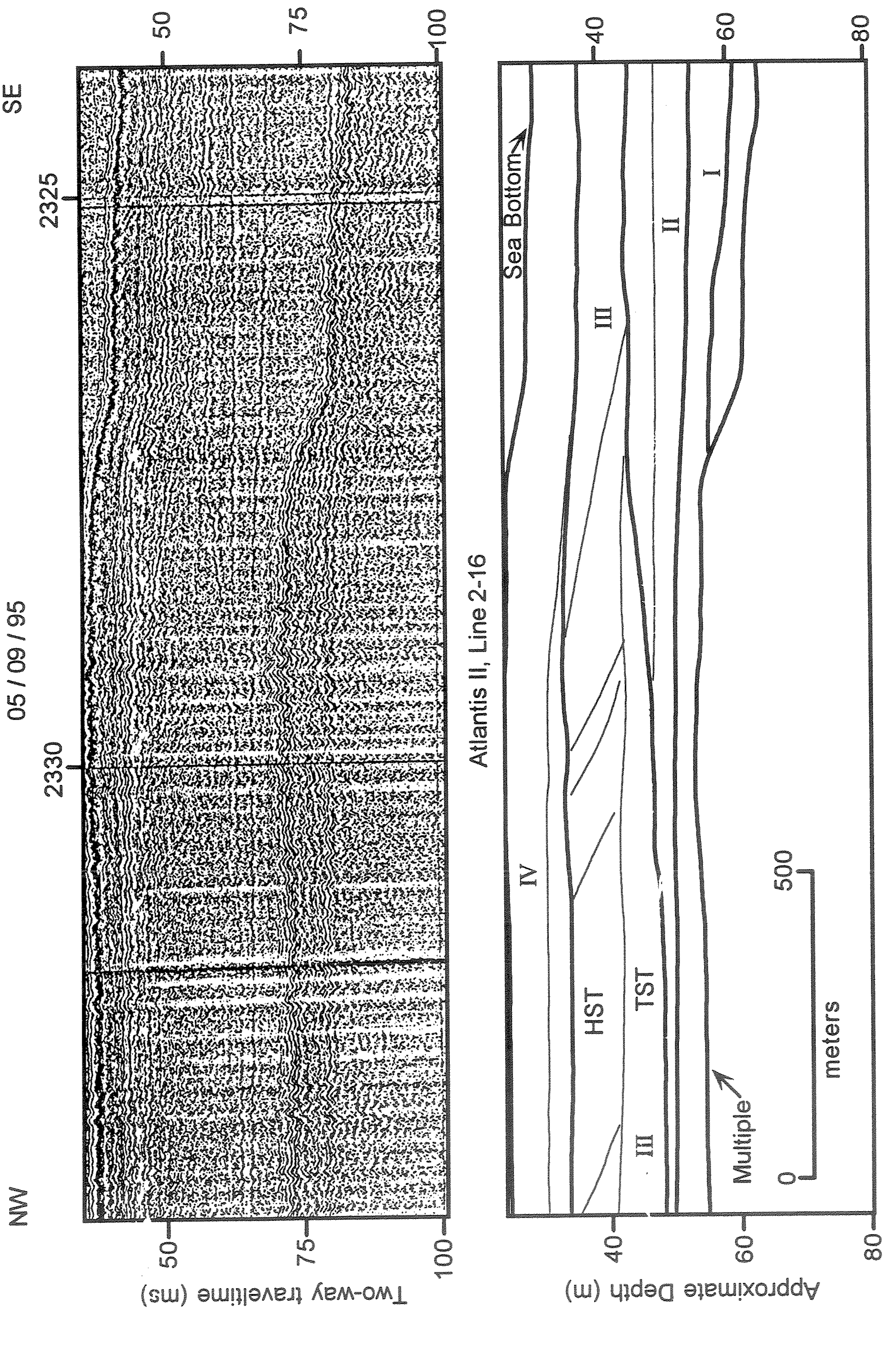

NW
05 / 09 / 95
SE
0635
0640
0645
Two-way traveltime (ms)
175
200
225
250
275
Atlantis II, Line 2-14
Approximate Depth (m)
120
140
160
180
200
220
Outer shelf Sediment Wedge
TST
Sea Bottom
South Tom's
River Canyon
Shelf Edge
IV
LST
IV
I
0
500
meters

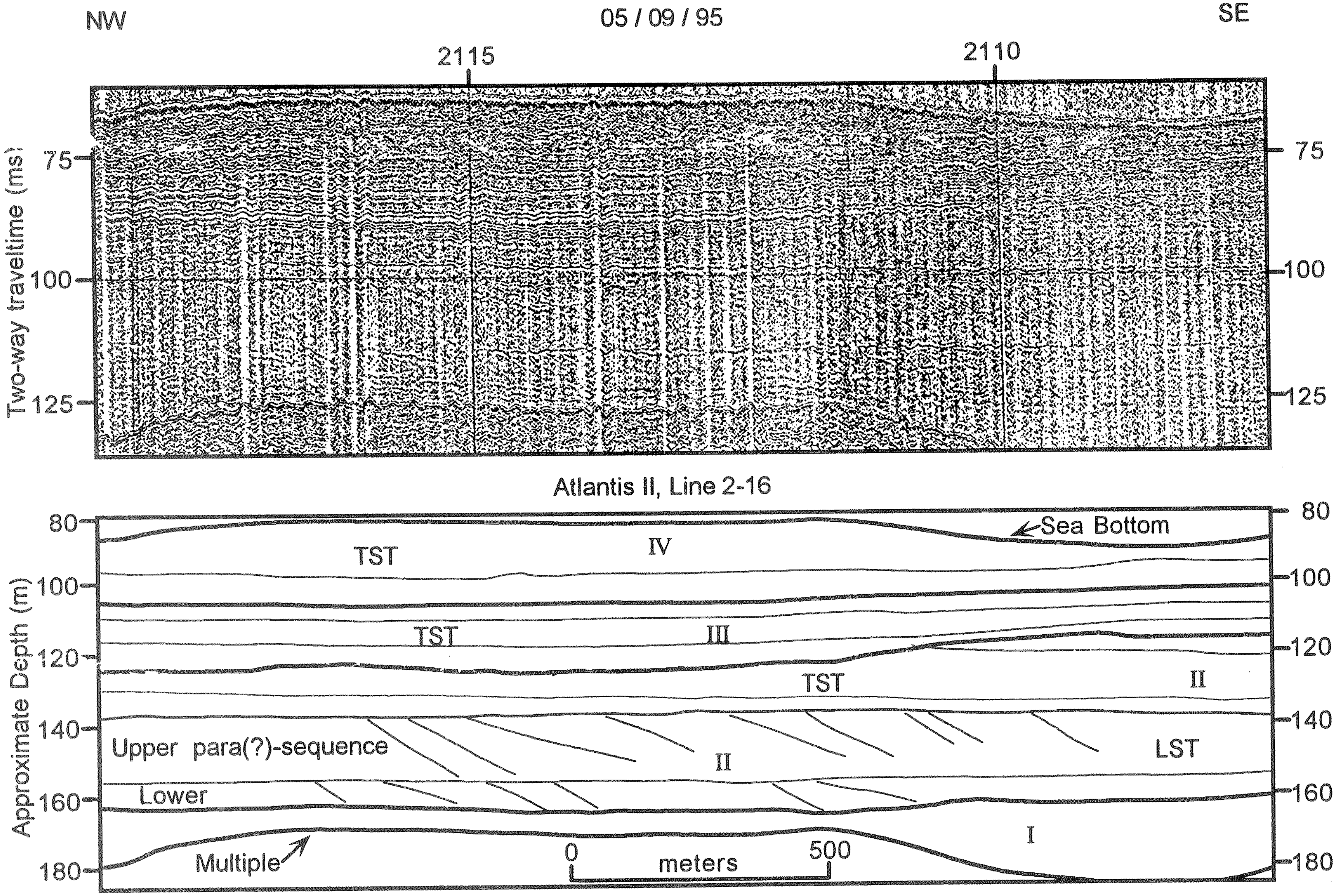

NW
SE
05 / 09 / 75
0110
0115
Two-way traveltime (ms)
50
75
100
125
Atlantis II, Line 2-14
Mid-shelf Sediment Wedge
Fortune Shore
Sea Bottom
IV
TST
III
LST
I
Multiple
0
500
meters
Approximate Depth (m)
40
60
80
100
120

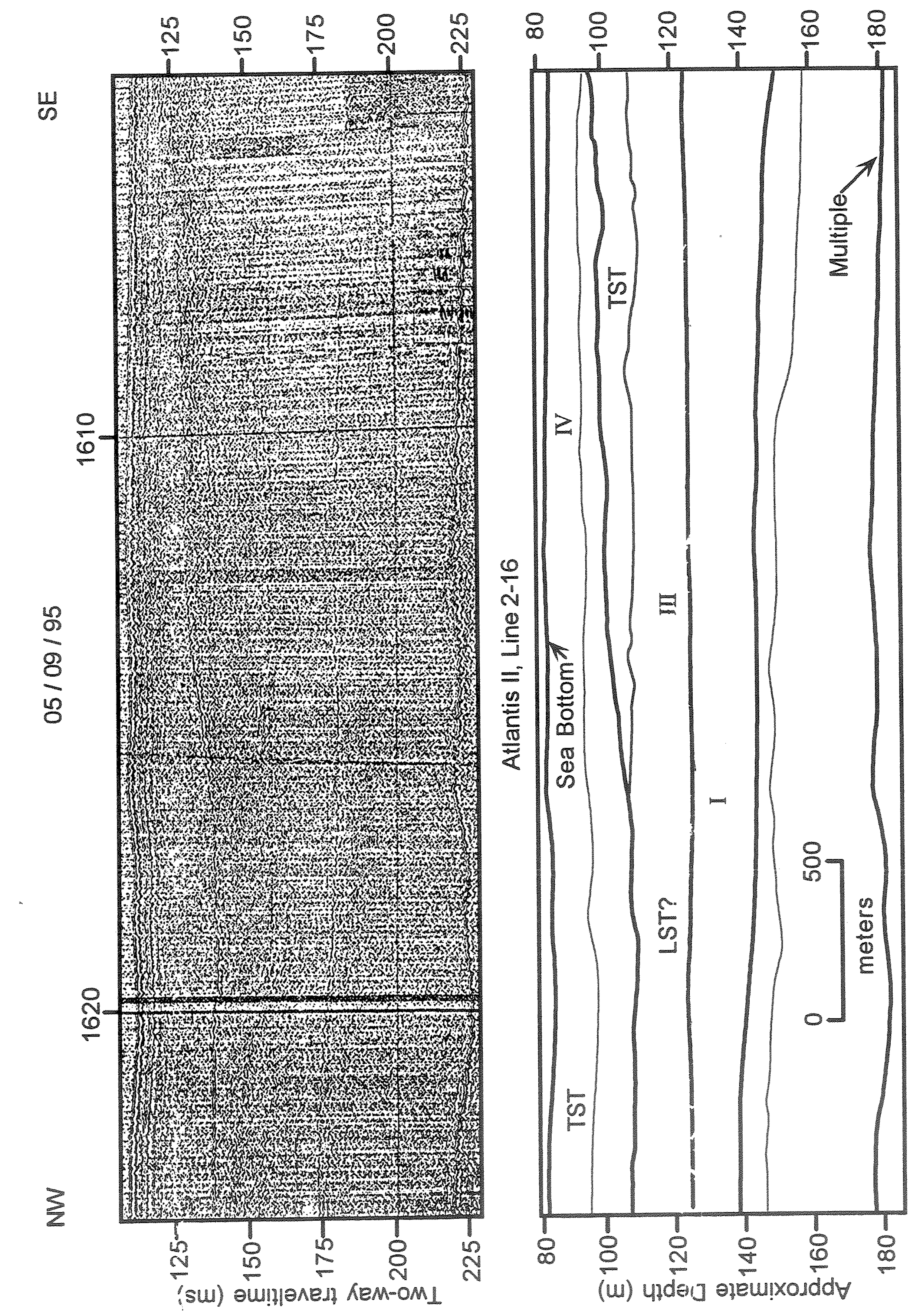

NW
SE
2125
2130
05 / 08 / 75
50
75
100
Two-way traveltime (ms)
Atlantis II, Line 2-14
Sea Bottom
Mid-shelf Sediment Wedge
TST
TST
TST
LST
LST
LST
LST
Prograding Shoreline
IV
III
II
I
Multiple
0
200
meters
40
60
80
Approximate Depth (m)

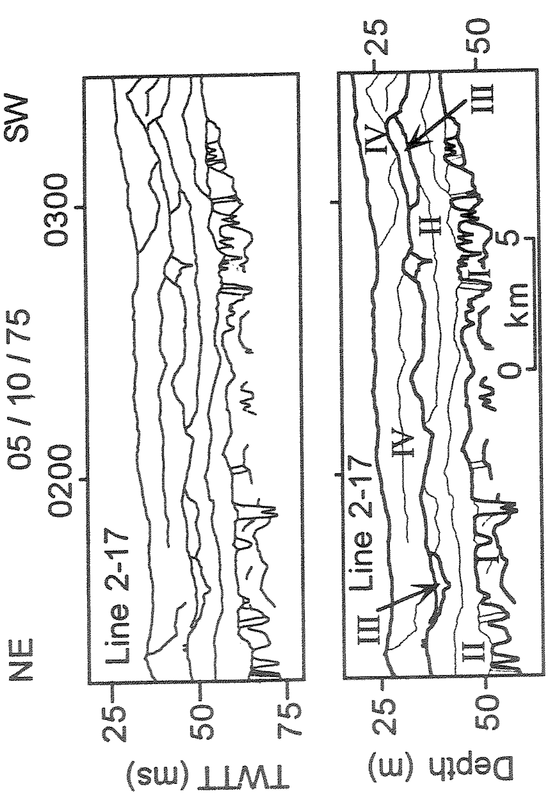


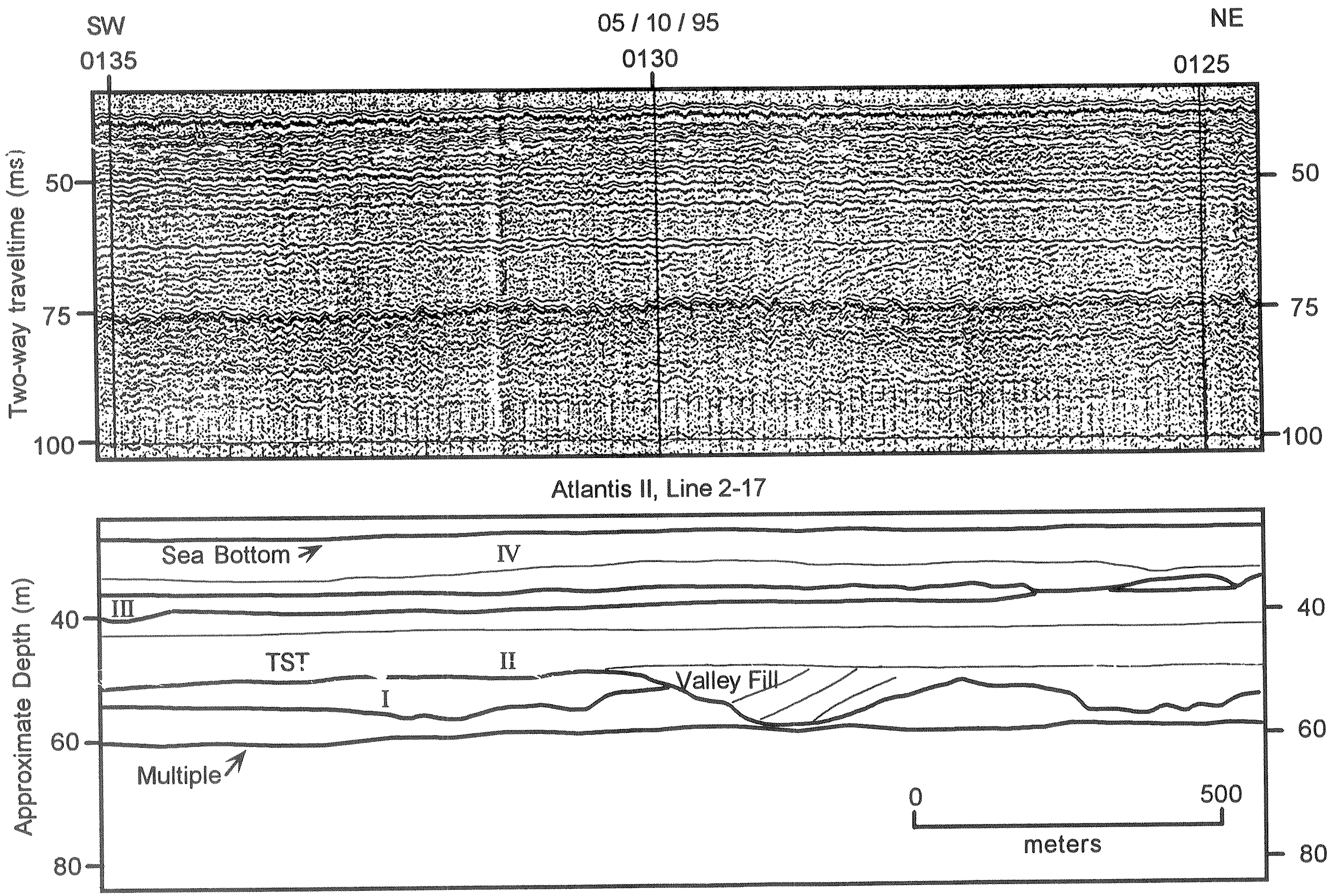
NW
SE
05 / 09 / 75
05 / 10/ 75
05/10/75
6020
6010
OUTER SHELF HIGH
kilometers
Line 2-16, northwestern part
Line 2-16, southeastern part
Approximate Depth (meters)

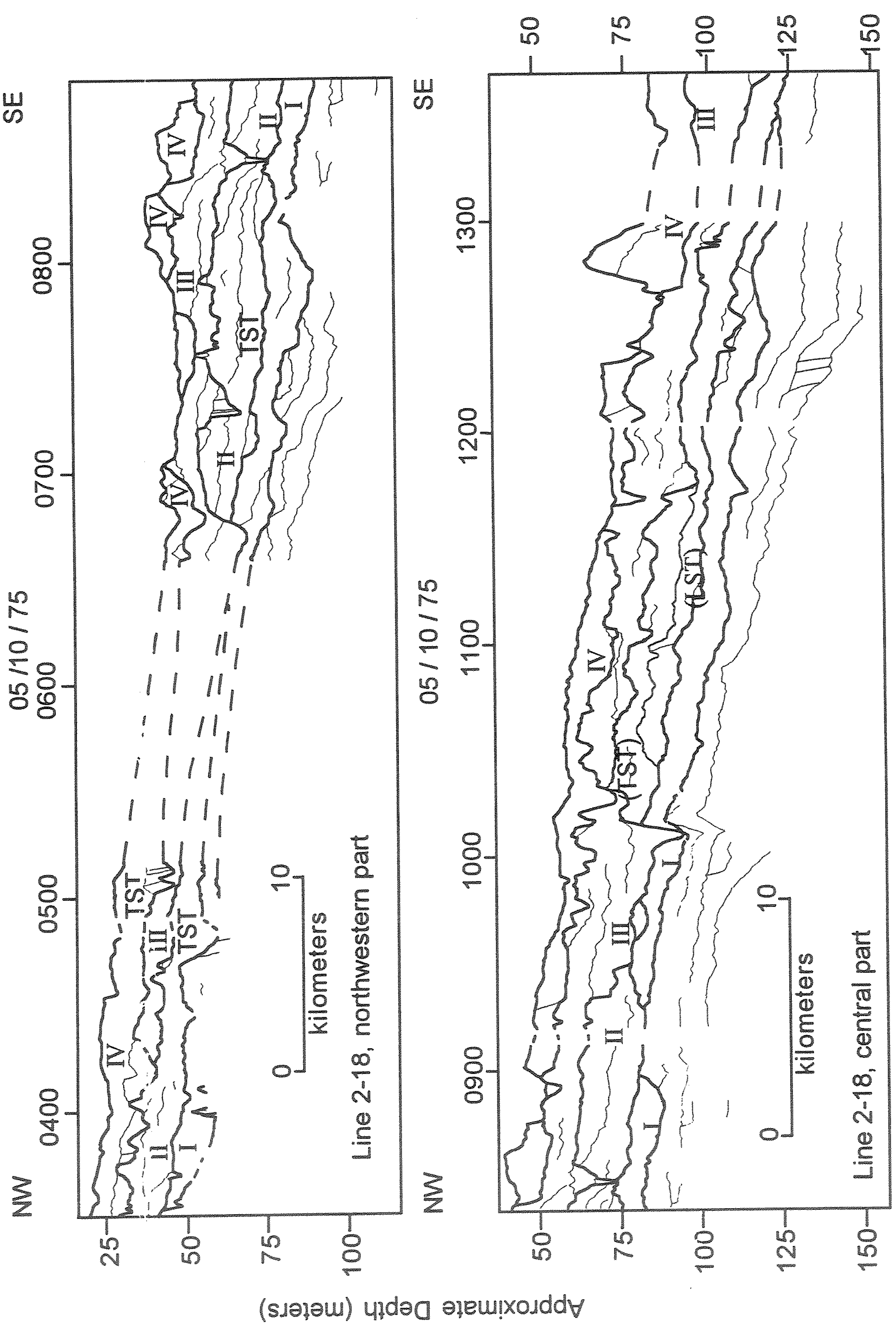

NW
SE
05 / 09 / 95
2325
2330
50
75
100
Two-way traveltime (ms)
Atlantis II, Line 2-16
Sea Bottom
IV
III
II
I
HST
TST
Multiple
0
500
meters
40
60
80
Approximate Depth (m)

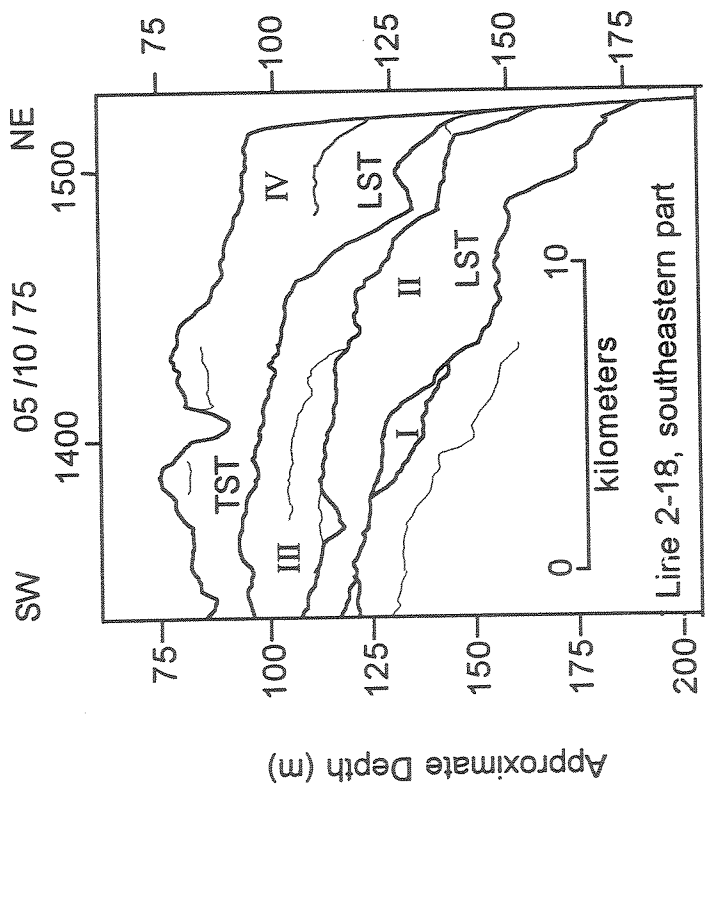

NW
05 / 09 / 95
SE
2115
2110
Two-way traveltime (ms)
75
100
125
Atlantis II, Line 2-16
Approximate Depth (m)
80
100
120
140
160
180
Sea Bottom
TST
IV
TST
III
TST
II
Upper para(?)-sequence
II
LST
Lower
I
Multiple
0
meters
500

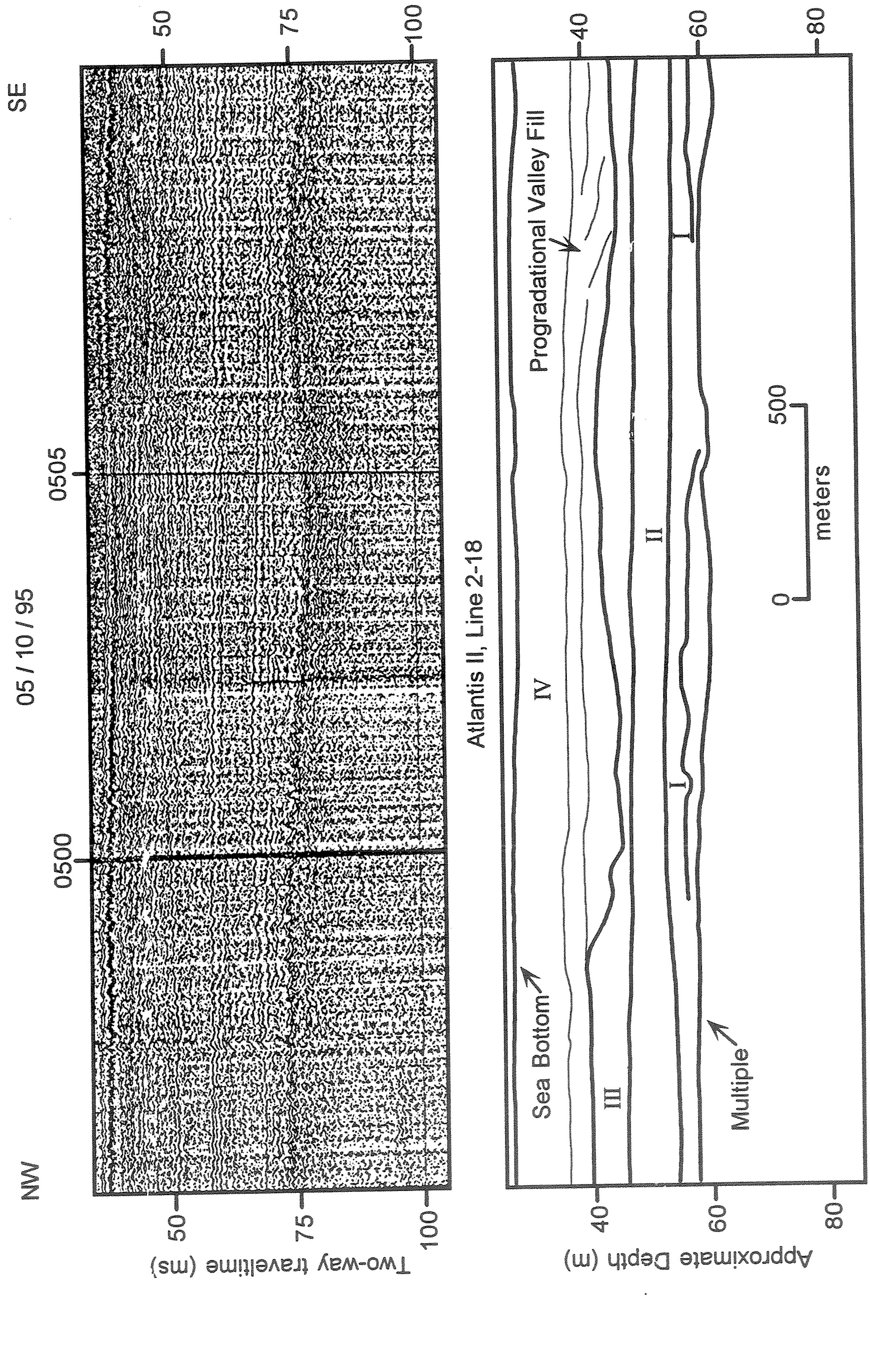

NW
SE
1620
1610
05 / 09 / 95
125
150
175
200
225
Two-way traveltime (ms)
Atlantis II, Line 2-16
80
100
120
140
160
180
Approximate Depth (m)
Sea Bottom
Multiple
TST
TST
LST?
IV
III
I
0
500
meters

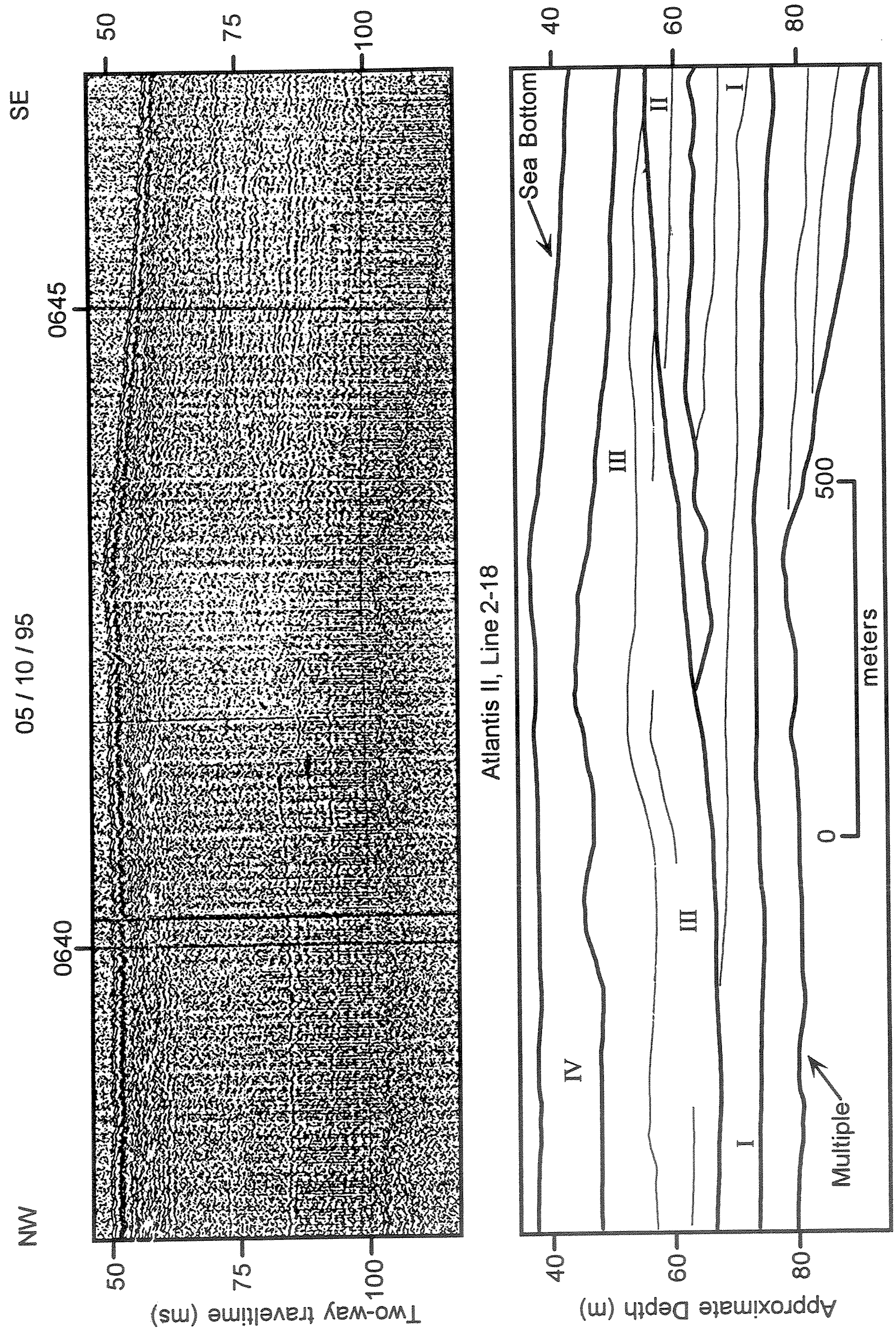

NE
05 / 10 / 75
SW
0200
0300
Line 2-17
TWTT (ms)
25
50
75
Line 2-17
Depth (m)
25
50
III
IV
II
I
0 km 5

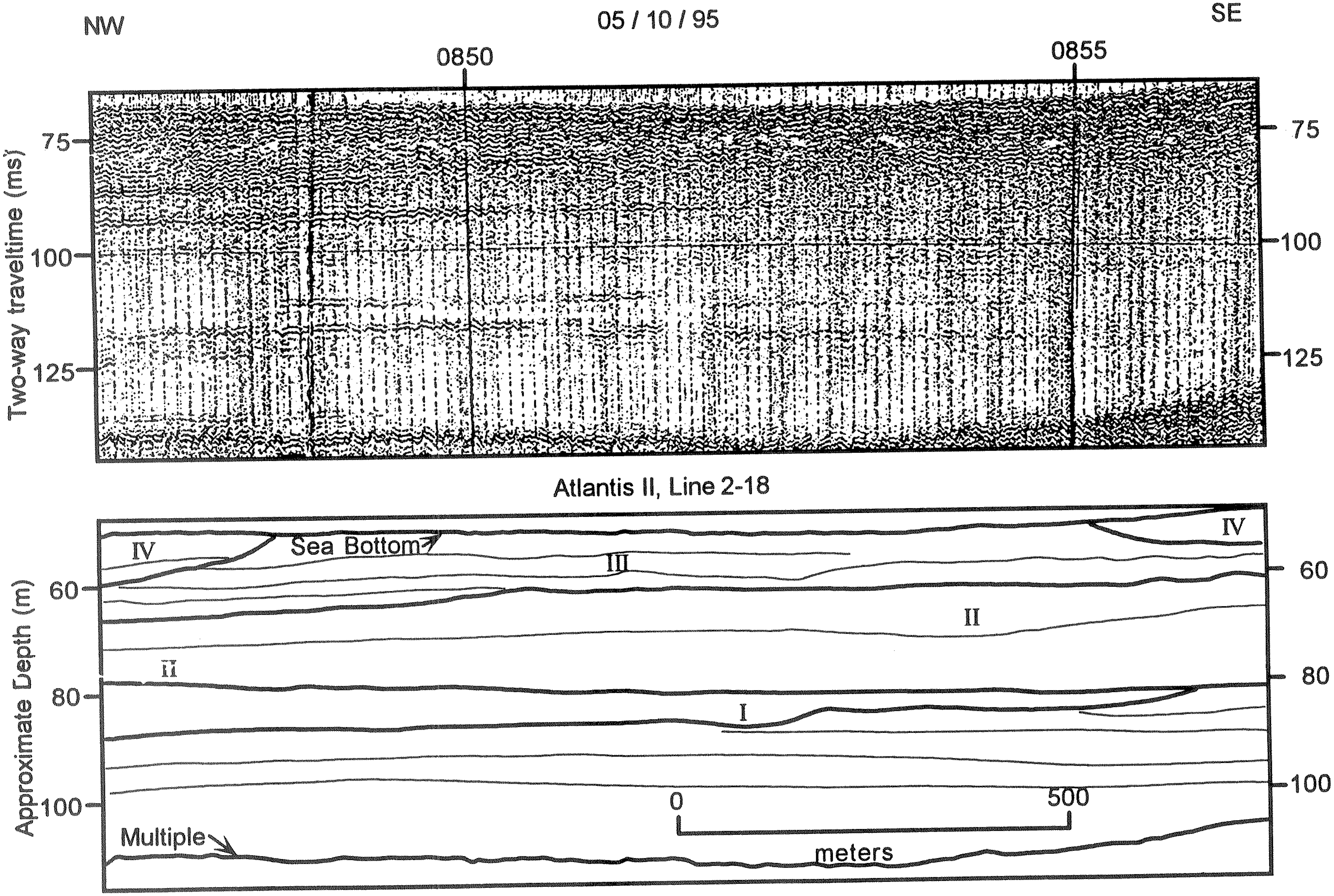

SW
05 / 10 / 95
NE
0135
0130
0125
Two-way traveltime (ms)
50
75
100
Atlantis II, Line 2-17
Sea Bottom
IV
III
TST
II
I
Valley Fill
Multiple
Approximate Depth (m)
40
60
80
0
500
meters

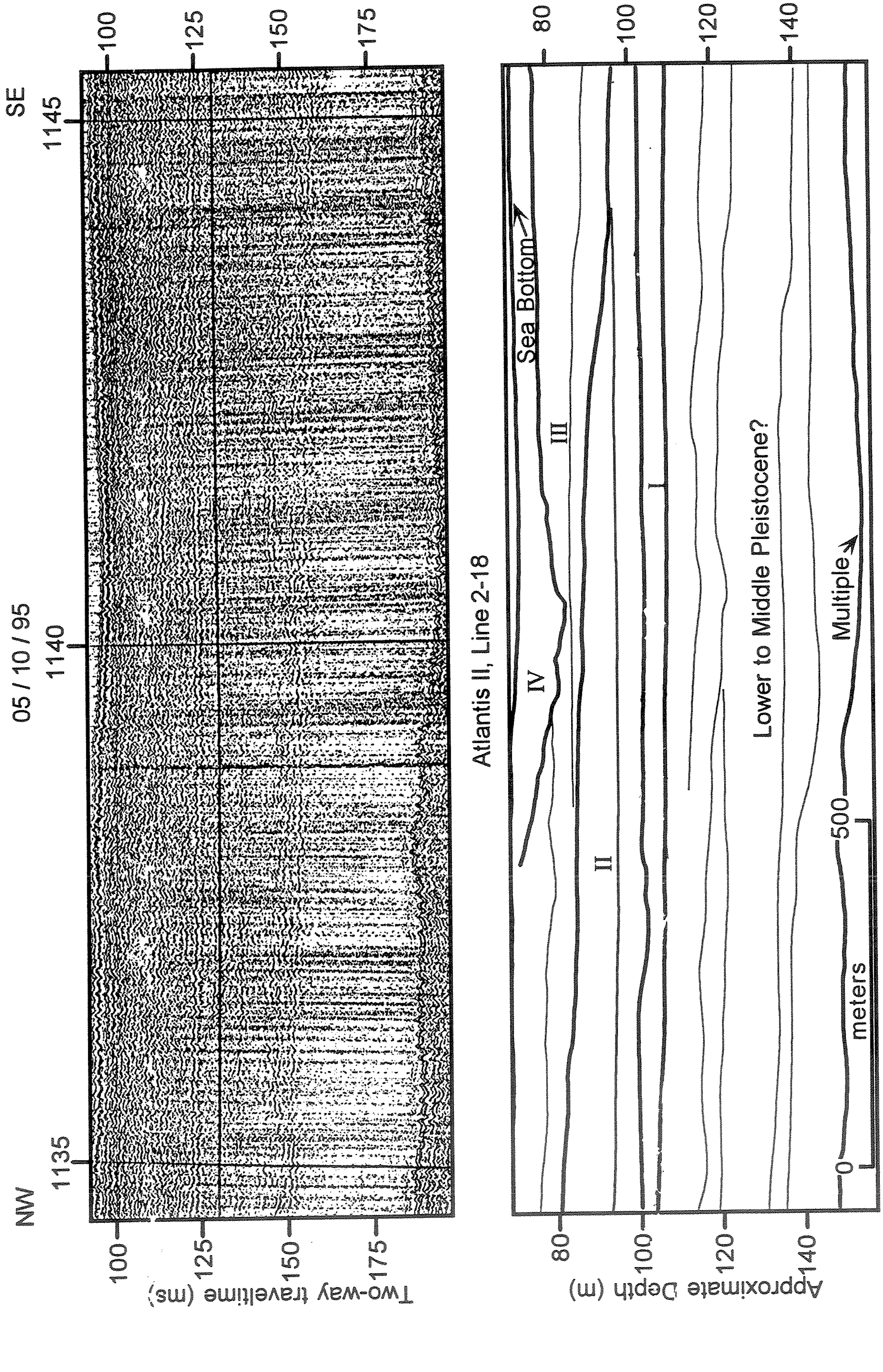

NW
SE
05/10/75
0400
0500
0600
0700
0800
25
50
75
100
IV
III
II
I
TST
0
10
kilometers
Line 2-18, northwestern part
NW
SE
05/10/75
0900
1000
1100
1200
1300
50
75
100
125
150
(TST)
(LST)
kilometers
Line 2-18, central part
Approximate Depth (meters)


SW
NE
1400
05 / 10 / 75
1500
75
100
125
150
175
200
Approximate Depth (m)
TST
IV
III
LST
II
LST
I
0
10
kilometers
Line 2-18, southeastern part

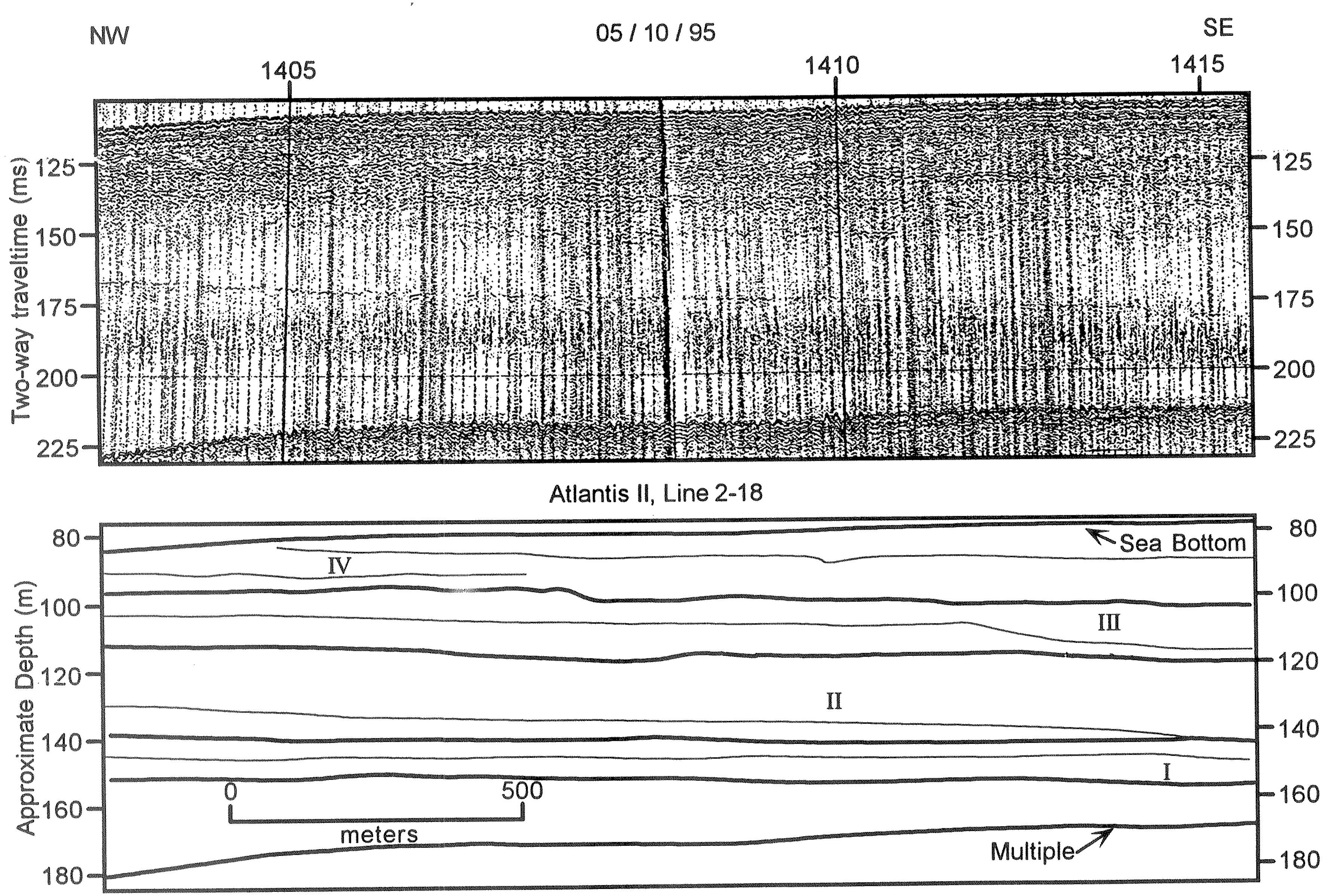

NW
SE
05 / 10 / 95
0500
0505
50
75
100
Two-way traveltime (ms)
Atlantis II, Line 2-18
Sea Bottom
Progradational Valley Fill
Multiple
IV
III
II
I
I
0
500
meters
40
60
80
Approximate Depth (m)

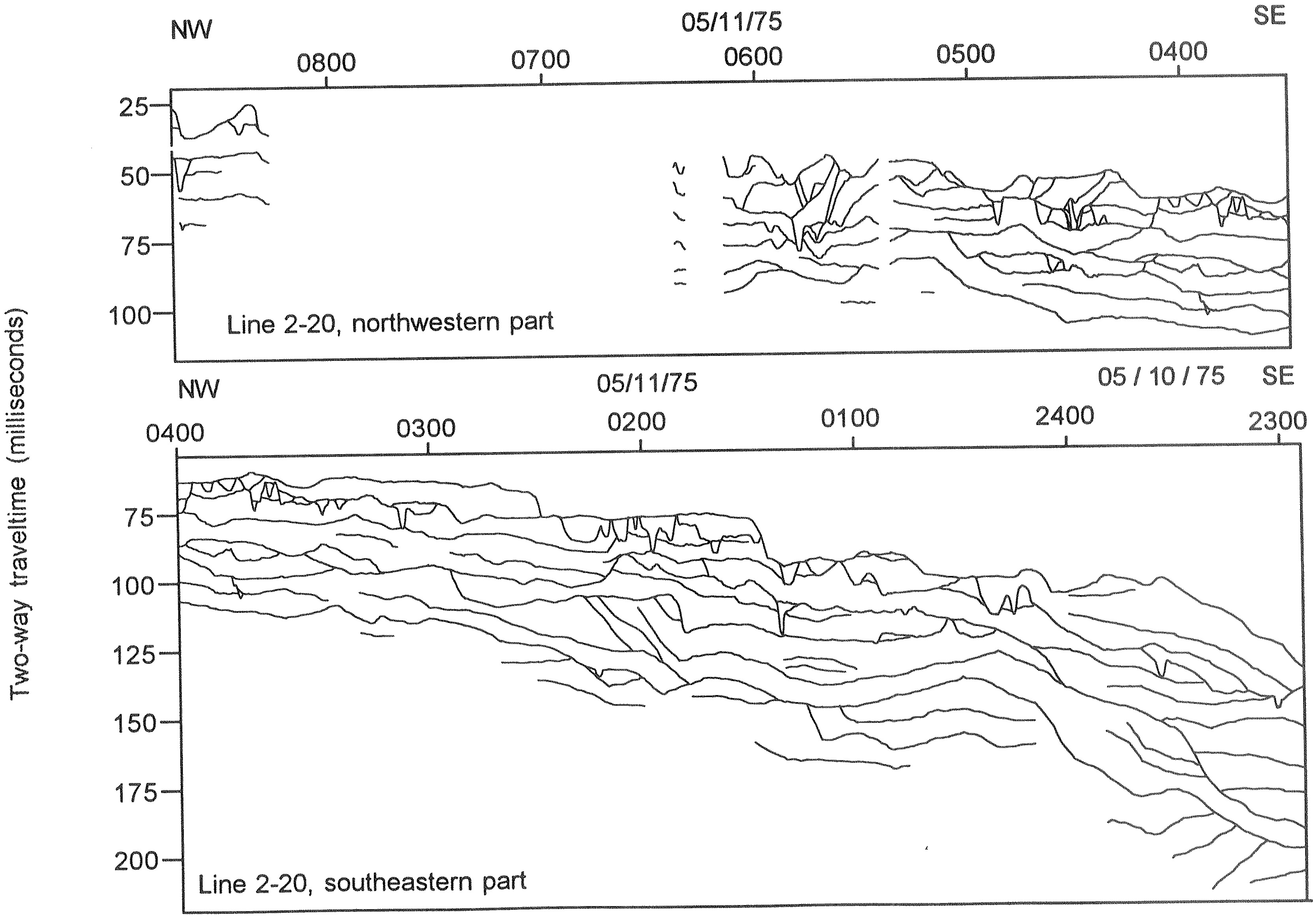

NW
SE
05 / 10 / 95
0640
0645
50
75
100
Two-way traveltime (ms)
Atlantis II, Line 2-18
Sea Bottom
IV
III
II
I
Multiple
0
500
meters
40
60
80
Approximate Depth (m)


NW
05 / 10 / 95
SE
0850
0855
75
100
125
Two-way traveltime (ms)
Atlantis II, Line 2-18
IV
Sea Bottom
III
II
I
60
80
100
Approximate Depth (m)
0
500
meters
Multiple

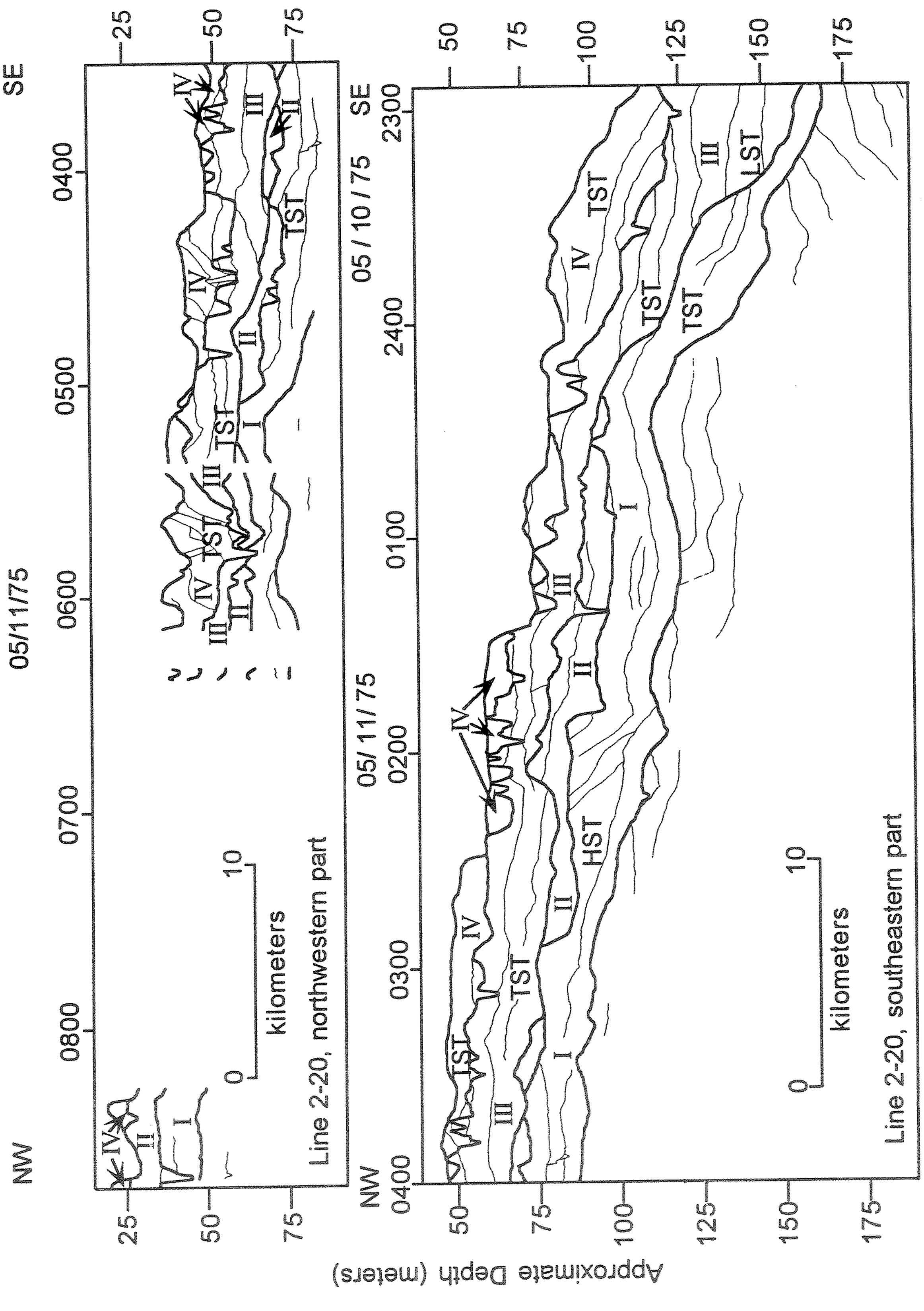


Atlantis II, Line 2-18

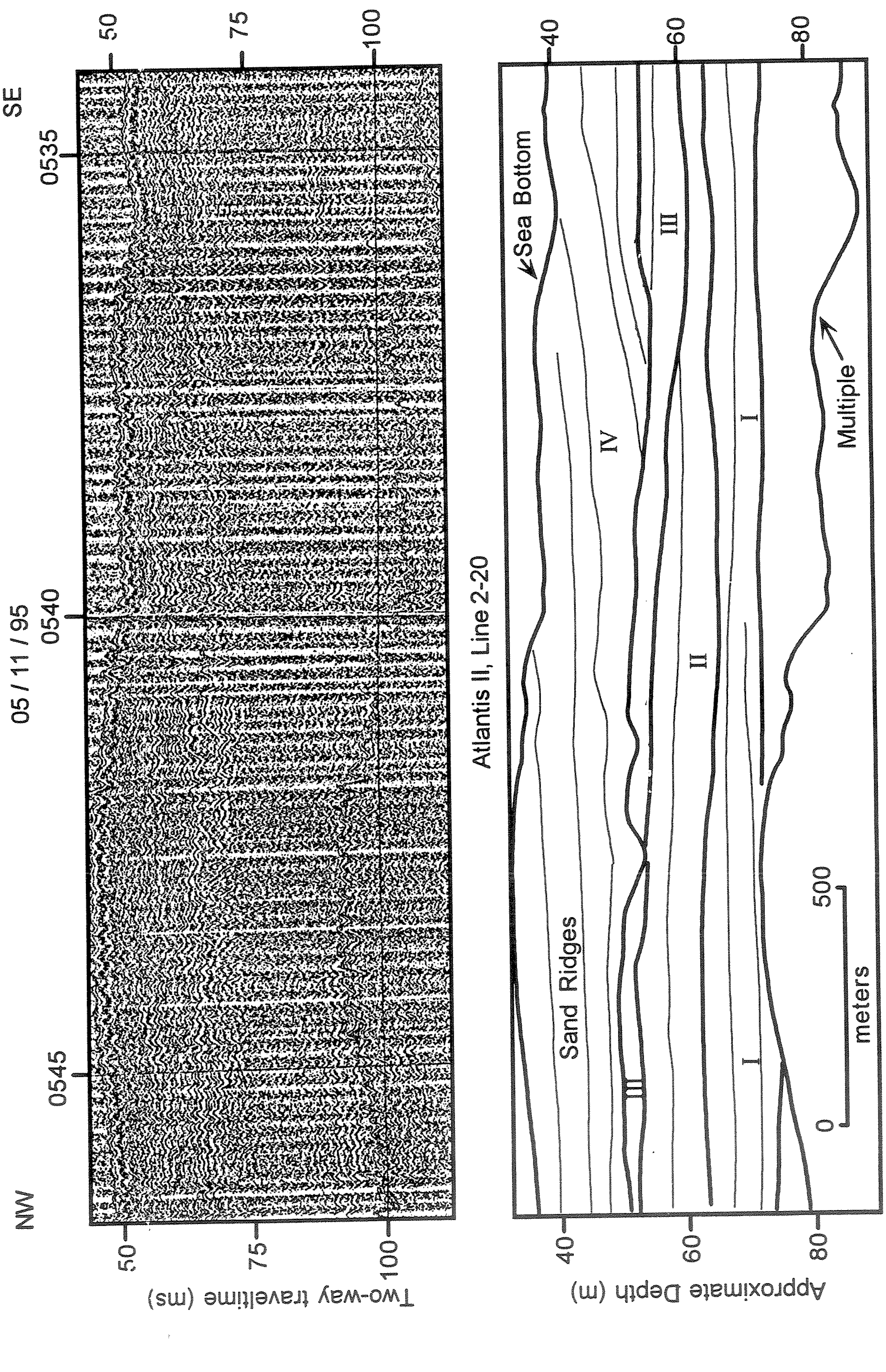

NW
05 / 10 / 95
SE
1405
1410
1415
Two-way traveltime (ms)
125
150
175
200
225
Atlantis II, Line 2-18
Approximate Depth (m)
80
100
120
140
160
180
Sea Bottom
IV
III
II
I
0
500
meters
Multiple



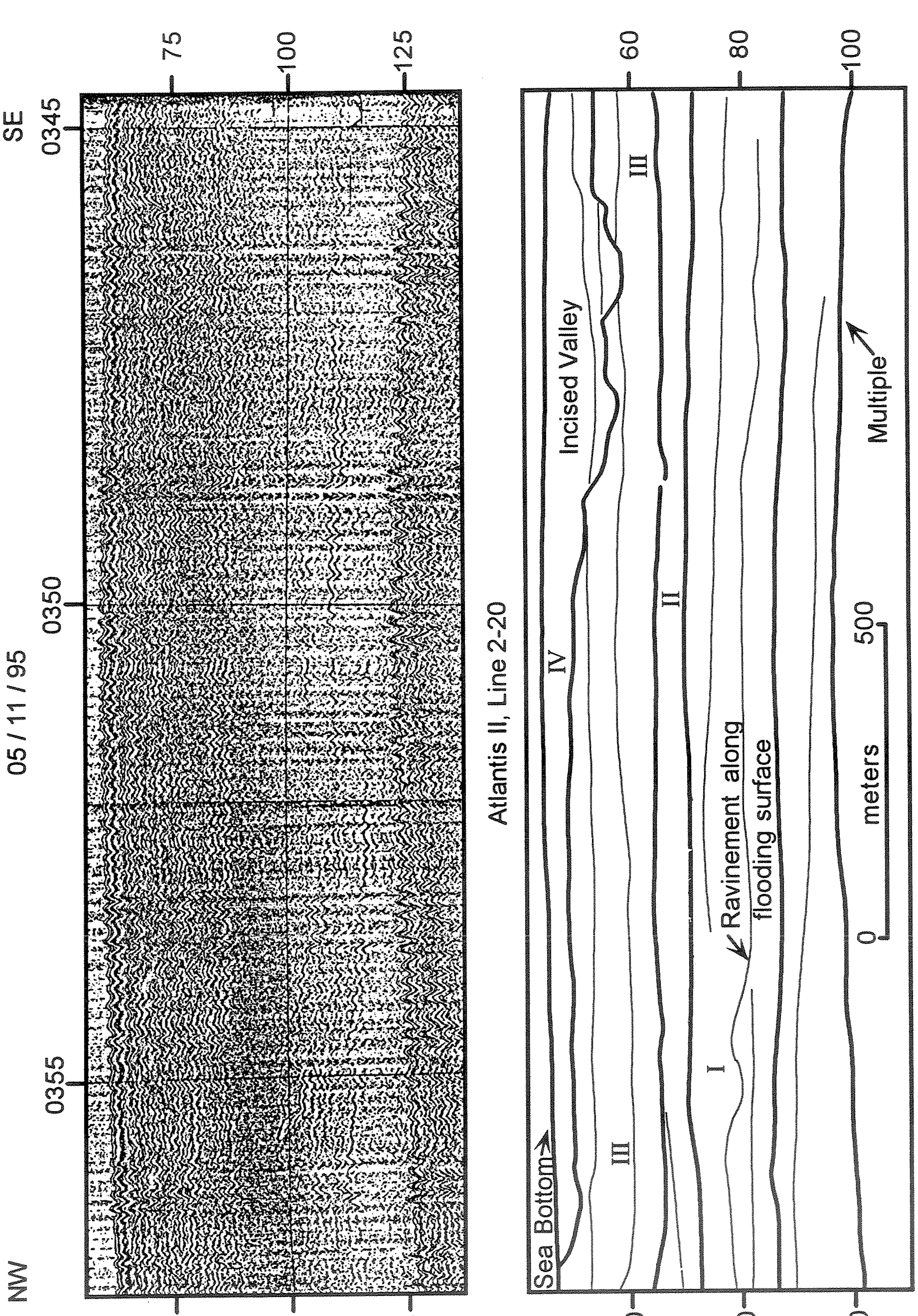
Two-way traveltime (milliseconds)
NW
05/11/75
SE
0800
0700
0600
0500
0400
25
50
75
100
Line 2-20, northwestern part
NW
05/11/75
05 / 10 / 75
SE
0400
0300
0200
0100
2400
2300
75
100
125
150
175
200
Line 2-20, southeastern part

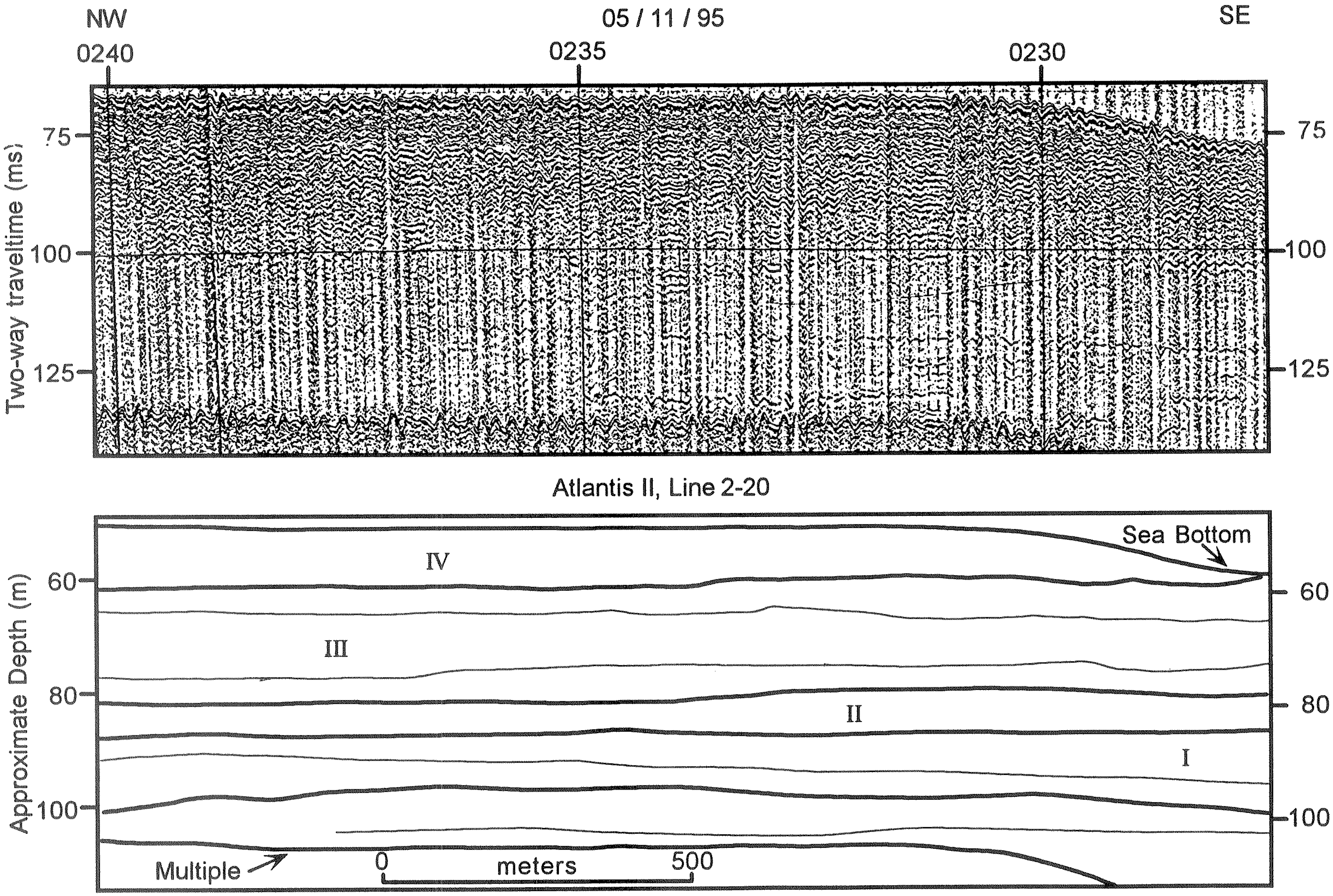


NW
SE
05/11/75
0400
0500
0600
0700
0800
kilometers
0
10
Line 2-20, northwestern part
Approximate Depth (meters)
25
50
75
IV
III
II
I
TST
05 / 10 / 75
05/ 11/ 75
2300
2400
0100
0200
0300
0400
Line 2-20, southeastern part
100
125
150
175
HST
LST

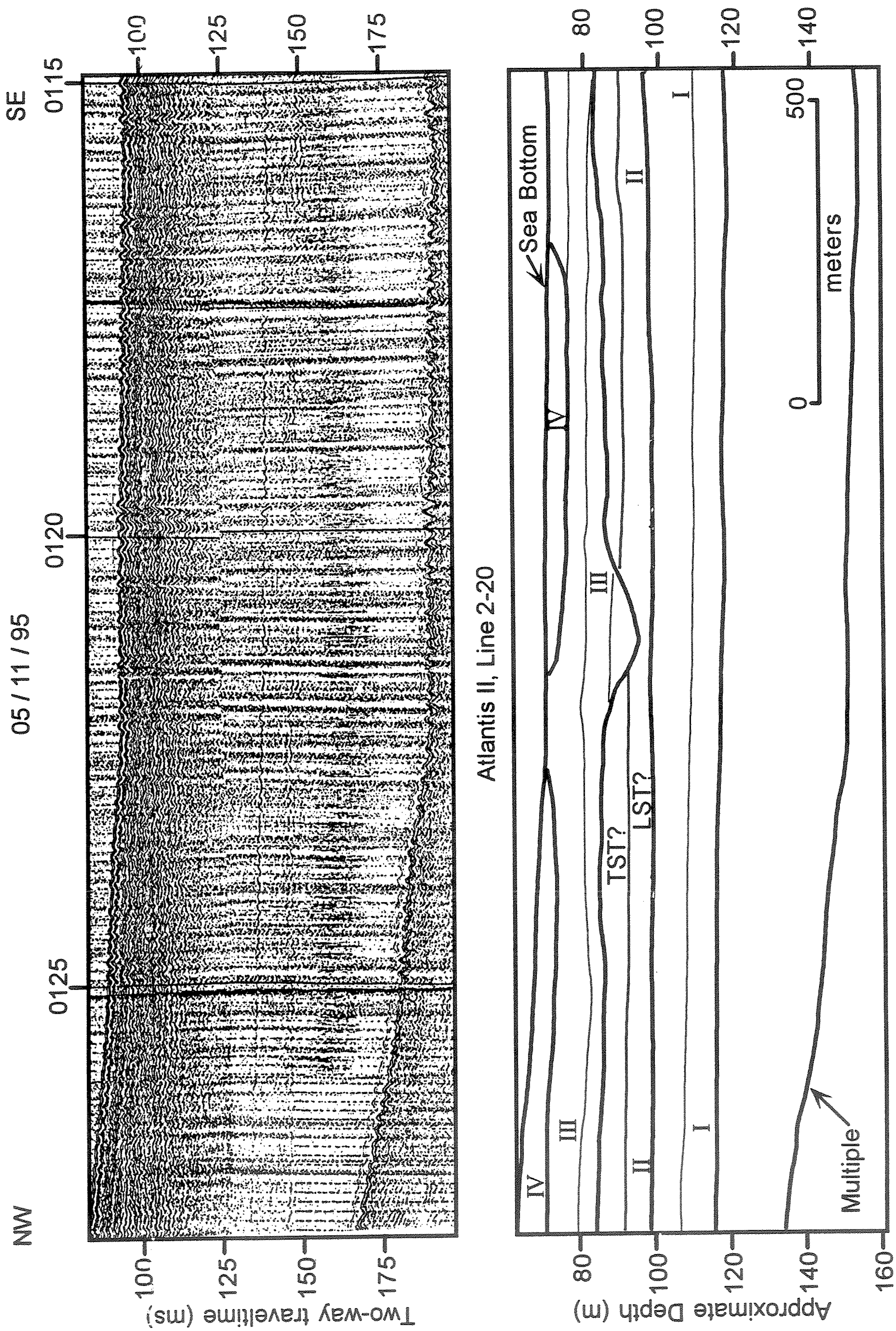


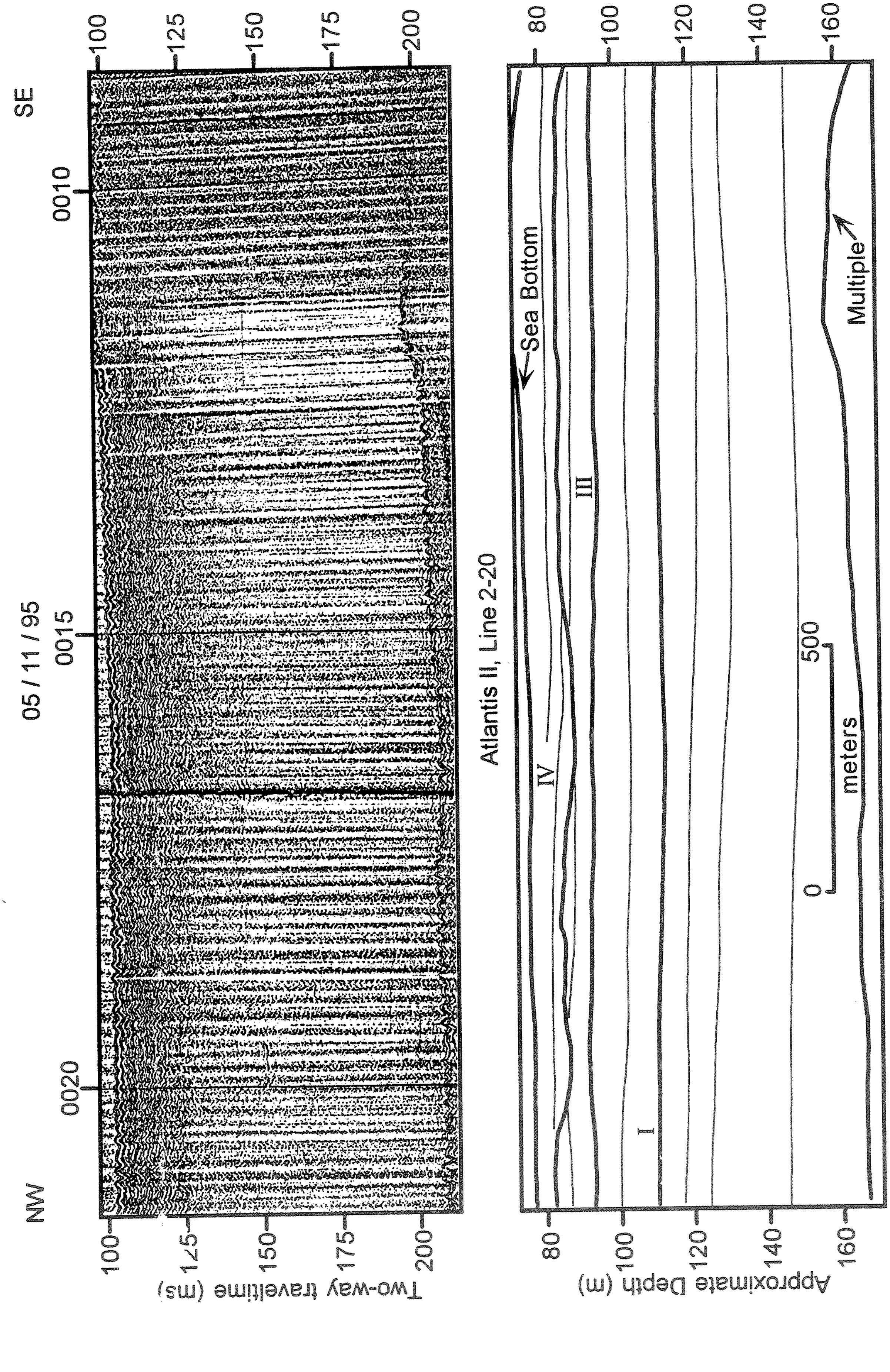
NW
SE
0545
05 / 11 / 95
0540
0535
50
75
100
Two-way traveltime (ms)
Atlantis II, Line 2-20
Sea Bottom
Sand Ridges
IV
III
II
I
Multiple
0
500
meters
40
60
80
Approximate Depth (m)


NW
SE
05 / 11 / 95
0345
0350
0355
75
100
125
Two-way traveltime (ms)
Atlantis II, Line 2-20
Sea Bottom
Incised Valley
IV
III
II
I
Ravinement along
flooding surface
0
500
meters
Multiple
60
80
100
Approximate Depth (m)



NW
05 / 11 / 95
SE
0240
0235
0230
Two-way traveltime (ms)
75
100
125
Atlantis II, Line 2-20
Sea Bottom
IV
III
II
I
Approximate Depth (m)
60
80
100
Multiple
0
meters
500


NW
SE
05 / 11 / 95
0115
0120
0125
Two-way traveltime (ms)
100
125
150
175
Atlantis II, Line 2-20
Sea Bottom
Multiple
TST?
LST?
IV
III
II
I
0
500
meters
Approximate Depth (m)
80
100
120
140
160


NW
SE
0020
05 / 11 / 95
0015
0010
Two-way traveltime (ms)
100
125
150
175
200
Atlantis II, Line 2-20
Approximate Depth (m)
80
100
120
140
160
Sea Bottom
Multiple
I
III
IV
0
500
meters

REFERENCES CITED

Aharon, P., and Chappell, J., 1986, Oxygen isotopes, sea-level changes and the temperature history of a coral reef environment in New Guinea over the past 10^5 years: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 56, p. 337-379.

Allen-Lafayette, Z., 1996, Mud dump 1996 seismic interpretation: New Jersey Geological Survey, unpublished map.

Ashley, G.M., and Sheridan, R.E., 1994, Depositional model for valley fills on a passive continental margin *in* Incised Valley Systems: Origin and Sedimentary Sequences: Tulsa, OK, SEPM Special Publication 51, p. 285-301.

Ashley, G.M., Wellner, R.W., Esker, D., and Sheridan, R.E., 1991, Clastic sequences developed during late Quaternary glacio-eustatic sea-level fluctuations on a passive continental margin: Example from the inner continental shelf near Barnegat Inlet, New Jersey: Geological Society of America Bulletin, v. 103, p. 1607-1621.

Belknap, D.F., and Kraft, J.C., 1985, Influence of antecedent geology on stratigraphic preservation potential and evolution of Delaware's barrier systems: Marine Geology, v. 63, p. 235-262.

Berggren, W.A., Burckle, L.H., Cita, M.B., Cooke, H.B.S., Funnell, B.M., Gartner, S., Hays, J.D., Kennett, J.P., Opdyke, N.D., Pastouret, L., Shackleton, N.J., and Takayanagi, Y., 1980, Towards a Quaternary time scale: Quaternary Research, v. 13, p. 277-302.

Blackwelder, B.W., 1980, Late Wisconsinan and Holocene tectonic stability of the United States Mid-Atlantic coastal region: Geology, v. 8, p. 534-537.

Blackwelder, B.W., Pilkey, O.H., and Howard, J.D., 1979, Late Wisconsinan sea level on in-place shoreline indicators: Science, v. 204, p. 618-620.

Bloom, A.L., 1983, Sea level and coastal changes *in* Wright, H.E., Jr., ed., Late-Quaternary Environments of the United States: Minneapolis, MN, University of Minnesota Press, p. 42-51.

Bloom, A.L., Broecker, W.S., Chappell, J.M.A., Matthews, R.K., and Mesolella, K.J., 1974, Quaternary sea-level fluctuations on a tectonic coast: new $^{230}Th/^{234}U$ dates from the Huon Peninsula, New Guinea: Quaternary Research, v. 4, p. 183-205.

Boulton, G.S., Smith, G.D., Jones, A.S., and Newsome, J., 1985, Glacial geology and glaciology of the last mid-continent ice sheets: Journal of the Geological Society of London, v. 142, p. 447-474.

Broecker, W.S., and Thurber, D.L., 1965, Uranium series dating of corals and oolites from Bahamian and Florida Keys limestones: Science, v. 149, p.55-57.

Calkin, P.E., Muller, E.H., and Barnes, J.J., 1982, The Gowanda Hospital interstadial site, New York: American Journal of Science, v. 282, p. 1110-1142.

Carey, J.S., Ashley, G.M., and Sheridan, R.E., 1996, Late Quaternary sequence stratigraphy of the New Jersey continental shelf: Inferred influences of continental glaciation: Buffalo, 1996 Abstracts with Programs, Geological Society of America, Northeast Section, p. 44.

Carey, J.S., Sheridan, R.E., and Ashley, G.M., 1995a, Late Pleistocene sequence stratigraphy of the New Jersey continental shelf in Official Program, 1995 Annual Convention of the American Association of Petroleum Geologists: Houston, TX, p. 15A

Carey, J.S., Ashley, G.M., and Sheridan, R.E. 1995b, Sea-level elevations, shoreline positions and Hudson valley migration during stage 3: 1995 Spring Meeting: American Geophysical Union, Eos, p. S170.

Chappell, J., 1974, Geology of coral terraces, Huon Peninsula, New Guinea: a study of Quaternary tectonic movements and sea-level changes: Geological Society of America Bulletin, v. 85, p. 553-570.

Chappell, J., 1983, Evidence for smoothly falling sea level relative to north Queensland, Australia, during the past 6000 years: Nature, v. 302, p. 406-408.

Chappell, J., and Shackleton, N.J., 1986, Oxygen isotopes and sea level: Nature, v. 324, p. 137-140.

Christensen, B.A., Hoppie, B.W., Thunnell, R.C., Miller, K.G., and Burckle, L., in press, Pleistocene age models: Proceedings of the Ocean Drilling Project, Scientific Results, v. 150.

Church, M., 1972, Baffin Island sandurs; a strudy of Arctic fluvial processes: Geological Survey of Canada Bulletin 216, 208 p.

Church, M., and Ryder, J.M., 1972, Paraglacial sedimentation: A consideration of fluvial processes conditioned by glaciation: Geological Society of America Bulletin, v. 83, p. 3059-3072.

Clark, J.A., Farrell, W.E., and Peltier, W.R., 1978, Global changes in post-glacial sea level: a numerical calculation: Quaternary Research, v. 9, p. 265-287.

Clark, J.A., and Lingle, C.S., 1979, Predicted relative sea-level changes (18,000 years B.P. to present) caused by late glacial retreat of the antarctic ice sheet: Quaternary Research, v. 9, p. 265-287.

Clarke, T.L., Swift, D.J.P., and Young, R.A., 1983, A stochastic modeling approach to the fine sediment budget of the New York Bight: Journal of Geophysical Research, v. 88, p. 9653-9660.

Clet, M., and Occhietti, S., 1994, Palynologie et paléoenvironnements des épisodes du Sable de Lotbinière et des Varves de Deschaillons (Pléistocène supérieur) de la vallée du Saint-Laurent: Canadian Journal of Earth Sciences, v. 31, p. 1474-1485.

Colman, S.M., Mixon, R.B., Bloom, A.L., and Johnson, G.H., 1989, Comment on Late Pleistocene barrier-island sequence along the southern Delmarva Peninsula: implications for middle Wisconsin sea levels: Geology, v. 17, p. 84-85.

Crowl and Sevon, 1980, Glacial border deposits of late Wisconsinan age in northeast Pennsylvania: Pennsylvania Bureau of Topographic and Geologic Survey, General Geology Report 71.

Curray, J.A., 1965, Late Quaternary history, continental shelves of the United States *in* Wright, H.E. and Frey, D.G., eds., The Quaternary of the United States: Princeton, NJ, Princeton University Press, p. 723-735.

Daddario, J., 1961, A lagoon deposit profile near Atlantic City, New Jersey: Bulletin of the New Jersey Academy of Science, v. 6, p. 7-14.

Dalrymple, R.W., Zaitlin, B.A., and Boyd, R., 1992, Estuarine facies models: conceptual basis and stratigraphic implications: Journal of Sedimentary Petrology, v. 62, p. 1130-1146.

Davies, T.A., Austin, J.A., Lagoe, M.B., and Milliman, J.D., 1992, Late Quaternary sedimentation off New Jersey: New results using 3-D seismic profiles and cores: Marine Geology, v. 108, p. 323-343.

Davis, J.L, and Mitrovica, J.X., 1996, Glacial isostatic adjustment and the anomalous tide gauge record of eastern North America: Nature, v. 379, p. 331-333.

Dillon, W.P., and Oldale, R.N., 1978, Late Quaternary sea-level curve: Reinterpretation based on glacio-tectonic influence: Geology, v. 6, p. 56-60.

Dodge, R.E., Fairbanks, R.G., Benninger, L.K., and Maurrasse, F., 1983, An early Wisconsin reef terrace at Barbados, West Indies and its climatic implications: Science, v. 219, p. 1422-1425.

Dreimanis, A., 1992, Early Wisconsinan in the north-central part of the Lake Erie basin: a new interpretation *in* Clark, P.U., and Lea, P.D., eds., The last interglacial-glacial transition in North America: Geological Society of America, Special Paper 270. p. 109-118.

Duane, D.B., Field, M.E., Miesberger, E.P., Swift, D.J.P., and Williams, S.J., 1972, Linear sand shoals on the Atlantic continental shelf, Florida to Long Island *in* Swift, D.J.P., Field, M.E., and Duane, D.B., eds., Shelf sediment transport: process and pattern: Stroudsburg, PA, Dowden, Hutchinson and Ross, p. 447-498.

Dyke, A.S., and Prest, V.K., 1986, Paleogeography of Northern North America 18000 - 12000 years ago: Ottawa, Geological Survey of Canada Map 1703A, 1:12,500,000, Sheet 1 of 3.

Edwards, R.L., Beck, J.W., Burr, G.S., Donahue, D.J., Chappell, M.A., Bloom, A.L, Daruffel, E.R.M., and Taylor, F.W., 1993, A large drop in atmospheric $^{14}C/^{12}C$ and reduced melting in the Younger Dryas, documented with ^{230}Th ages of corals: Science, v. 260, p. 962-967.

Ellsworth, J.M., 1986, Sources and sinks for fine-grained sediment in the lower Hudson River: Northeastern Geology, v. 8, p. 141-155.

Eyles, C.H., and Eyles, N., 1983, Sedimentation in a large lake: a reinterpretation of the late Pleistocene stratigraphy at Scarborough Bluffs, Ontario, Canada: Geology, v. 11, p. 146-152.

Fairbanks, R.G., 1989, A 17,000-year glacio-eustatic sea-level record: influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation: Nature, v. 342, p. 637-640.

Fairbridge, R.W., 1961, Eustatic changes in sea level *in* Ahrens, L.H., Press, F., Rankema, K., and Runkorn, S.K., eds., Physics and Chemistry of the Earth, v. 4: Oxford, Pergamon, p. 99-185.

Farran, M., and Maldonado, A., 1990, The Ebro continental shelf: Quaternary seismic stratigraphy and growth patterns: Marine Geology, v. 95, p. 289-312.

Finkelstein, K. and Kearney, M.S., 1988, Late Pleistocene barrier-island sequence along the southern Delmarva peninsula: Implications for middle Wisconsin sea levels: Geology, v. 16, p. 41-45.

Flint, R.F., 1971, Glacial and Quaternary Geology: New York, Wiley.

Frakes, L.A., 1979, Climates Through Geologic Time: Amsterdam, Elsevier, 310 p.

Frazier, D.E., 1974, Depositional episodes: their relationship to the Quaternary stratigraphic framework in the northwestern portion of the Gulf basin: University of Texas at Austin, Bureau of Economic Geology Circular 74-1, 28p.

Fullerton, D.S., 1986, Stratigraphy and correlation of glacial deposits from Indiana to New York and New Jersey: Quaternary Science Reviews, v. 5, p. 23-36.

Fulton, R.J., 1990, Foreword *in* Fulton, R.J., ed., Quaternary geology of Canada and Greenland: Boulder, Colorado, Geological Society of America, Geology of North America, v. K-1, p. 1-11.

Galloway, W.E., 1989, Genetic stratigraphic sequences in basin analysis I: Architecture and genesis of flooding-surface bounded depositional units: AAPG Bulletin, v. 73, p. 125-142.

Gallup, C.D., Edwards, R.L., and Johnson, R.G., 1994, The timing of high sea levels over the past 200,000 years: Science, v. 263, p. 796-800.

Greenlee, S.M., Schroeder, F.W., and Vail, P.R., 1988, Seismic stratigraphic and geohistory analysis of Tertiary strata from the continental shelf off New Jersey; calculation of eustatic fluctuations from stratigraphic data *in* Sheridan, R.E. and Grow, J., eds., The Atlantic Continental Margin: Boulder, Colorado, Geological Society of America, Geology of North America V. I-2,, p. 437-444.

Groot, J.J., Benson, R.N., and Wehmiller, J.F., 1995, Palynological, foramaniferal and aminostratigraphic studies of Quaternary sediments from the U.S. middle Atlantic upper continental slope, continental shelf and coastal plain: Quaternary Science Reviews, v. 14, p. 17-49.

Grow, J.A., 1980, Deep structure and evolution of the Baltimore Canyon Trough in the vicinity of the COST No. B3 well *in* Scholle, P.A., ed. Geologic studies of the COST No. B-3 well, United States Mid-Atlantic Continental Slope area: U.S. Geological Survey Circular 833, p. 117-125.

Grow, J.A., Klitgord, K.D., and Schlee, J.S., 1988, Structure and evolution of the Baltimore Canyon Trough *in* Sheridan, R.E. and Grow, J., eds., The Atlantic Continental Margin: Boulder, Colorado, Geological Society of America, Geology of North America V. I-2,, p. 269-290.

Hack, J.T., 1982, Physiographic divisions and differential uplift in the Piedmont and Blue Ridge: Reston, VA, United States Geological Survey, Professional Paper 1265, 49 pp.

Haq, B.U., Hardenbol, J., and Vail, P.R., 1987, The chronology of fluctuating sea level since the Triassic: Science, v. 235, p. 1156-1167.

Harris, J.D., 1983, Pleistocene events from the shallow structure of the middle and outer continental shelf of New Jersey [Masters Thesis]: Providence, University of Rhode Island, 112p.

Hathaway, J.C., Schlee, J.S., Poag, C.W., Valentine, P.C., Weed, E.G.A., Bothner, M.H., Kohout, F.A., Manheim, F.T., Schoen, R., Miller, R.E., and Schultz, D.M., 1976, Preliminary Summary of the 1976 Atlantic Margin Coring Project: United States Geological Survey Open File Report No. 76-844, 217p.

Hays, J.D., Imbrie, J., and Shackleton, N.J., 1975, Variations in earth's orbit: Pacemaker of the ice ages: Science, v. 194, p. 1121-1132.

Hicock, S.R., and Dreimanis, A., 1989, Sunnybrook drift indicates a grounded early Wisconsin glacier in the Lake Ontario basin: Geology, v. 17, p. 169-172.

Imbrie, J., 1985, A theoretical framework for the Pleistocene ice ages: Journal of the Geological Society of London, v. 142, p. 417-432.

Imbrie, J., Shackleton, N.J., Pisias, N.G., Morley, J.J., Prell, W.L., Martinson, D.G., Hays, J.D., MacIntyre, A.D., Mix, A.C., 1984, The orbital theory of Pleistocene climate: support from a revised chronology of the marine δO^{18} record *in* Berger, A., ed., Milankovitch and Climate, Part I, Hingham, Massachusetts, Reidel, p. 269-305.

James, N.P., Mountjoy, E.W., and Omura, A., 1971, An early Wisconsin reef terrace at Barbados, West Indies, and its climatic implications: Geological Society of America Bulletin, v. 82,p. 2011-2018.

Jervey, M.T., 1988, Quantitative geologic modeling of siliciclastic rock sequences and their seismic expression *in* Wilgus, C.K., Hastings, B.S., Kendall, C.G.St.C., Posamentier, H.W., Ross, C.A., and Van Wagoner, J.C., eds., Sea-level Changes; An Integrated Approach: SEPM Special Publication No. 42, p. 47-70.

Karrow, P.F., 1967, Pleistocene geology of the Scarborough area: Toronto, Ontario: Ontario Department of Mines, Geological Report 46, 108 p.

Karrow, P.F., and Occhietti, S., 1990, Quaternary geology of the St. Lawrence Lowlands of Canada *in* Fulton, R.J., ed., Quaternary Geology of Canada and Greenland: Boulder, Colorado, Geology of North America v. K-1, p. 321-389.

Knebel, H.J., and Spiker, E.C., 1977, Thickness and age of surficial sand sheet, Baltimore Canyon Trough area: American Association of Petroleum Geologists Bulletin, v. 61, p. 861-870.

Knebel, H.J., Wood, S.A., and Spiker, E.C., 1979, Hudson River; evidence for extensive migration on the exposed continental shelf during Pleistocene time: Geology, v. 17, p. 254-258.

Koteff, C. and Pessl, F., Jr., 1985, Till stratigraphy in New Hampshire: Correlations with adjacent New England and Quebec: Boulder, Colorado, Geological Society of America, Special Paper 197, 12 p.

Kraft, J.C, Chrzastowski, M.J., Belknap, D.F., Toscano, M.A., and Fletcher, C.H., III, 1987, The transgressive barrier lagoon coast of Delaware: morphostratigraphy, sedimentary sequences and responses to relative rise in sea level *in* Sea-level Fluctuation and Coastal Evolution: Society of Economic Paleontologists and Mineralogists, Special Publication 41, p. 129-143.

Ku, T., Kimmel, M.A., Easton, W.H., and O'Neill, T.J., 1974, Eustatic sea-level 120,000 years ago on Oahu, Hawaii: Science, v. 183, p. 959-962.

Kutzbach, J.E., 1987, Model simulations of the climatic patterns during the deglaciation of North America *in* North America and Adjacent Oceans During the Last Deglaciation: Boulder, Colorado, Geology of North America, v. K-3, p. 447-462.

Lamothe, M., 1989, A new framework for the Pleistocene stratigraphy of the central St. Lawrence Lowland, southern Quebec: Géographie Physique et Quaternaire, v. 43, p. 119-129.

Linsley, B., 1996, Oxygen-isotope record of sea level and climate variations in the Sulu Sea over the past 150,000 years: Nature, v. 380, p. 234-237.

Ludwig, K.R., Muhs, D.R., Simmons, K.R., Halley, R.B., Halley, R.B., and Shinn, E.A., 1996, Sea-level records at ~80 ka from tectonically stable platforms, Florida and Bermuda: Geology, v. 24, p. 211-214.

Lyles, S.D., Hickman, L.E., Jr., and Debaugh, H.A., Jr., 1987, Sea level variations for the United States 1855-1986: Rockville, MD, NOAA National Ocean Service, 182 pp.

MacIntyre, I.G., Pilkey, O.H. and Stuckenrath, R., 1978, Relict oysters on the United states Atlantic continental shelf: a reconsideration of their usefulness in understanding late Quaternary sea-level history: Geological Society of America Bulletin, v. 89, p. 277-282.

Mathews, W.H., 1975, Cenozoic erosion and erosion surfaces of eastern North America: American Journal of Science, v. 275, p. 818-824.

McClennen, C.E., 1973, Nature and Origin of the New Jersey Continental Shelf Topographic Ridges and Depressions [Ph.D. dissert.]: Providence, University of Rhode Island, 93 p.

Miall, A.D., 1986, Eustatic sea level changes interpreted from seismic stratigraphy: a critique of the methodology with particular reference to the North Sea Jurassic record: American Association of Petroleum Geologists Bulletin, v. 70, p. 131-137.

Milliman, J.D., and Emery, K.O., 1968, Sea levels during the past 35,000 years: Science, v.162, 1121-1123.

Milliman, J.D., Jiezao, Z., Anchun, Li, and Ewing, J.I, 1990, Late Quaternary sedimentation on the outer and middle New Jersey continental shelf: Result of two local deglaciations?: Journal of Geology, v. 98, p. 966-976.

Mitchum, R.M., Jr., and Vail, P.R., 1977, Seismic stratigraphy and global changes of sea level, Part 7: Seismic stratigraphic interpretation procedure *in* Payton, C.E., ed., Seismic Stratigraphy--Applications to Hydrocarbon Exploration: Tulsa, OK, American Association of Petroleum Geologists, Memoir 96, p. 135-143.

Mitchum, R.M., Jr., Vail, P.R., and Thompson, S., 1977a, Seismic stratigraphy and global changes of sea level, Part 2: The depositional sequence as a basic unit for stratigraphic analysis *in* Payton, C.E., (ed.), Seismic Stratigraphy--Applications to Hydrocarbon Exploration: Tulsa, OK, American Association of Petroleum Geologists, Memoir 96, p. 53-62.

Mitchum, R.M., Jr., Vail, P.R., and Sangree, J.B., 1977b, Seismic stratigraphy and global changes of sea level, Part 6: Stratigraphic interpretation of seismic reflection patterns in depositional sequences *in* Payton, C.E., ed., Seismic Stratigraphy--Applications to Hydrocarbon Exploration: Tulsa, OK, American Association of Petroleum Geologists, Memoir 96, p. 117-133.

Mitterer, R.M., 1975, Ages and diagenetic temperatures of Pleistocene deposits of Florida based upon isoleucine epimerization in *Mercenaria*: Earth and Planetary Science Letters, v. 28, p. 275-282.

Mixon, R.B., Hazel, J.E., Rubin, M., Sirkin, L.A., and Szabo, B.J., 1974, Geologic framework of the southernmost Delmarva peninsula, Accomack and Northampton counties, Virginia: Geological Society of America, Abstracts with Program (Northeastern Section), v. 6, p. 56-57.

Mountain, G.S., Miller, K.G., and 26 others, 1994, Proceedings of the Ocean Drilling Program, Initial Reports, v. 150 : College Station, TX: Ocean Drilling Program, 872 pp.

Muhs, D.R., Kennedy, G.L., and Rockwell, T.K., 1994, Uranium-series ages of marine terrace corals from the Pacific coast of North America: implications for last-interglacial sea-level history: Quaternary Research, v. 42, p. 72-87.

Muller, E.H., and Calkin, P.E., 1993, Timing of Pleistocene glacial events in New York state: Canadian Journal of Earth Sciences, v. 30, p. 1829-1845.

Murray-Wallace, C.V., Belperio, A.P., Gostin, V.A., and Cann, J.H., 1992, Amino acid racemization and radiocarbon dating of interstadial marine strata (oxygen isotope stage 3) Gulf of St. Vincent, South Australia: Marine Geology, v. 110, p. 83-92.

Mylroie, J.E., and Carew, J.L., 1988, Solution conduits as indicators of late Quaternary sea level position: Quaternary Science Reviews, v. 7, p. 55-64.

Nakiboglu, SM., and Lambeck, K., 1990, Secular sea-level change *in* Sabadini, R., Lambeck, K., and Boschi, E., eds., Glacial Isostasy, Sea-Level and Mantle Rheology: NATO ASI Series C: Mathematical and Physical Sciences, v. 334, p. 237-258.

Oertel, G.F., Kearney, M.S., Leatherman, S.P., and Woo, H.J., 1989, Anatomy of a barrier platform, outer barrier lagoon, Southern Delmarva Peninsula, Virginia: Marine Geology, v. 88, p. 303-318.

Oertel, G.F., Henry, V.J., and Foyle, A.M., 1991, Implications of tide-dominated lagoonal processes on the preservation of buried channels on a sediment-starved continental shelf *in* Swift, D.J.P., Oertel, G.F., Tillman, R.W., and Thorne, J.A., eds., Shelf sand and sandstone bodies--geometry, facies and sequence stratigraphy: International Association of Sedimentologists, Special Publication 14, p. 379-394.

Oldale, R.N., and Colman, S.M., 1992, On the age of the penultimate full glaciation of New England *in* Clark, P.W., and Lea, P.D., eds., The last interglacial-glacial transition in North America: Geological Society of America, Special Paper 370, p. 163-170.

Pekar, S.F., and Miller, K.G., (in press), New Jersey coastal plain Oligocene sequences, ODP Leg 150X *in* Miller, K.G., S.W. Snyder, et al., Proceedings of the Ocean Drilling Program, Scientific Results, v. 150X

Peltier, W.R., 1986, Deglaciation-induced vertical motion of the North American continent and transient lower mantle rheology: Journal of Geophysical Research, v. 91, p. 9099-9123.

Peltier, W.R., 1994, Ice age paleotopography: Science, v. 265, p. 195-201.

Poag, C.W., 1979, Stratigraphy and depositional environments of the Baltimore Canyon Trough: American Association of Petroleum Geologists Bulletin, v. 63, p. 1452-1466.

Poag, C.W., and Sevon, W.D., 1989, A record of Appalachian denudation in postrift Mesozoic and Cenozoic sedimentary deposits of the U.S. middle Atlantic continental margin: Geomorphology, v. 2, p. 119-157.

Poppe, L.J., 1981, Data File-the 1976 Atlantic Margin Coring (AMCOR) Project of the U.S. Geological Survey: U.S. Geological Survey Open-File Report 81-239: U.S. Department of the Interior, Reston, Va., 96p.

Posamentier, H.W., and Vail, P.R., 1988, Eustatic controls on clastic deposition II; Sequence and systems tract models *in* Kilgus, C.W., Hastings, B.S., Kendall, C.G.St.C., Posamentier, H.W., Ross, C.A., and Van Wagoner, J.C., eds., Sea Level Changes: An Integrated Approach: Tulsa, OK, SEPM Special Publication 42, p. 125-154.

Psuty, N.P., 1985, Holocene sea level in New Jersey: Physical Geography, v. 7, p. 156-167.

Raymo, M.E., Ruddiman, W.F., Shackleton, N.J., and Oppo, D.W., 1990, Evolution of global ice volume and Atlantic-Pacific $\delta^{13}C$ gradients over the past 2.5 M.Y.: Earth and Planetary Science Letters, v. 97, p. 353-368.

Richards, D.A., Smart, P.L., and Edwards, R.L., 1994, Maximum sea levels for the last glacial period from U-series ages of submerged speleothems: Nature, v. 367, p. 357-360.

Richmond, G.M., and Fullerton, D.S., 1986, Introduction to Quaternary glaciations in the United States of America *in* Sibrava, V., Bowen, D.Q., and Richmond, G.M., eds., Quaternary Glaciations in the Northern Hemisphere: Oxford, Pergamon Press, p. 3-10.

Ridge, J.C., Duane, D.D., and Evenson, E.B., 1990, Does Altonian drift exist in Pennsylvania and New Jersey: Quaternary Research, v. 33, p. 253-258.

Rine, J.M., Tillman, R.W., Culver, S.J., and Swift, D.J.P., 1991, Generation of Late Holocene ridges on the middle continental shelf of New Jersey, USA--evidence for formation in a mid-shelf setting based upon comparison with a nearshore ridge *in* Swift, D.J.P., Oertel, G.F., Tillman, R.W., and Thorne, J.A., eds., Shelf Sands and Sandstone Bodies: Oxford, United Kingdom, International Association of Sedimentologists, Special Publication 14, p. 395-426.

Shackleton, N., 1967, Oxygen isotope analyses and Pleistocene temperatures reassessed: Nature, v. 215, p. 15-17.

Shackleton, N.J., 1987, Oxygen isotopes, ice volume and sea level: Quaternary Science Reviews, v. 6, p. 183-190.

Sheridan, R.E., Ashley, G.M., and Carey, J.S., 1993, Seismic stratigraphic evidence of late Pleistocene sea levels controlling coastal plain evolution: Geological Society of America Abstracts With Program, v. 25?

Snedden, J.W., Tillman, R.W., Kreisa, R.D., Schweller, W.J., Culver, S.J., and Winn, R.D., Jr., 1994, Stratigraphy and genesis of a modern shoreface-attached sand ridge, Peahala Ridge, New Jersey: Journal of Sedimentary Research, v. B64, p. 560-581.

Stahl, L., Koczan, J., and Swift, D.J.P., 1974, Anatomy of a shoreface-connected sand ridge on the New Jersey shelf: implications for the development of the surficial sand sheet: Geology, v. 2, p. 117-120.

Stanford, S.D., 1993, Late Cenozoic surficial deposits and valley evolution of unglaciated northern New Jersey: Geomorphology, v. 7, p. 267-288.

Steckler, M.S., Watts, A.B., and Thorne, J.A.1988, Subsidence and basin modeling at the U.S. Atlantic passive margin *in* Sheridan, R.E., and Grow, J.A., eds., The Atlantic Continental Margin: U.S.: Geological Society of America, The Geology of North America V. I-2, p. 399-416.

Steckler, M.S., Reynolds, D.J., Coakley, B.J., Swift, B.A., and Jarrard, R., 1993, Modelling passive margin sequence stratigraphy *in* Posamentier, H.W., Summerhayes, C.P., Haq, B.U., and Allen, G.P., eds., Sequence Stratigraphy and Facies Associations: Oxford, United Kingdom, International Association of Sedimentologists, Special Publication 14, p. 19-41.

Steinen, R.P., Harrison, R.S., and Matthews, R.K., 1973, Eustatic low stand of sea level between 125,000 and 105, 000 years B.P.: evidence from the subsurface of Barbados, West Indies: Geological Society of America Bulletin, v. 84, p. 63-70.

Stubblefield, W.L., McGrail, D.W., and Kersey, D.G., 1983, Recognition of transgressive and post-transgressive sand ridges on the New Jersey continental shelf: American Association of Petroleum Geologists Bulletin, v. 67, p. 817-830.

Stuiver, M., and Daddario, J.J., 1963, Submergence of the New Jersey coast: Science, v. 142, p. 951-953.

Suter, J.R., Berryhill, H.L., and Penland, S., 1987, Late Quaternary sea-level fluctuations and depositional sequences, southwest Louisiana continental shelf *in* Nummedal, D., Pilkey, O.H., and Howard, J.D., eds., Sea-Level Fluctuation and Coastal Evolution: SEPM Special Publication No. 41, p. 199-222.

Swift, D.J.P., Kofoed, J.W., Saulsbury, F.P. and Sears, P., 1972, Holocene evolution of the shelf surface, south and central Atlantic shelf of North America *in* Swift, D.J.P., Duane, D.B., and Pilkey, O.H., eds., Shelf Sediment Transport: Process and Pattern: Stroudsburg, PA, Dowden, Hutchinson and Ross, p. 499-574.

Swift, D.J.P., Young, R. A., Clarke, T. L., Vincent, C.E., Niedoroda, A., and Lesht, B., 1981, Sediment transport in the Middle Atlantic Bight of North America: synopsis of recent observations *in* Nio, S.D., Shuttenhelm, R.T.E., and Van Weering, Tj.C.E., eds., Holocene Marine Sedimentation in the North Sea Basin: Oxford, United Kingdom, International Association of Sedimentologists, Special Publication 5, p. 361-383.

Sydow, J., and Roberts, H.H., 1994, Stratigraphic framework of a late Pleistocene shelf-edge delta, northeast Gulf of Mexico: American Association of Petroleum Geologists Bulletin, v. 78, p. 1276-1312.

Sykes, G., 1991, Amino Acid Dating, *in* P.L. Smart and P.D. Frances, Quaternary Dating Methods--A User's Guide: Quaternary Research Organization Technical Guide No.4, Cambridge, p. 161-176.

Teller, J.T., 1987, Proglacial lakes and the southern margin of the Laurentide Ice Sheet *in* North America and Adjacent Oceans During the Last Deglaciation: Boulder, Colorado, Geology of North America, v. K-3, p. 39-70.

Tesson, M., Gensous, B., Allen, G.P., and Ravenne, Ch., 1990, Late Quaternary deltaic lowstand wedges on the Rhône Continental Shelf, France: Marine Geology, v. 91, p. 325-332.

Thierstein, H.R., Geitzenauer, K., Molfino, B., and Shackleton, N.J., 1977, Global synchroneity of Late Quaternary coccolith datums; validation by oxygen isotopes: Geology, v. 5, p. 400-404.

Thom, B.G., 1973, Dilemma of high interstadial sea levels during the last glaciation: Progress in Geography, v. 5, p. 167-246.

Thomas, M.A., 1992, Clastic sequences developed during late Quaternary glacio-eustatic sea-level fluctuations on a passive continental margin: Example from the inner continental shelf near Barnegat Inlet, New Jersey: Discussion: Geological Society of America Bulletin, v. 104, p. 1386-1388.

Thomas, M.A., and Anderson, J.B., 1989, Glacial eustatic controls on seismic sequences and parasequences of the Trinity/Sabine incised valley, Texas continental shelf: Gulf Coast Association of Geological Societies Transactions, v. 39, p. 563-570.

Thorne, J.A., and Swift, D.J.P., 1991, Continental margin sedimentation--regime model *in* Swift, D.J.P., Oertel, G.F., Tillman, R.W., and Thorne, J.A., eds., International Association of Sedimentologists, Special Publication 14: Shelf Sand and Sandstone Bodies--Geometry, Facies and Sequence Stratigraphy: Blackwell Scientific, Oxford, p. 189-255.

Thunnell, R.C., Anderson, D., Gellar, D., and Miao, Q., 1994, Sea surface temperature-estimates for the Tropical Western Pacific during the last deglaciation and their implications for the Pacific Warm Pool: Quaternary Research, v. 41, p. 255-264.

Toscano, M.A., 1989, Comment on "Late Pleistocene barrier-island sequence along the southern Delmarva Peninsula: Implications for middle Wisconsin sea levels: Geology, v. 17, p. 85-86.

Toscano, M.A. and York, L.L., 1992, Quaternary stratigraphy and sea-level history of the U.S. Middle Atlantic Coastal Plain: Quaternary Science Reviews, v. 11, p. 301-328.

Tushingham, A.M., and Peltier, W.R., 1992, Validation of the ICE-3G model of Wurm-Wisconsin deglaciation using a global data base of relative sea level histories: Journal of Geophysical Research, v. 97, p. 3285-3304.

Vail, P.R., Hardenbol, J., and Todd, R.G., 1984, Jurassic unconformities, chronostratigraphy and sea-level changes from seismic stratigraphy and biostratigraphy *in* J.S. Schlee, ed., Interregional unconformities and hydrocarbon accumulation: American Association of Petroleum Geologists Memoir 36, p. 129-144.

Vail, P.R., Mitchum, R.M., Jr., Todd, R.G., Widmier, J.M., Thompson, S., III, Sangree, J.B., Bubb, J.N, and Hatlelid, W.G., 1977, Seismic stratigraphy and global changes of sea level *in* Payton, C.E., ed., Seismic Stratigraphy--Applications to Hydrocarbon Exploration: Tulsa, OK, American Association of Petroleum Geologists, Memoir 96, p. 49-211.

Van Wagoner, J.C., Posamentier, H.W., Mitchum, R.M., Jr., Vail, P.R., Sarg, J.F., Loutit, T.S., and Hardenbol, J., 1988, An overview of the fundamentals of sequence stratigraphy and key definitions in Wilgus, C.K., Hastings, B.S., Kendall, C.G.St.C., Posamentier, H.W., Ross, C.A., Van Wagoner, J.C., eds., Sea-level Changes; An Integrated Approach: SEPM Special Publication No. 42, p. 39-46.

Veeh, H.H., Schwebel, D., Van de Graaff, W.J.E, and Denman, P.P., 1979, Uranium-series ages of coralline terrace deposits in Western Australia: Journal of the Geological Society of Australia, v. 26, p. 285-292.

Vincent, C.E., Swift, D.J.P., and Hillard, B., 1981, Sediment transport in the New York Bight, North Atlantic Shelf: Marine Geology, v. 42, p. 369-398.

Walker, R.G., 1992, Facies, facies models and modern stratigraphic concepts *in* R.G. Walker and N.P. James (eds.), Facies Models--Response to Sea Level Change: Geological Association of Canada, St. John's, p. 1-14.

Wehmiller, J.F., 1984, Relative and absolute dating of Quaternary mollusks with amino acid racemization: evaluation, applications, and questions *in* Mahaney, W.C., ed., Quaternary Dating Methods: Elsevier, p. 171-194.

Wehmiller, J.F., and Belknap, D.F., 1982, Amino acid age estimates: Quaternary Atlantic Coastal Plain, Comparisons with U-series dates, biostratigraphy and paleomagnetic control: Quaternary Research, v. 18, p. 311-336.

Wehmiller, J.F., York, I.I., Belknap, D.F., and Snyder, S.W., 1992, Theoretical correlations and lateral discontinuities n the Quaternary aminostratigraphic record of the U.S. Atlantic Coastal Plain: Quaternary Research, v. 38, p. 275-291.

Wellner, R.W., Ashley, G.M., and Sheridan, R.E., 1993, Seismic stratigraphic evidence for a submerged middle Wisconsinan barrier: Implications for sea-level history: Geology, v. 21, p. 109-112.

Vita

John S. Carey

1985-89	B.Sc.(Hons.) in Geology & Physical Geography Queen's University, Kingston, Ontario Canada.
1989-91	Teaching Assistant, Dept. of Geological Sciences, University of Alberta, Edmonton, Alberta, Canada.
1990	R.W. Dalrymple, and Carey, J.S., Water-level fluctuations in Lake Ontario over the last 4500 years as recorded in the Cataraqui River Lagoon, Kingston, Ontario: Canadian Journal of Earth Sciences, v. 27, p. 1330-1338.
1991	Summer Exploration Geologist, Texaco Canada Petroleum Inc., Calgary, Alberta, Canada.
1992	M.Sc. in Geology, University of Alberta, Edmonton, Alberta, Canada .
1994-95	Teaching Assistant, Department of Geological Sciences.
1995	J.S. Carey, T.F. Moslow, and J.V. Barrie, Origin and distribution of Holocene temperate carbonates, Hecate Strait, western Canada continental shelf: Journal of Sedimentary Research, v. B65, p. 185-194.
1996	Research Assistant, Institute of Marine and Coastal Sciences.
1996	Ph.D. in Geology.