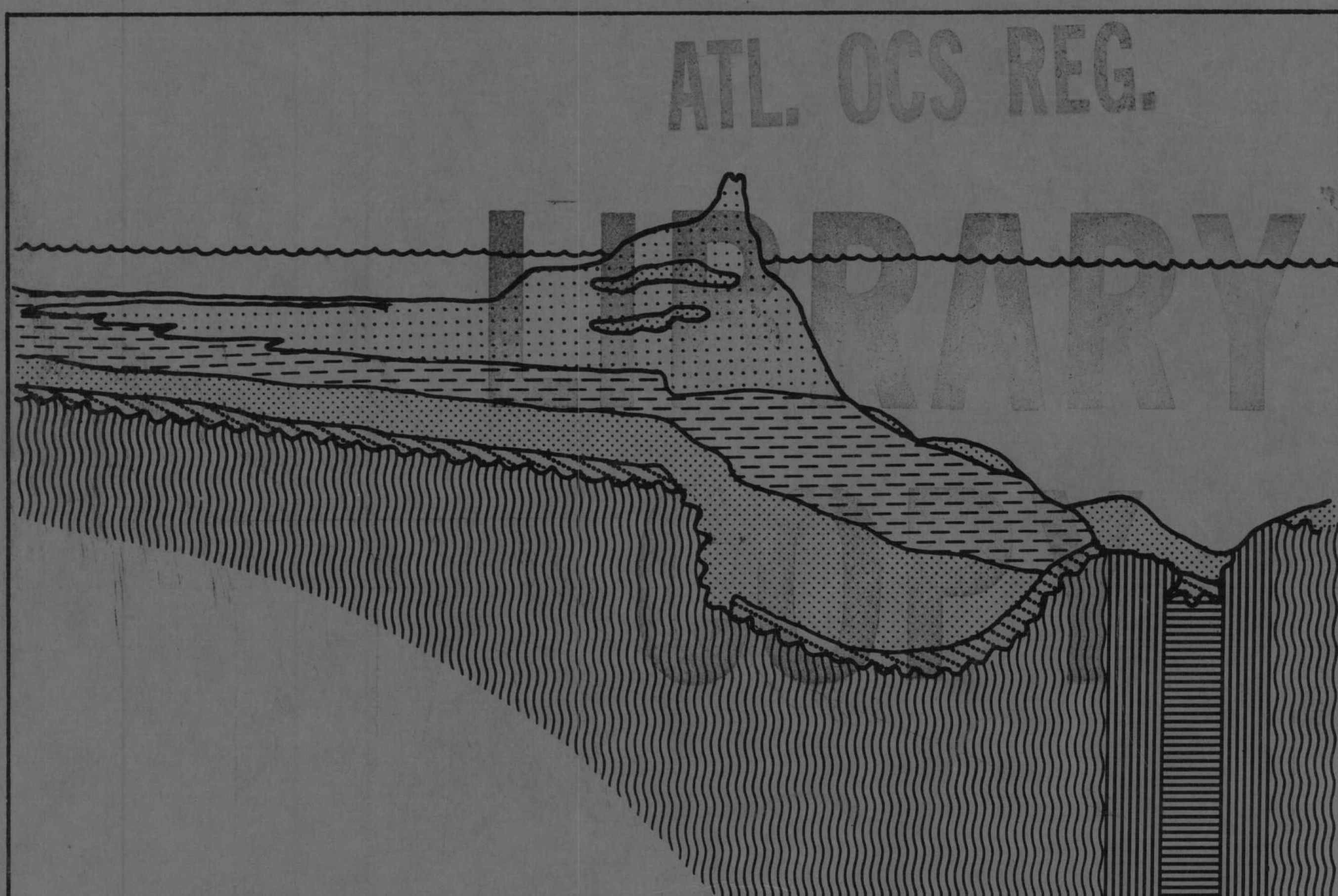
Summary and Analysis of Cultural Resource Information on the Continental Shelf from the Bay of Fundy to Cape Hatteras

FINAL REPORT

Volume I – Physical Environment



prepared by
Institute for Conservation Archaeology
Peabody Museum
Harvard University

for the Bureau of Land Management under contract number AA 551-CT8-18

A SUMMARY AND ANALYSIS OF CULTURAL RESOURCE INFORMATION ON THE CONTINENTAL SHELF FROM THE BAY OF FUNDY TO CAPE HATTERAS

FINAL REPORT

Volume I - Physical Environment

This study was funded by the New York Outer Continental Shelf Office of the Bureau of Land Management, Department of the Interior, under contract number AA 551-CT8-18.

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PREFACE

This is the first in a series of four volumes entitled "Summary and Analysis of Cultural Resource Information on the Continental Shelf from the Bay of Fundy to Cape Hatteras" which were prepared for the Bureau of Land Management (BLM) by the Institute for Conservation Archaeology (ICA) of the Peabody Museum at Harvard University. These four volumes, their accompanying chart sets, a computer-compatible tape documenting the accumulated inventories, and a set of large scale (1:125,000) maps showing the inventory and the results of our analysis constitute the final report for a project performed under contract number AA551-CT8-18 for the BLM. The purpose of this project is to provide the BLM with information about the existence of known or expected prehistoric sites and historically important sunken ships, as well as appropriate methods for locating the same, and planning recommendations for both offshore and onshore land use. The principal challenge of this project lies in the fact that the project's scope-of-work demands that individual lease blocks be classified according to whether or not they have the potential for containing cultural resources. It is for that reason that considerable emphasis is placed on the transgressional geological processes on the Continental Shelf (CS). These processes and their implications for the project are the subject of this volume.

Archaeologists and historians generally agree that given the length of time the CS was above sea level (about 15,000 years) and the intensity of European and other shipping along the northeastern coast of the USA in the period after the CS was inundated, there is probably no area on the Shelf that does not have the possibility for containing remains of either prehistoric peoples or sunken shipping. All other things being equal, this would mean that whenever federal funds were involved in land-modifying projects anywhere on the CS, federal antiquities legislation would apply to these activities (see 36 CFR 800 for a summary of the necessary procedures). On the other hand the cost of looking for and recovering data from possible properties which might be impacted in many cases exceeds the cost of exploring for the resources that are considered necessary for the economic well-being of the nation. It is possible at this point that predictions about early planning with respect to possible cultural resources on the CS will assist land users not only to meet their legal responsibilities in terms of historic preservation but to use varying levels of survey intensity to locate those sites or wrecks which may be endangered by land use.

It is important to stipulate here that, using the data presently available, nobody in the historic preservation community could, in good conscience, ever entirely eliminate any area from consideration for further work. This study attempts to give planning and management guidance to potential land users and those having jurisdiction over the use of lands on or abutting the CS from the Bay of Fundy to Cape Hatteras.

Volume I, "Physical Environment," describes the processes that influenced the preservation of subaerial surfaces (those surfaces that were exposed to the air before submergence). The differential preservation of these surfaces provides guidance to those making recommendations about the kind of survey (locational study) that may be needed to insure that various land-using projects minimize impact to important prehistoric and historic properties.

This volume then assesses the geological processes on the CS during the Holocene and Late Pleistocene that may have influenced the preservation of land surfaces containing evidence of past human activity. This preservation, as will be pointed out below, may be complete, partial, negligible, or non-existent. The expected degree of preservation will to a large extent dictate the recommended type of locational studies which will be appropriate after a program of pilot studies validates or modifies our models. The additional element of this volume is the location and identification of various physiographic regions on the CS. These data assist the archaeologists on the team to differentiate zones of prehistoric land use. These physiographic regions are therefore an influence on the type of sites that may be found in these zones (see Volume II).

Thus this volume is the primary element in predicting the integrity of archaeological sites, which then leads to management recommendations concerning cultural resources on and abutting the Continental Shelf.

Special thanks are due to our consultants, Dr. Donald J.P. Swift, Dr. Jerry R. Schubel, and Dr. Alan Niederoda; The BLM Geologists Review Committee; and David Hirschberg, Research Assistant. Thanks also to the Harvard University Geology Library Staff; Dr. Robert N. Oldale and William P. Dillon; and to the report production and office staffs of the Institute for Conservation Archaeology, Janet Johnson, Georgess McHargue, Gretchen Neve, Lynne Perrotte, Whitney Powell, Elizabeth Wahle, Mary Beth Zickefoose,without whom this volume could not have been completed.

Michael E. Roberts
Project Manager

1.0 INTRODUCTION

The Atlantic Continental Shelf of North America is one of the best studied marine regions in the world (Emery and Uchupi 1972). There exists for this area a large volume of geological literature spanning over half a century of intensive research. Despite the large volume of research, only a few individuals have addressed the geology of the Continental Shelf from an archaeological perspective (Emery and Edwards 1966; Edwards and Emery 1977; Edwards and Merrill 1977). Many studies have focused on sea-level change during the Holocene (for example Belknap and Kraft 1977; Bloom and Stuiver 1963; Kaye and Barghoorn 1964; Newman and Rusnak 1965; Redfield and Rubin 1962; Stuiver and Borns 1975 to name only a few) and during the Late Pleistocene (for example Curray 1965; Emery and Garrison 1967; Milliman and Emery 1968; and Shepard 1963). Little attention, however, has been paid to the fate of the subaerial surface during marine transgression. Exactly what happens to the subaerial land surface during erosional shoreface retreat is of paramount interest to archaeologists.

Several geologists have given some attention to the process of shoreface retreat (Bruun 1962; Curray 1964; Swift 1968, 1970; Swift and others 1972) and their work emphasizes the role erosion plays during landward migration of the shoreface. During the course of this project, it was recognized that the development of a conceptual model regarding the fate of the subaerial surface would be extremely useful in assessing the archaeological potential of the Continental Shelf. A synthesis of many geological publications covering a diversity of subjects has laid the initial framework for constructing this model. Major investigations concerned with sea-level change, continental shelf shallow structure, shelf morphology, sedimentation, and nearshore environments have been reviewed with the following objectives in mind. First, attention has been devoted to reconstructing the level of the ocean in the study area for the last twenty millennia. Second, specific attention has been given to the effect sea-level change has had upon the subaerial land surface. Last of all, several reconstructions of geological environments have been attempted for those areas for which sufficient data are available. On the basis of these reconstructions, we have undertaken an assessment of the likelihood of encountering former subaerial surfaces that are now either exposed or buried beneath the underwater surface of the Continental Shelf.

The term "subaerial surface" refers to that portion of the continental margin which was exposed to terrestrial weathering and biological action at the time when the CS was above sea level. The term does not indicate the length of time for which the surface was exposed and, consequently, a soil horizon does not have to be well defined. Active flood plains along many rivers would be classified as subaerial surfaces although part of their time was spent beneath flood waters. The surfaces of a beach or sand dune may also be classified as subaerial. But because these landforms constantly change as they are formed and subsequently eroded, their subaerial surface was seldom preserved intact. The amount of reworking of a subaerial surface is important to archaeologists since it corresponds with the potential integrity of archaeological resources should any be present. The amount of probable disruption or distortion of the archaeological record is proportionate to the amount of erosion experienced by a subaerial surface. In uneroded sites, the spatial context (positional interrelation) of the artifacts has a greater chance of retaining its full scientific value, whereas artifacts recovered from extensively eroded or transported subaerial deposits (for example, slopewash) have lost a greater percentage of their archaeological interest. In areas where our present archaeological knowledge is limited, artifacts from either type of context can provide important information. As a section of this report points out, some geological features of the shelf may have acted as artifact traps. Recovery of archaeological material from these features would help to expand our knowledge of an area about which, archaeologically speaking, we know next to nothing.

2.0 PROJECT ORIENTATION

The geographical limits for this project were established by the Bureau of Land Management. The project area consists of that portion of the continental margin lying between mean high water (mhw) and a depth of 200 m below mean sea level (msl), and extending from Cape Hatteras, North Carolina northward to the United States - Canadian Border. Within this area, geological investigations are intended to focus on the evolution of the shelf surface between 45,000 and 3000 B.P. in order to examine the distribution of cultural resources in the region.

Only those aspects of shelf geology which play a role in the distribution and preservation of prehistoric and historic cultural resources are addressed. Many aspects of continental shelf geology are of no direct use to this study. Under no conditions should the material in this report be construed as a synthesis of all aspects of the continental shelf geology.

Individuals interested in aspects of continental shelf geology not covered here should consult Emery and Uchupi (1972), TRIGOM (1974, 1976), URI (1973), USGS (1978), and similar general references for additional information.

Several aspects of this project are topics of ongoing research, and will require further field work for their resolution. For example, one of the most important problems affecting paleo-reconstruction is the problem of determining successive Holocene sea-level positions. Sea-level curves developed from coastal and inland sources significantly diverge from those obtained for the mid- or outer shelf (see Belknap and Kraft 1977; Milliman and Emery 1968; Newman 1977). In addition, few reliable points have been obtained which correspond to positions older than 9000 B.P. These and similar problems are recognized in this report.

3.0 METHODS AND MATERIALS

This report is based on a review of pertinent literature and discussions with individuals knowledgeable in the geology of the project area. Several geologists served as consultants during the course of this project and their expertise and guidance were extremely useful.

At the beginning of this study, a computerized literature search was used for a portion of the project area. The New York Bight sector was used to test the efficiency of such an approach for gathering data Inappropriate references were identified by title and removed from the computer literature list. A comparison of this list against the bibliographies found in major works (for example Emery and Uchupi 1973; CNA 1977; TRIGOM 1974, 1976; URI 1973) indicated that few additional sources were uncovered. The biggest problem with this approach was the choice of key words used in extracting references. The geological evolution of the Continental Shelf covers a broad variety of subjects and makes it very difficult to select a narrow list of key words for computerized search. Because over 80% of the references obtained were not useful, this method was abandoned as not cost effective. Computerized literature searches may be more useful for projects focusing on easily definable topics.

For the remainder of our study, articles and books were selected from the bibliographies found in major sources recommended to us by the BLM (for example CNA 1977; TRIGOM 1974, 1976; URI 1973) and from important references available from or recommended to us by our consultants (such as Edwards and Merrill 1977; Emery and Uchupi 1972; Field and Duane 1976; Kraft 1977; Newman 1977; Sheridan and others 1974; Stanley and Swift 1976; Stubblefield and others 1977; Swift and Sears 1974; Swift and others 1972; Swift and others 1977).

The scope of this project did not include field investigations. Some of the results of this study, however, suggest where future field work may be best located in order to answer some of the questions facing cultural resource managers. Furthermore, field investigations are necessary in order to verify some of the conclusions of this report. These investigations are mentioned in the recommendation section.

We have attempted to locate features as precisely as possible to allow for accurate locations of possible intact archaeological sites. The accuracy of most bathymetric data and easily definable buried features on the Continental Shelf, we feel, allow for a level of accuracy for site location to within three nautical miles. Some of the geological features discussed in this report, however, are not as accurately located because of the sampling methods used. The path of the buried Delaware River valley is one example. Twichell, Knebel and Folger (1977) mapped about 120 km of this feature on the middle and outer continental shelf off Delaware Bay, using 11 seismic profile transects. Transect spacing ranged from 3 km to about 26 km. As Twichell and others (1977) readily admit in their article, the sinuosity of this buried valley is not well known from the data at hand. For this reason, when discussing management policies, we have used slightly larger boundaries for outlining important buried geological features in our project area. Throughout this volume, however, we have delimited most features with the aid of those boundaries generally found in the geological literature; whereas in the concluding section we have used slightly larger boundaries in order to conform to the objectives placed before us by the BLM. Essentially, their guidelines for this project require that we make a go/no-go recommendation on a lease-block-by-lease-block basis. In order to minimize unnecessary destruction of archaeological resources on the Shelf, we have elected to draw boundaries which conserve these resources when not enough information is available. For example, the buried Delaware Valley is redrawn to fall inside of a large valley corridor which has a width slightly greater than the "height" (peak to trough) of the sinuous buried feature.

Larger boundaries for some buried features are also more realistic, since many of them have no sharp limits but rather consist of gradual facies changes which grade into another unit. Delimiting sharp boundaries from seismic profiles is often not an easy task. Several of the seismic profiles shown in Fig. I-1 vividly illustrate this problem. Consequently, it is important to keep these considerations in mind when each of the shelf compartments is reviewed and its buried geological features illustrated.

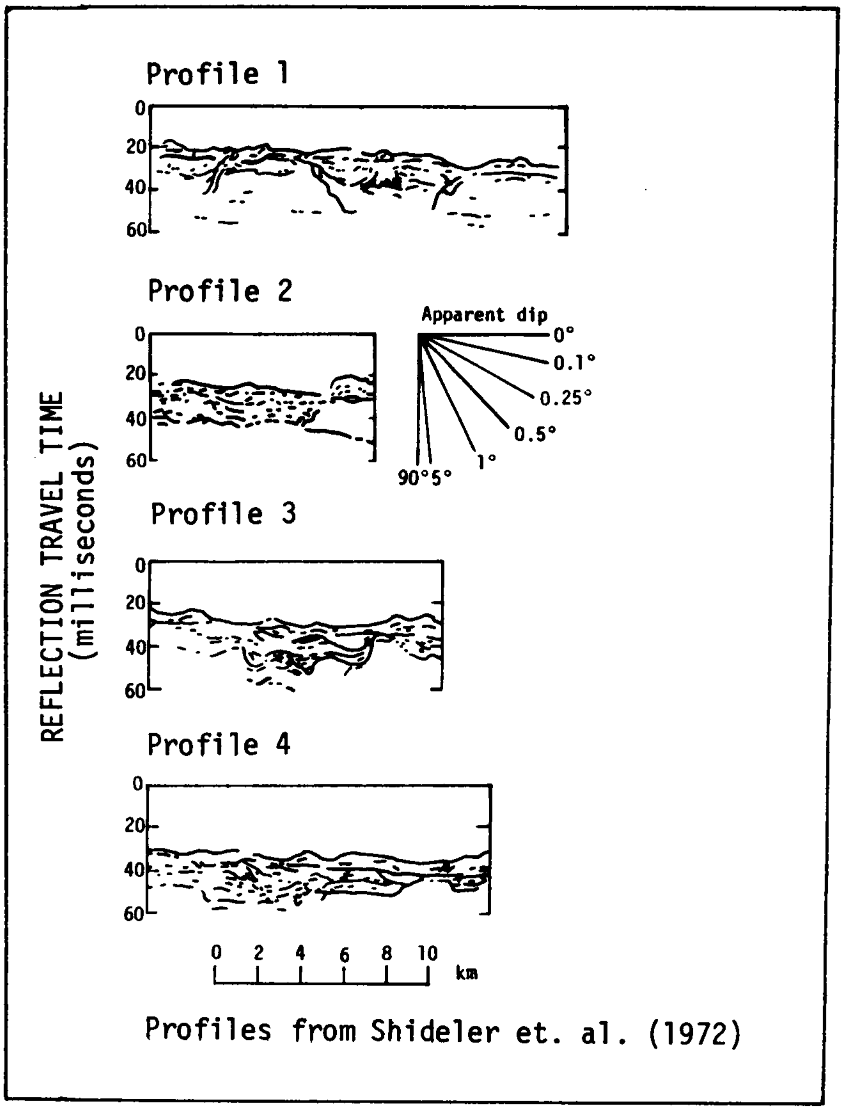


Fig. I-1

Line drawings of seismic profiles across the Virginia Beach buried valley. Showing fill sequences. After Swift and others (1977).

4.0 RESULTS OF RESEARCH

The results of this study can be grouped into the following general areas: formulation of a conceptual model regarding the fate of the subaerial surface during shoreface retreat, review of relative sea-level changes during the last 20,000 years, delineation of important geomorphological landforms, reconstruction of some aspects of the Continental Shelf's geological environment since the last lowstand (that is, since about 16,000 to 20,000 B.P.), and an assessment of the likelihood of encountering a Late Pleistocene/Holocene subaerial landsurface submerged or buried within the project area.

4.1 Conceptual Model

The conceptual model presented is an important tool for the understanding of the recent evolution of the Continental Shelf. For the past 15,000 - 20,000 years the net change in the level of the ocean throughout the project area has been a rise of from somewhere between 80 and 160 m (Dillon and Oldale 1978; Milliman and Emery 1968). This rise in sea level has resulted in the migration of the shoreline from a position many kilometers eastward to its position today. Fig. I-2 shows several positions of the shoreline during the Holocene.

During the migration of the shoreline from its easternmost (lowest) position at the end of the Pleistocene to its present position, all subaerial landsurfaces once exposed must have been passed over by the surf zone and the storm flows of the inner shelf. Some surfaces evolved through a series of inland and coastal environments which laid down new sediment over the earlier subaerial surface (such as flood plains, encroaching swamps and marshes, lower slopes covered by talus, or downhill creep). Other subaerial surfaces were eroded before the coastline migrated over them.

In order to understand the fate of a subaerial surface, whether it was buried (inactive) or exposed (active) when the coastline reached its location, it is necessary to look at some shoreface and nearshore processes.

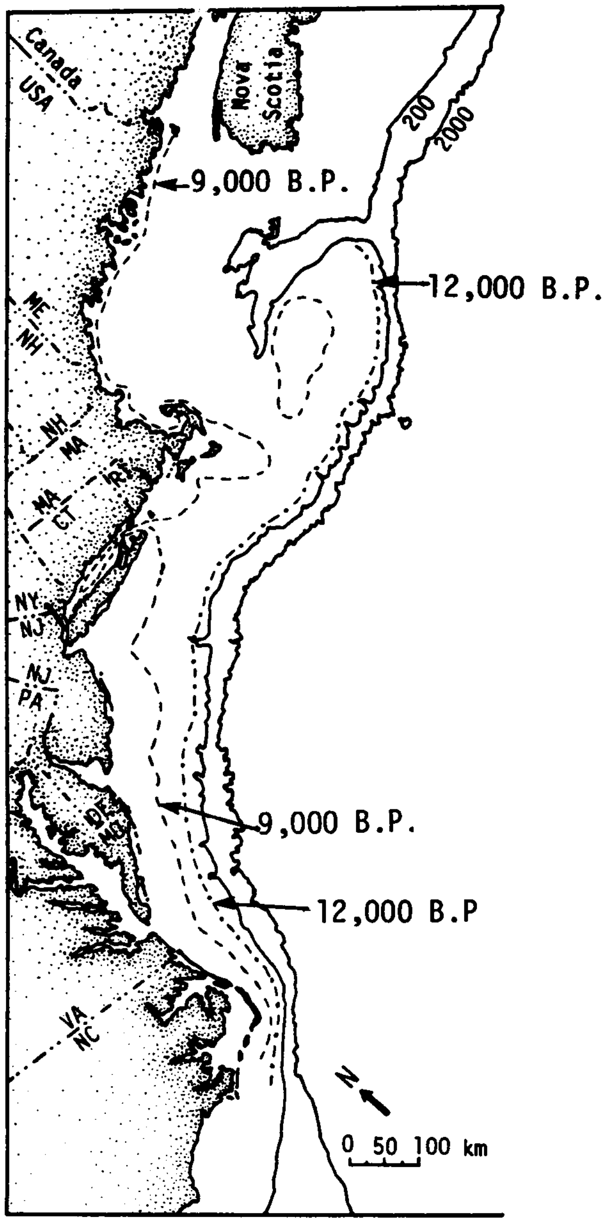


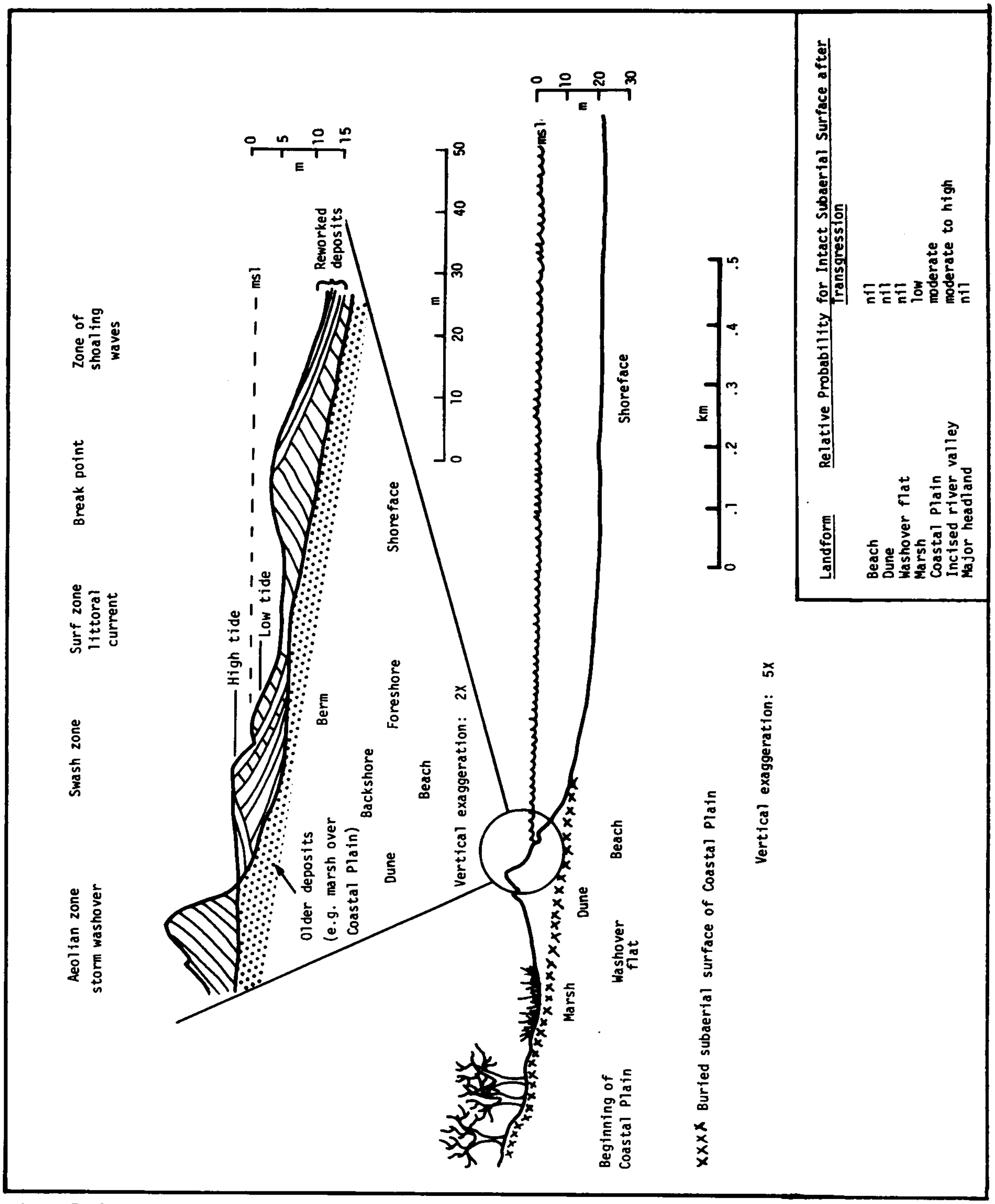
Fig. I-2

Maximum position of shoreline at 9,000 and 12,000 B.P. Please note that for the Gulf of Maine this represents about the maximum extent of emergence due to isostatic readjustment. South of Cape Cod, the maximum emergence was sometime between 15,000 and 20,000 B.P. and would have exposed an additional 10 to 15 km of the Continental Margin. See text for additional discussions. Shoreline positions based upon some data from Milliman and Emery 1968; Dillon and Oldale 1978; and other references found in text.

Shoreface and nearshore geologic processes have been studied by many researchers. The shore zone consists of several morphologic elements as shown in Fig. I-3 (Swift 1976a). The beach and shoreface zones are the recipients of energy expended by shoaling and breaking waves, and by intense coastal currents. The amount of energy expended along any particular coastline is a function of many variables, several of which are wave fetch, nearshore topography, tides, climate, and shoreface slope. These variables are not static but interactive, forming a dynamic system in which complex feedback may occur. The profile of a beach and shoreface responds to these variables daily, seasonally, and over greater periods of time. Shoreface erosion is common along the Atlantic coast and considerable landward migration of the shoreline usually takes place during major storms or specific seasons. The eroding surface generally appears in profile as an exponential curve, concave side up, with the steepest portion nearest the beach (Fig.I-4).

During the Holocene, sea level rose throughout the project area, with the result that the shoreline migrated landward. In most areas, this phenomenon has been accompanied by landward translation of the shoreface. However, in areas with high sediment influx, the shoreface may be able to migrate seaward given the right conditions. This may have occurred along the shorefaces of some early Holocene deltas now submerged on the mid-Continental Shelf. Along portions of eastern Massachusetts (Redfield 1965) and northeastern Delaware Bay (Meyerson 1972), marsh areas have migrated over shallow bays during the last few thousand years. Consequently, erosion of the shoreline is not mandatory given sea-level rise. Sometimes local conditions may interact to produce a different scenario.

Marine transgression is essentially a re-leveling process. Landward migration of the shoreline results in the reduction of relief in a region. Pre-transgressive hills and high areas are eventually truncated by waves and currents. Valleys, basins and other low areas act as depositional centers, receiving sediment removed from these higher areas. The detailed studies done by several researchers (such as Kraft 1971, 1977; Kraft and Allen 1975; Kumar and Sanders 1975; Sheridan and others 1974) indicates that marine transgression usually removes or redistributes between 10 and 20 m of unconsolidated sediment. Profiles across coastal and nearshore zones along the Atlantic coast illustrate this process and generally exhibit a drop of 10 to 15 m over several kilometers seaward of the beach. This change in relief is easily seen in the bathymetry along the northern New Jersey coastline, for example. Fig. I-5 illustrates eight families of profiles from the Delaware coast whose positions indicate net erosion. The shoreface profile may be idealized as representing an exponential curve. This curve consists of two main subdivisions: the beach portion above mean sea level and the lower section seaward. Fig.I-5 shows changes occurring within the lower or seaward portion of the profile. Changes within 9 beach sections of the arc during a 5-month period are shown in Fig.I-6. As this figure illustrates, the beach represents an extremely active landform being reworked by storms, tides, and wind. There is virtually no chance of an intact beach surface's being preserved after transgression.



Landform	Relative Probability for Intact Subaerial Surface after Transgression
Beach	nil
Dune	nil
Washover flat	nil
Marsh	low
Coastal Plain	moderate
Incised river valley	moderate to high
Major headland	nil

Fig. I-3 An example of coastal landforms found along an open coast. The buried subaerial surface is destroyed as erosional shoreface retreat proceeds landward. Adapted from Swift 1976a. The table inset above right lists relative probabilities for the preservation of several landform types.

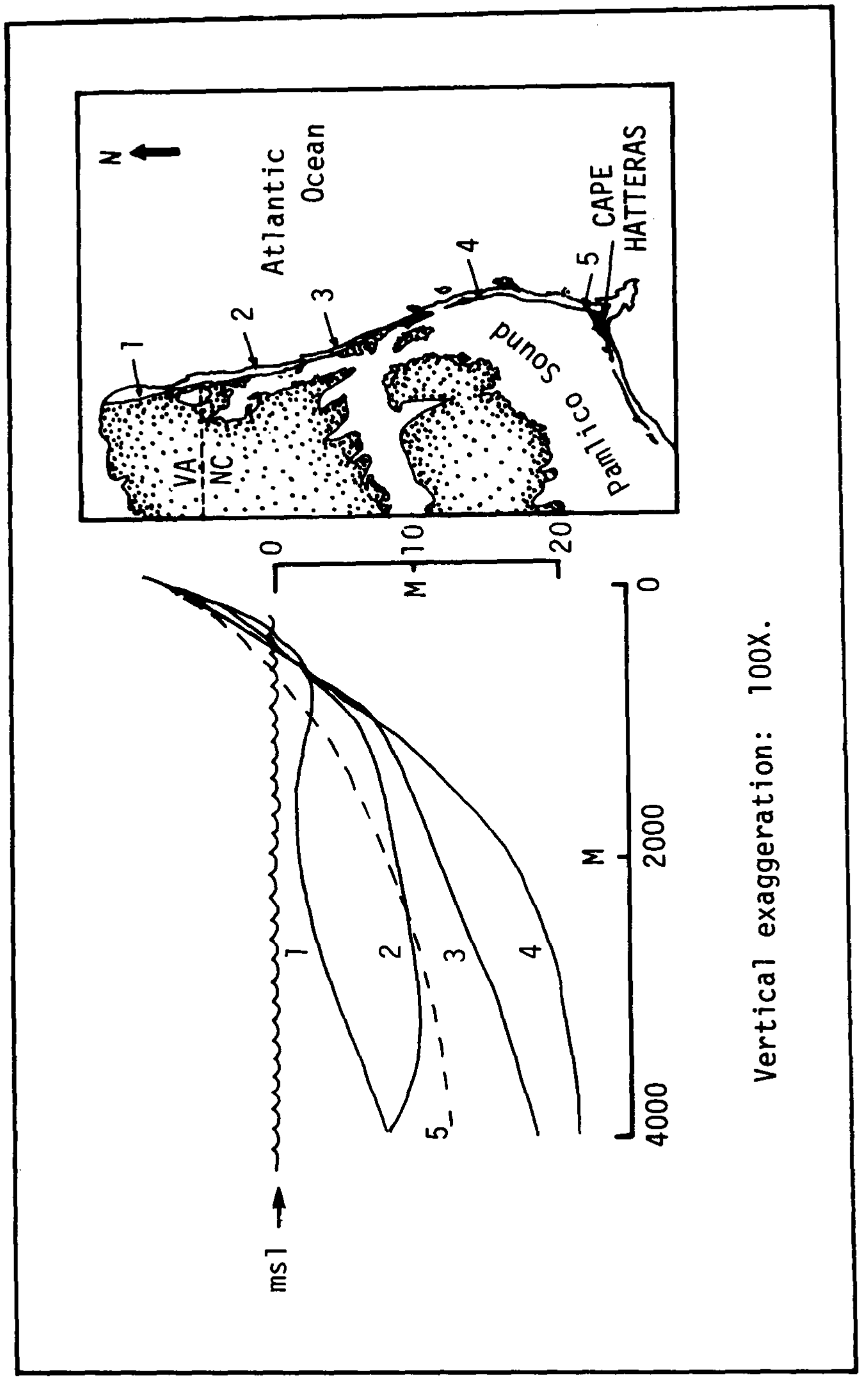


Fig. I-4

Examples of coastal profiles along the North Carolina-Virginia Coastal Compartment (after Swift 1975a).

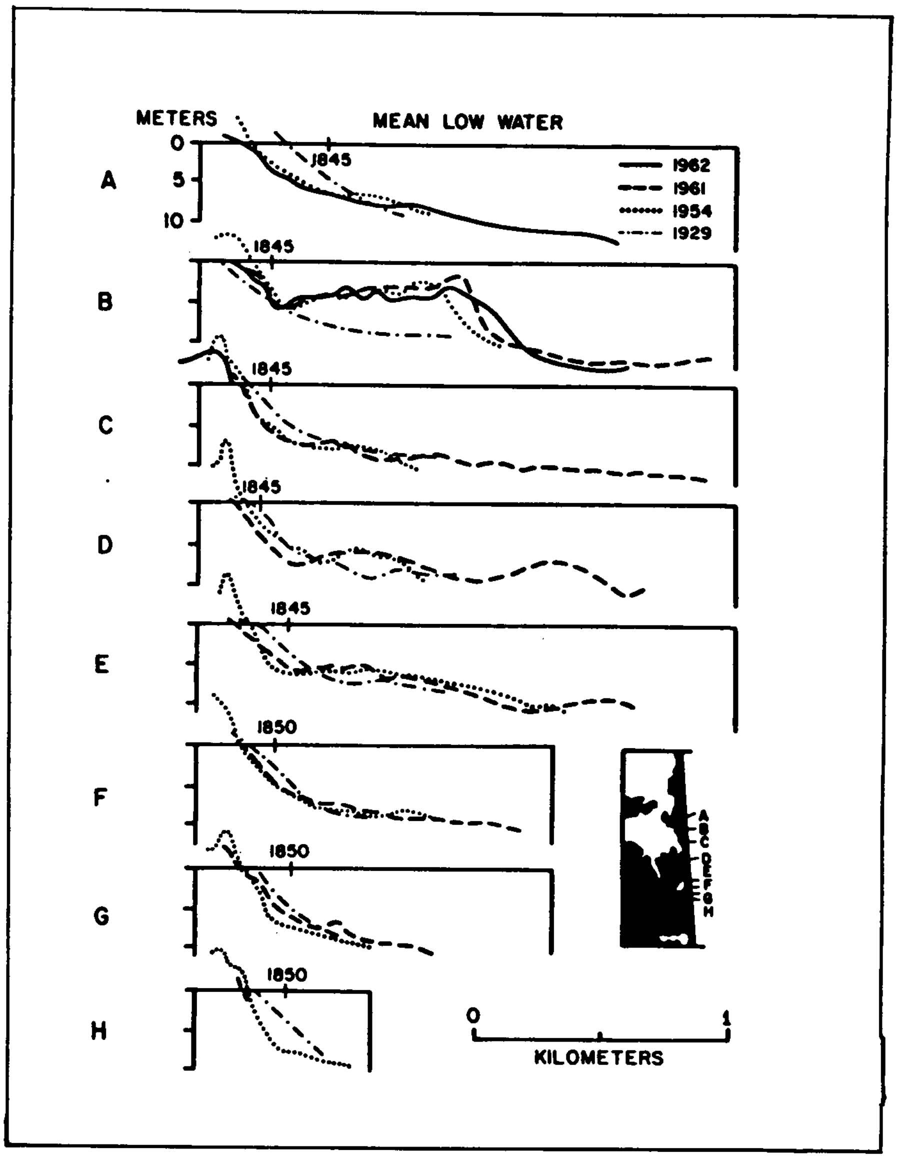


Fig. I-5

Several examples of shoreface envelopes of erosion showing coastal retreat along Delaware (after Swift 1976a). Time series showing change in shoreface profile at eight locations along the Delaware Coast. Mid-nineteenth century dates indicate approximate position of beach face at that time. Vertical exaggeration: 75x.

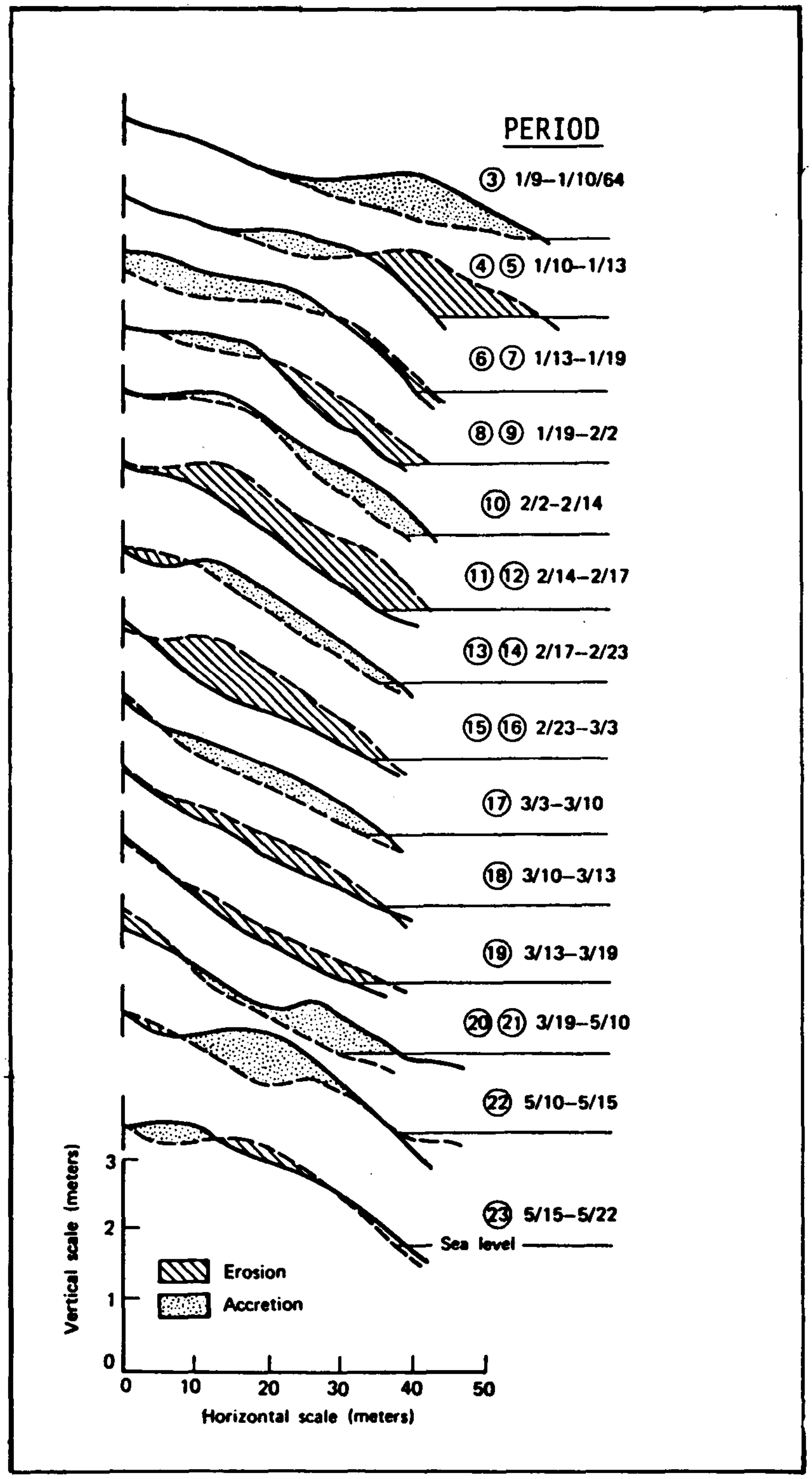


Fig. I-6

Characteristic change within a beach profile (Swift 1976a). Change within a beach profile during a 5 month period. Amount of erosion or accretion indicated by symbols. Vertical exaggeration: 10x.

Swift (1975a) has suggested some developmental histories for positions of the Virginia-northern North Carolina coast which represent various types of recent Holocene transgression (Fig. I-7). As indicated on each profile, a certain amount of erosion has taken place during landward migration of the shoreface at each station. Swift (1976a) has also identified several types of translation of the shoreface. Sea-level change, of course, plays an important role in shoreface translation. Fig. I-8 gives several modes of shoreface development as a function of sea-level change. Of interest to this project are profiles A, B, and C since they portray transgressional shoreface types. Except for several protected areas along the coastline under study, Holocene transgression has rarely involved type D (Fig. I-8: depositional regression with rising sea-level). The inner Delaware Bay shoreline north of Cape May, however, is one of the few exceptions where type D has occurred in recent times (Meyerson 1972).

Fig. I-9 is an example of erosional shoreface retreat and transgression of a barrier island coastline. As this figure illustrates, the pre-transgressive subaerial surface escapes destruction by a protective covering of lagoonal muds. Some erosion of the subaerial surface may occur along the lagoon shore or beneath tidal channels and between sand ridges seaward of the barrier. Tidal inlets are probably the most destructive element along barrier island coastlines. Fig. I-10 illustrates the depth to which these features may erode; note that erosion usually extends well into the pre-transgressive deposits. Tidal inlets may reach depths 3 to 4 times as great as that of the adjacent sea floor. Fig. I-10 shows the stratigraphy commonly found along low coasts that are undergoing erosional transgression. In this particular case, erosion removes the subaerial surface on the seaward side of the barrier. The peat deposits shown in the sequence are frequently encountered along barrier coastlines and beneath lagoonal deposits. Rate of sea-level rise and amount of lagoonal deposition play an important part in allowing for the retention of peat deposits and the underlying subaerial surface (Fig. I-11). In some cases, peat may not form or be retained and organic silt may be found in its place at the basal section of the lagoonal sequence (Sanders and Kumar 1975a and b).

It has become apparent during the course of this project that few coastal landforms enter the geological record intact. The subaerial portions of barriers, spits, beaches, and dunes do not become submerged without being truncated and reworked. To date, the few examples of a "submerged" spit encountered in the project area represent truncated sequences of their basal sections (Kraft 1971; Kraft and Maurmeyer 1978; McMaster and Garrison 1967). Consequently, these coastal landforms (spits, barrier islands, open ocean beaches, nearshore dunes, bay mouth barriers, etc.) and similar subaerial environments have little chance of being represented by intact deposits on the Continental Shelf. Many of these coastal landforms are illustrated in Fig. I-12 . In no case do these landforms and their subaerial surfaces become "submerged" or "drowned" in the classical sense as some researchers have suggested in the past. Relict subaerial surfaces do not exist on the Continental

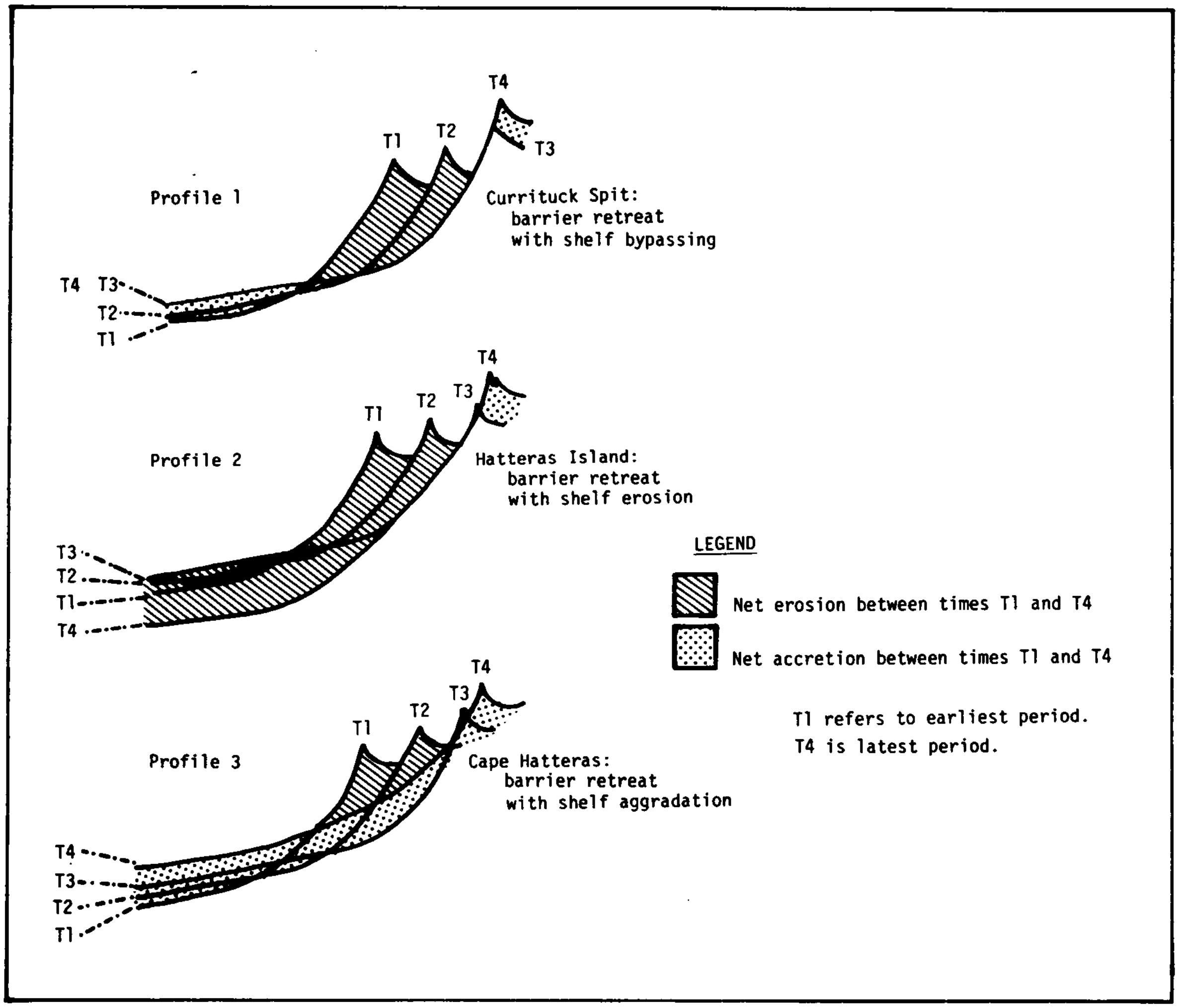


Fig. I-7
Hypothetical sequences of shoreface migration for sections of the Virginia-northern North Carolina Coast (after Swift 1975a).

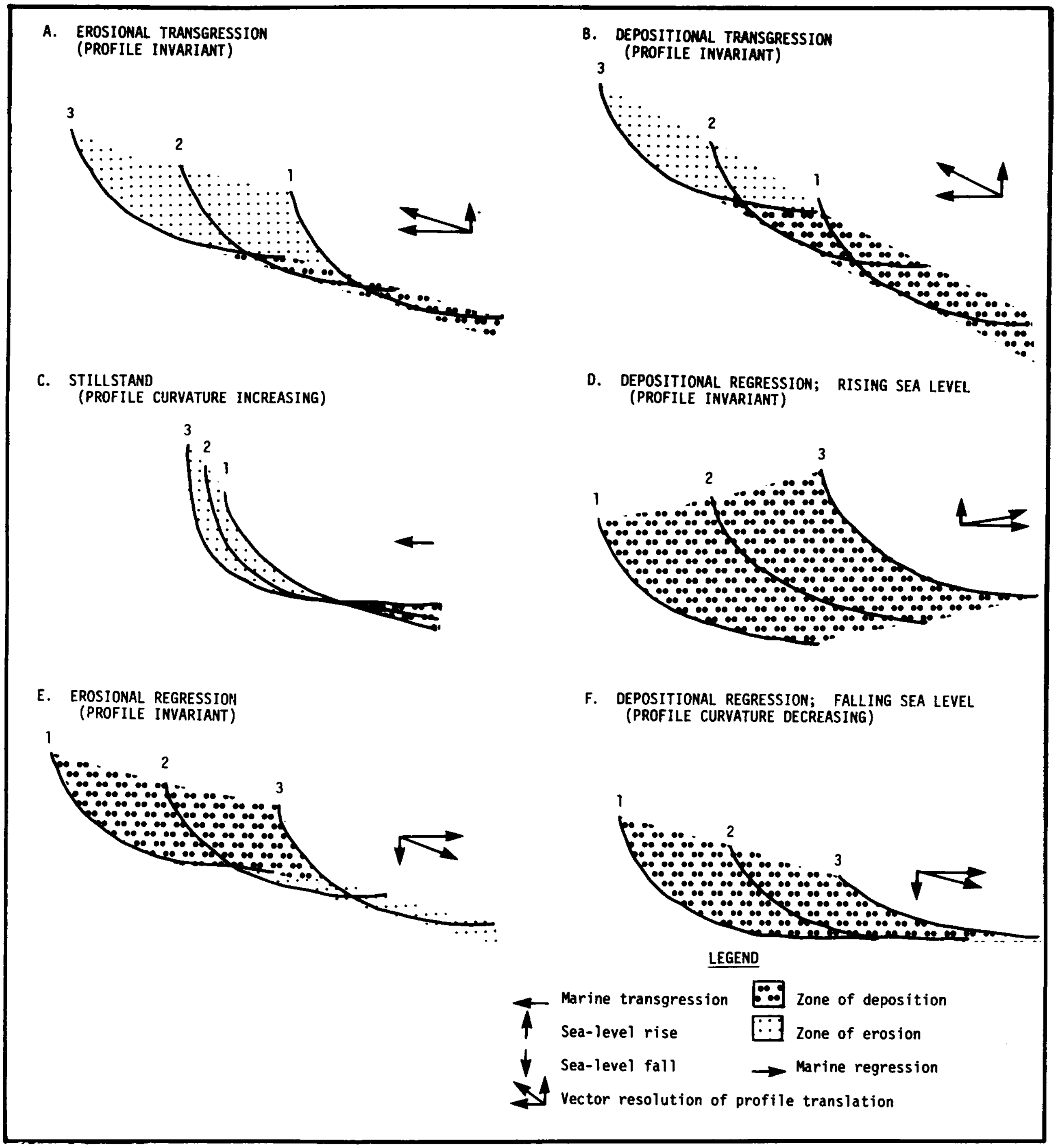


Fig. I-8

Shoreface profiles as a function of sea-level change and sedimentation (after Swift 1976a). Six examples of shoreface translation as a result of sea-level change (↑ rise, ↓ fall) and sediment availability.

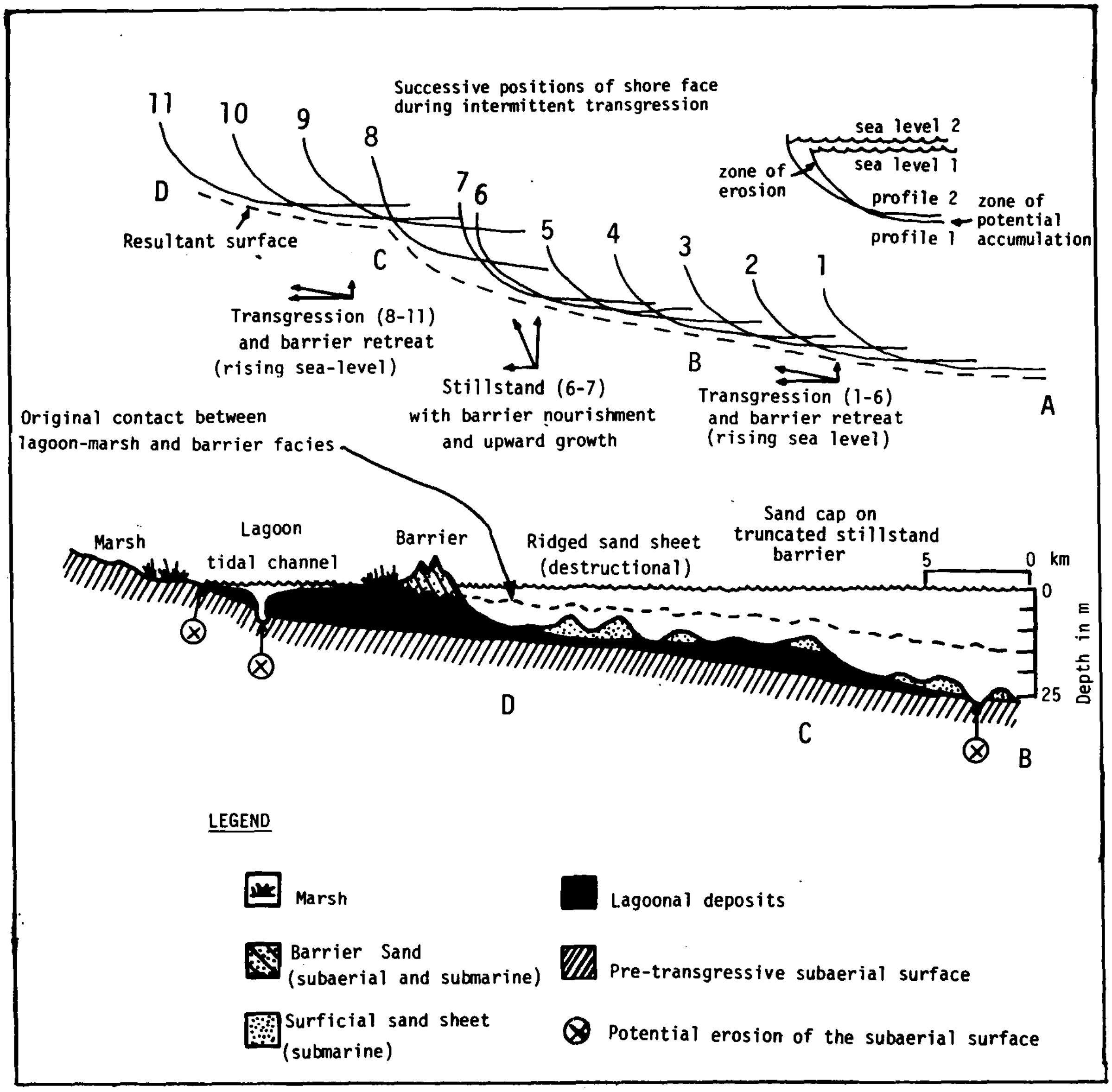


Fig. I-9
An example of shoreface erosion and landward migration of the shoreface profile (adapted from Swift 1975a). Vertical exaggeration: 200x

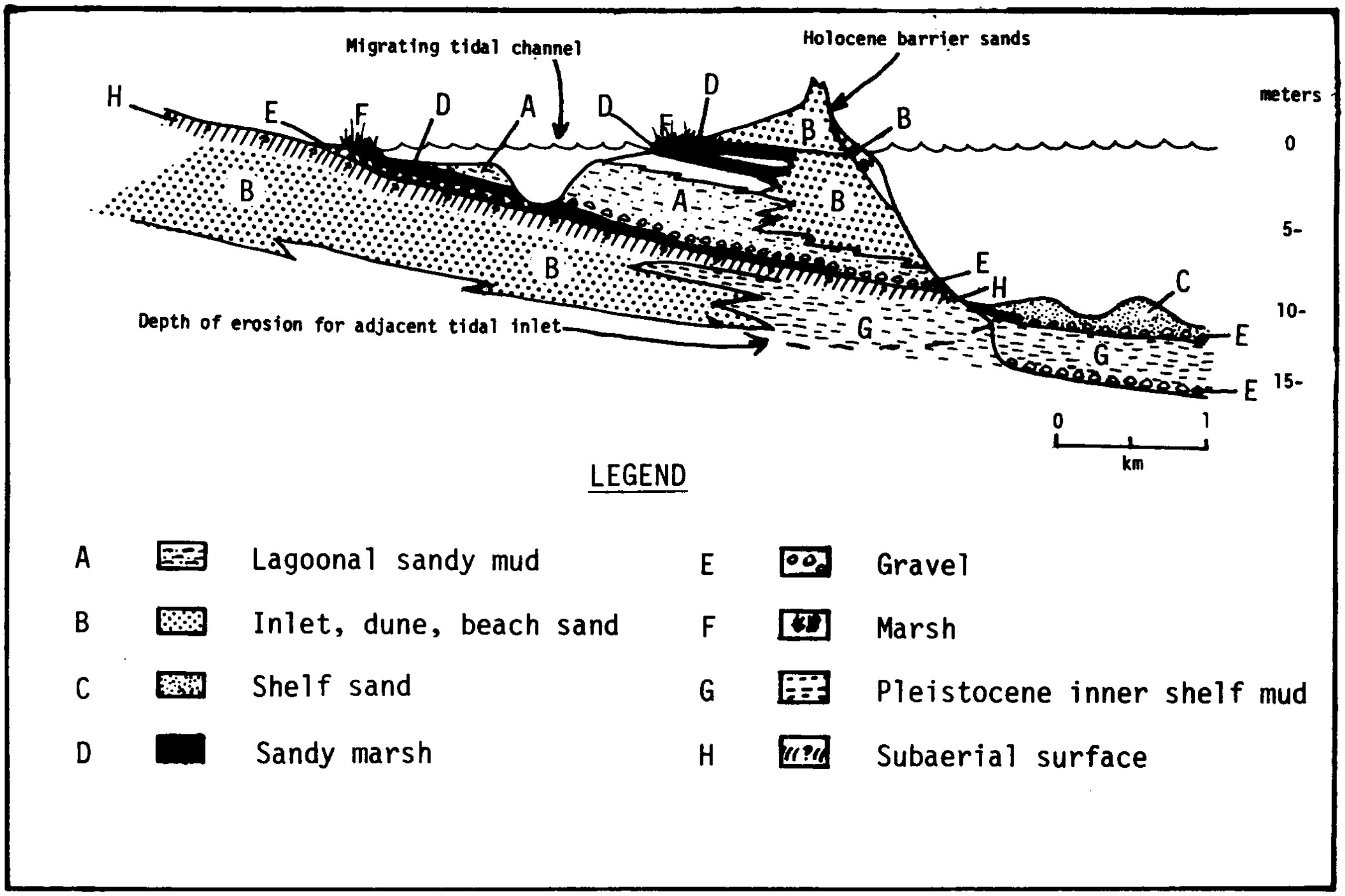
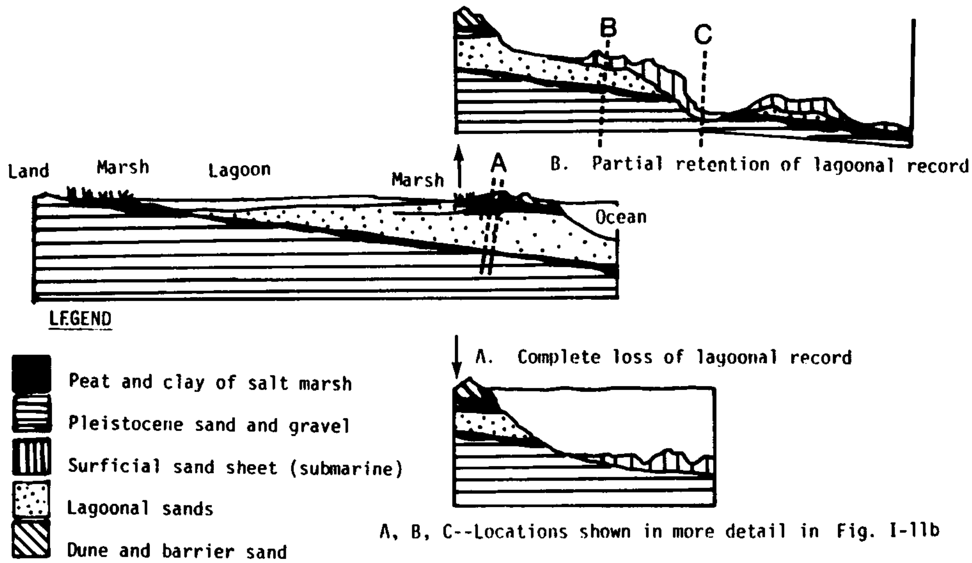
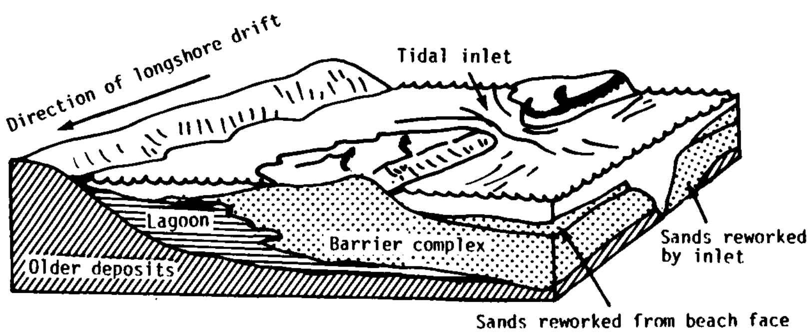


Fig. I-10

Stratigraphic model for a low coast undergoing erosional shoreface retreat (adapted from Swift 1976b). Schematic diagram illustrating the stratigraphy of a barrier coast undergoing erosional shoreface retreat. In this particular example, complete erosion of the subaerial surface has occurred about one kilometer seaward of the barrier beach. Disconformable surface on which sand sheet rests is cut by waves, and in some areas by tidal and inlet channels. Vertical exaggeration: 100 x.



Two simplified examples of erosional shoreface retreat. Tidal marsh peats and clays are followed by lagoonal sediments, these in turn by tidal marsh peats and clays, and these by dune sands and beach sands of the barrier beach. Further rise in sea-level and landward shift of barrier beach involve complete (A) or partial (B) destruction of marginal record, before overlap by offshore sands (after Fischer 1961).



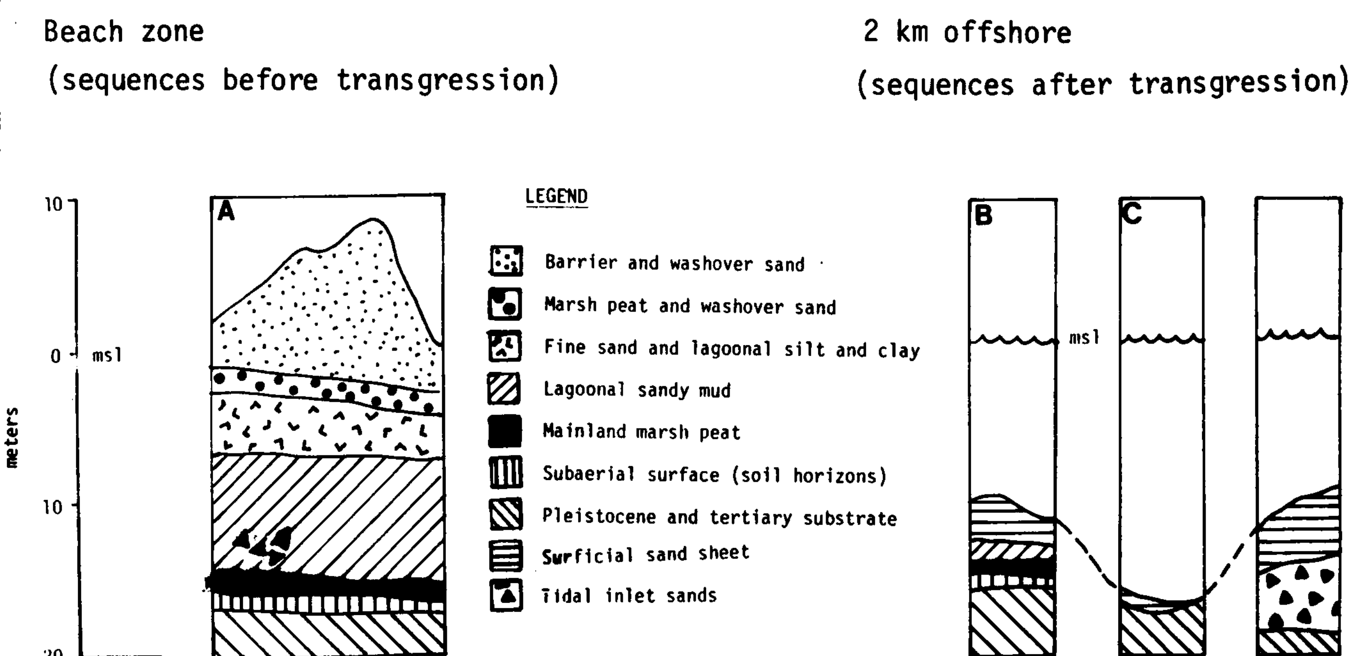
Simplified stratigraphic relationships caused by landward-retreating barriers with migrating inlets (after Swift 1968).

Fig. I-11a

Several examples of nearshore stratigraphy generated by transgressing barriers and erosional shoreface retreat.

Fig. I-11b

Several examples of nearshore stratigraphy generated by transgressing barriers and erosional shoreface retreat.



Erosional shoreface retreat along a barrier shoreline and three hypothetical sequences encountered 2 km offshore. Vertical exaggeration: 40X.

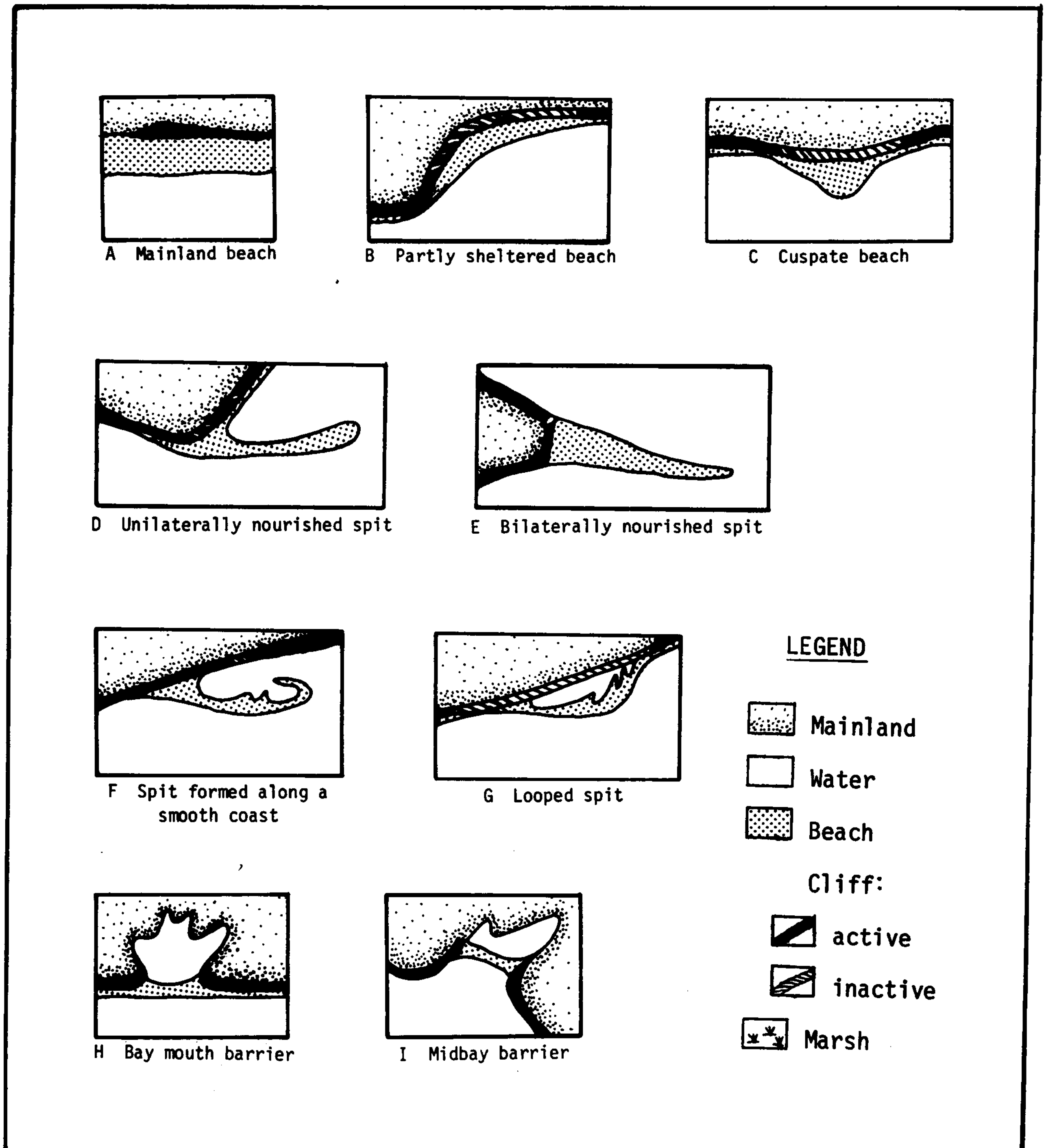


Fig. I-12a

Examples of coastal land forms that generally do not have their subaerial surfaces preserved after marine transgression. Adapted from Swift (1976a).

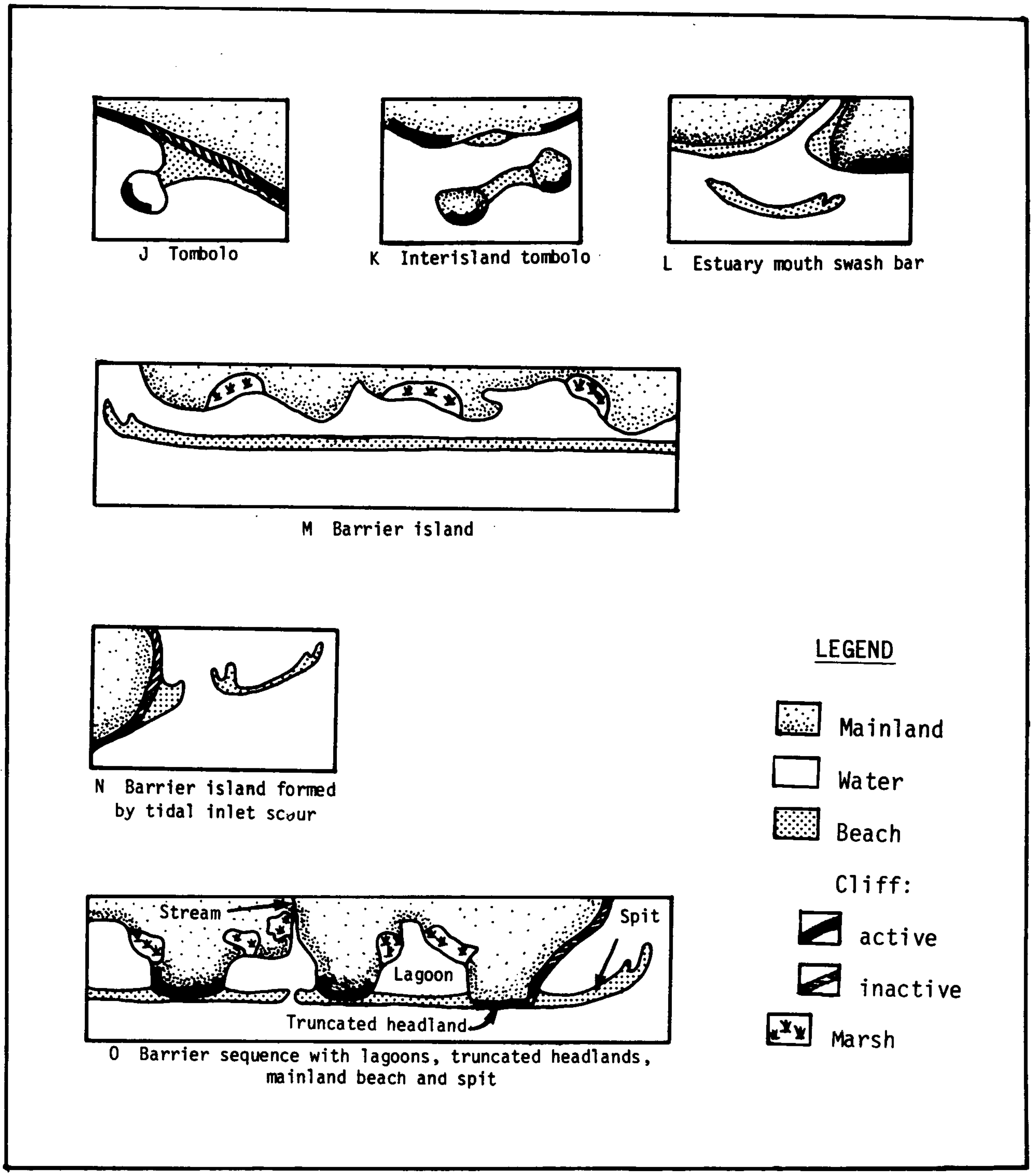


Fig. I-12b

Examples of coastal land forms that generally do not have their subaerial surfaces preserved after marine transgression. Adapted from Swift (1976a).

Shelf except where they have been buried by overburden during transgression. The transgressive "sand sheet" found on the Continental Shelf today is also not relict but responds to tidal and storm-generated currents. These concepts form the framework around which the remainder of this report is structured.

4.2 Physiographic Overview

Most of the region under study in this report lies within the continental shelf portion of the Atlantic Continental Margin. However, a thin strip along the eastern boundary, generally less than 10 km wide, falls within the upper portion of the Continental Slope. The shelf break separates the outer edge of the Continental Shelf from the Continental Slope and is a zone where more steeply sloping submarine topography occurs (slopes of 2 - 15^{o}). The shelf break lies deeper as one moves northward (Emery and Uchupi 1972:22). It is found at about -50 km at Cape Hatteras, N.C. Off northern New England, it occurs at about -150 m.

The Continental Shelf is that physiographic region lying between the present coastline and the shelf break. It varies in width from about 40 km off Cape Hatteras and becomes increasingly wider toward the north. Off the coast of Delaware the shelf width is about 100 km, while south of Nantucket it is approximately 130 km. North and east of Nantucket, the shelf has been glacially eroded and is topographically less uniform as a result of Pleistocene events.

The 3 major physiographic regions, as defined for the purposes of this report, are shown in Fig. I-13. From south to north, these are the Middle Atlantic Bight, Georges Bank, and the Gulf of Maine. The Middle Atlantic Bight has been subdivided into 5 compartments and the Gulf of Maine into 3 subregions (Fig. I-13). Also shown on this figure are the major sounds and bays and the approximate position of the fall line separating the emerged and submerged coastal plain from the crystalline rocks of the Piedmont, New England Uplands, and Gulf of Maine. The Middle Atlantic Bight is that portion of the Atlantic coast stretching from Cape Hatteras northward to Cape Cod. It may be subdivided into 5 shelf compartments (Swift 1970). South of the Hudson Valley, the shelf consists of 3 coastal compartments separated by 2 large bays (Delaware and Chesapeake Bays). The compartments are referred to as the North Carolina-Virginia shelf, Delmarva shelf, and the New Jersey shelf. North of the Hudson Valley, the Middle Atlantic Bight has been divided into 2 shelf compartments: the Long Island shelf and the southern New England or Block Valley to Cape Cod shelf. A general discussion of these three major physiographic regions follows.

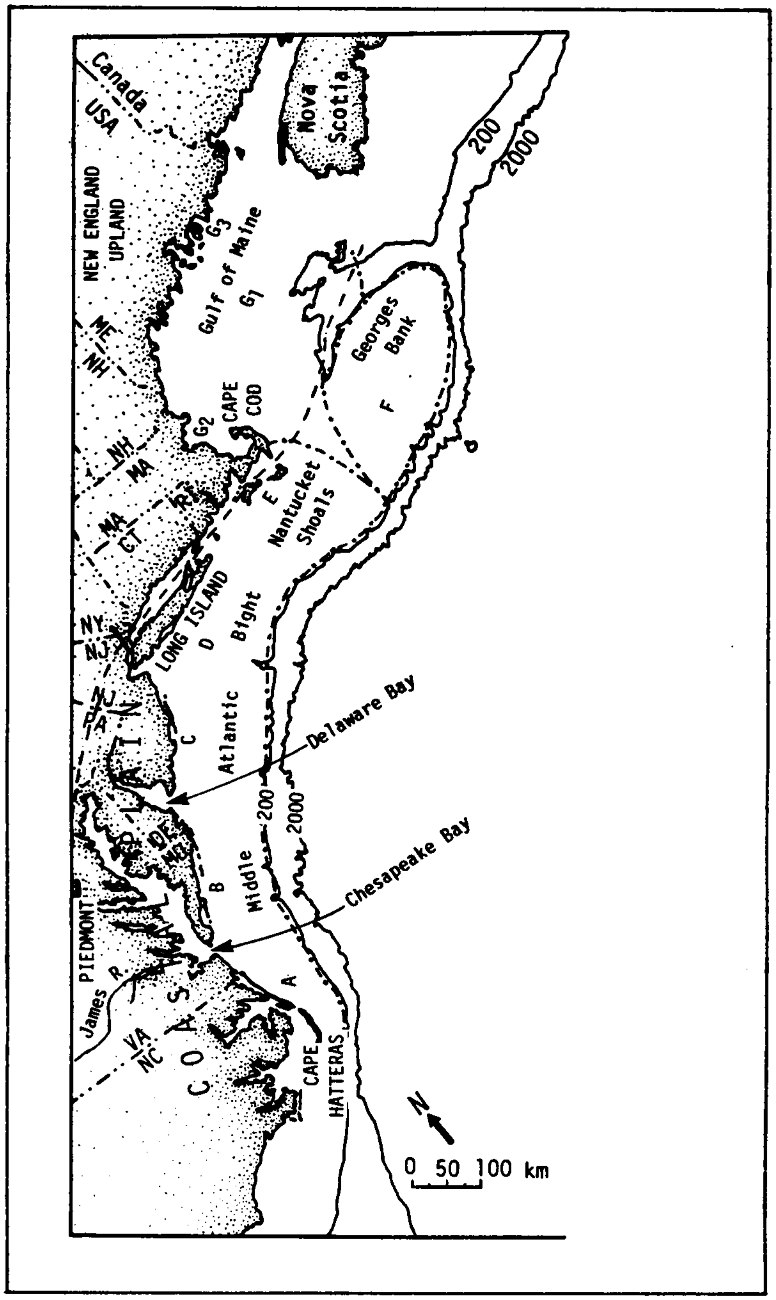


Fig. I-13

Physiographic regions used in this report. From south to north these are as follows: (A) Northern North Carolina--Virginia Shelf; (B) Delmarva Shelf; (C) New Jersey Shelf; (D) Long Island Shelf; (E) Southeastern New England Shelf; (F) Georges Bank Shelf; (G_1) Central Gulf of Maine; (G_2) Southern Gulf of Maine Mainland Shelf; (G_3) Northern Gulf of Maine Mainland Shelf.

It is useful to draw upon this classification because each compartment contains the continuation of at least 1 major drainage divide. During the Holocene, these divides extended onto the (then exposed) Continental Shelf. South of the Hudson, 2 broad bays (ancestral Delaware and Chesapeake Bays) separated the compartments and were the termini of two large river systems. As transgression occurred during the Holocene, these river valleys became broad estuary retreat paths (see Swift 1973). The higher ground on the flanks of these valleys would have been the location for minor streams and rivers. The two shelf compartments north of the Hudson shelf valley are referred to in this report as the Long Island shelf and the southern New England shelf. The Block Valley shelf separates the 2 compartments.

These shelf compartments make up the Middle Atlantic Bight. It has been set apart from the Gulf of Maine and Georges Bank regions because its Quarternary evolution was quite different. In the Gulf of Maine and Georges Bank regions, topography has been strongly influenced by Pleistocene glacial events. Since deglaciation, these two northern regions have had estuary systems which probably carried less suspended sediment during most of the Holocene than they had in earlier periods. Their nearshore regions are much steeper when compared to the southern portion of the Continental Shelf which has a wider, more gently sloping shelf profile. Because the Shelf in the southern region has a more gradual slope, large bays and sounds were able to exist during most of the Holocene. The nearshore portion of the Shelf east of Long Island, however, also experienced glaciation during the Late Wisconsin (Pratt and Schlee 1969). But because its mid-shelf slope is much more gentle and Late Wisconsin glaciation did not penetrate very far southward it is grouped with the southern region. Its topography is predominantly influenced by fluvial systems similar to those for the southern portion of the project area. Partly because of the topographical differences, barrier island and coastal marshes are more extensive along the Middle Atlantic Bight than they are to the north. This contrast probably held true throughout most of the Holocene (see for example Field and Duane 1976).

The landward migration of barrier island-marsh complexes has left beneath much of the present transgressive sand deposits patches of what is sometimes referred to as a carpet of lagoon sediments. The origin of barrier islands along this region may have been predominantly that of mainland beach detachment (see Swift 1975a). In comparison, the Gulf of Maine coastline experienced far less extensive barrier island development during the Holocene. Those barrier islands which did develop probably owe their origin to coastwise spit progradation as summarized by Swift (1975a:12-19). As our model indicates, little evidence is left in the sedimentary record by a migrating barrier island to enable one to reconstruct its origin.

Because of the steeper nearshore topography along the Gulf of Main region, marshes and lagoons are much smaller in size. Considerable relief, however, has allowed narrow but deep estuaries to penetrate many tens of kilometers inland as sea level rose.

Another important difference between the Gulf of Maine and the Hatteras-Cape Cod coast during the Holocene was the extent of glacial isostacy. This will be discussed in detail in the section on sea level. Besides differences in relative sea-level changes between the Gulf of Maine and the Hatteras-Cape Cod regions, tidal ranges also differ significantly at present. How far back these differences can be extrapolated is not easily determined. Fig. I-14 gives the range in meters for tides in general throughout the study area. From this figure, it is apparent that tides are normally less than 1.5 m along the Hatteras-Cape Cod coast, except for Long Island Sound. The Gulf of Maine experiences mean tide ranges at least twice as great as those of the southern regions (excluding Long Island Sound). Greater variation in tidal range affects the distribution of some nearshore biotic communities which in turn may have attracted prehistoric groups to specific locations for the purpose of exploiting these resources.

Fig.I-15 gives the variation in tidal currents presently found in the project region. The distribution and intensity of these currents give an indication as to where bottom sediments may have been subjected to intense currents in recent time. As the figure suggests, the mouth of Chesapeake and Delaware Bays, Nantucket Shoals, eastern Long Island Sound, and all of Georges Bank have experienced significant tidal currents during the late Holocene. This current activity would have had a good chance of severely altering or destroying portions of the submerged subaerial surface and any associated archaeological remains.

Of importance to archaeologists concerned with the Holocene evolution of the Continental Shelf is a reconstruction of major river systems and their drainage basins. In order to begin to reconstruct the location of these river systems during periods of lower sea level, it is useful to know the major drainage systems along with their basin size and present day discharge (Table I-1). To the extent that climatic effects other than deglaciation (i.e., rainfall rather than glacial meltwater) affected this region uniformly during the last 10,000 years, some generalizations can be offered regarding paleo-drainage systems.

During the Early Holocene, each of these river systems drained more terrain since much of the CS was exposed. The Hatteras-Cape Cod region had a substantially larger increase in subaerial surface area than the Gulf of Maine section because of the more extreme withdrawal of the sea (i.e., absence of isostatic downwarping). Glacial meltwater, however, greatly affected the northern streams, making it extremely difficult to determine river discharge between 11,000 and 20,000 B.P.

The average intensities of wind and waves in the study area differ only slightly along this portion of the Atlantic coast. Both wind and wave intensity increase slowly toward the north from Cape Hatteras. The greatest difference is in the frequency of waves higher than 3.5 m, which nearly doubles as one moves northward from Cape Hatteras to Nova Scotia (Emery and Uchupi 1972: 250). Since the landward translation of the shoreface is partly a consequence of expended wave energy, wave

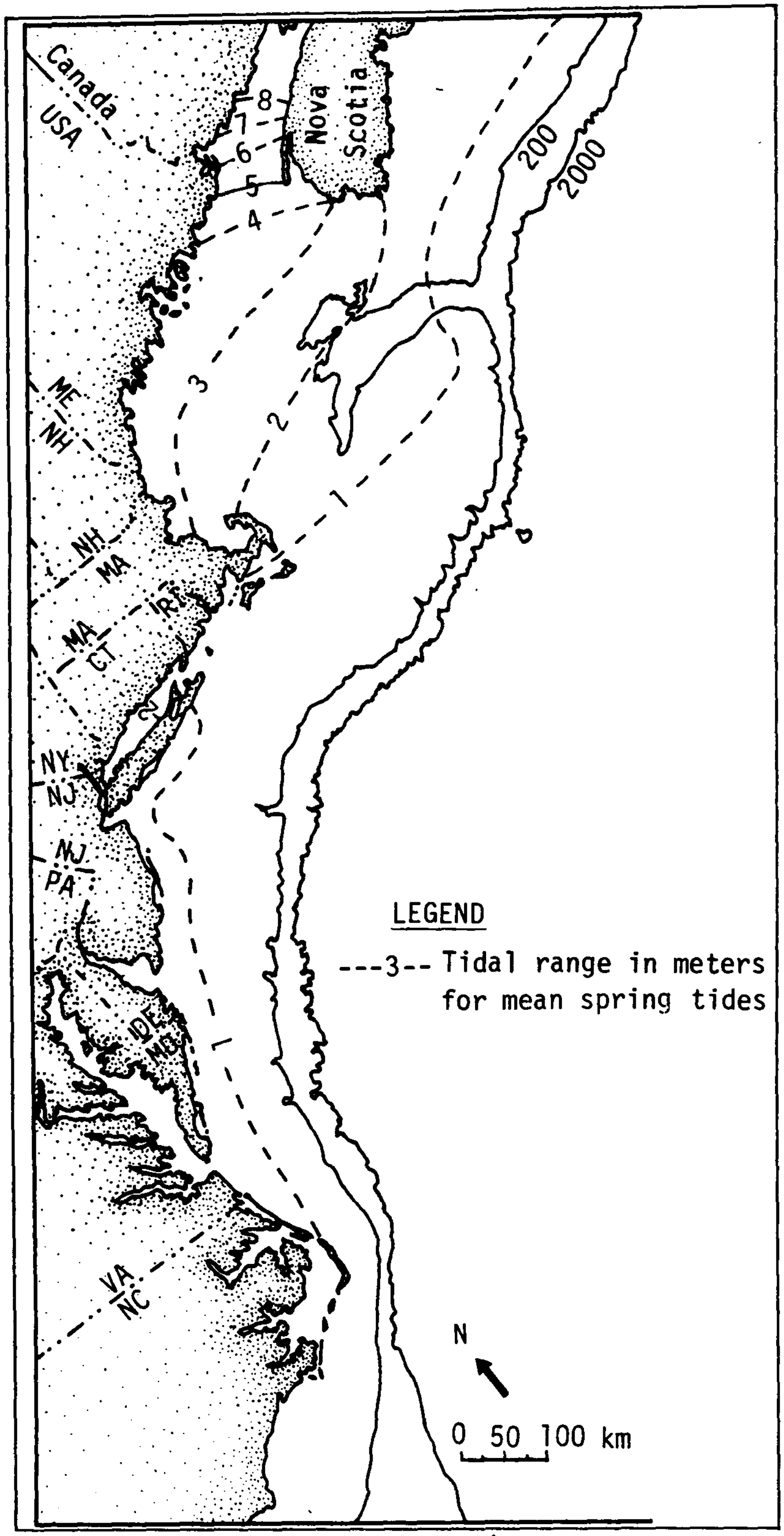


Fig. I-14 Co-range of mean spring tides (after Emery and Uchupi 1972).

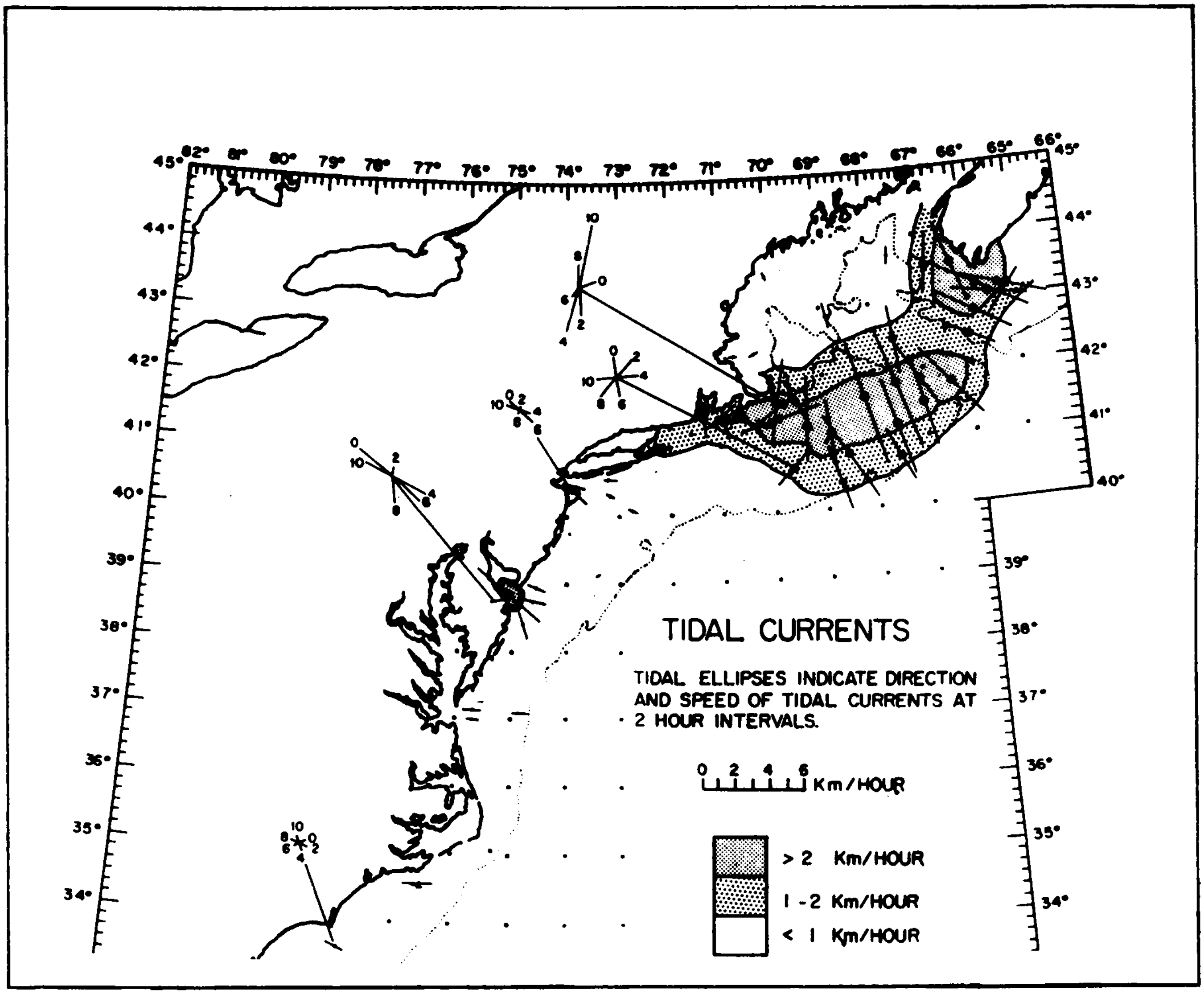


Fig. I-15
Intensity of surface tidal currents (after Emery and Uchupi 1972).

Table I-1: Major drainage systems along the project area and their areal extent and discharge. Data from URI (1973:10-6) and TRIGOM (1974b:5-16).

DRAINAGE SYSTEM	BASIN AREA (km^2)	WATER DISCHARGE km^3/yr.
Connecticut River	28,416	19.1
Hudson River	34,304	20.3
Delaware River	33,536	17.8
Susquehanna River	62,419	32
Potomac River	24,996	12.2
James River	26,624	8.9
Roanoke River	24,960	7.7
Penobscot River	40, 652	13
Kennebec River	24,976	15
Merrimack River	22,319	7

difference may have contributed significantly toward altering the subaerial surface during transgression. Typical beach profiles of the area also increase slightly in height towards the north (Emery and Uchupi 1972). Coarse sediment texture is one important factor which contributes to steeper northern beach profiles.

The direction of nearshore currents is shown in Fig. I-16 as deduced from beach shapes. These currents can be grouped to form at least 6 cells of convergence within the study region. Each cell is focused around a major bay mouth or estuary. Offshore bottom drift is shown in Fig. I-17 and also shows clustering toward major bays and estuaries as well as heading toward the shoreline.

4.3 Sea-Level Change from 20,000 B.P.

Of great importance to prehistorians is the issue of shoreline position during the Late Quaternary. Since there is little convincing evidence that human groups entered the New World before 18,000 years ago (Newman and Salwen 1977), it becomes unnecessary to review sea-level positions during earlier periods. Knowledge of sea-level positions before the Late Wisconsin period is poor and based on only a few scattered data points (Dillon and Oldale 1978; Emery and Merrill 1978).

The position of the shoreline along the North American Continent played an important role in determining the amount of land open to occupation by prehistoric groups entering the seaboard region. The shoreline formed a natural barrier that limited the eastern extent of occupation along the Atlantic coast. As the level of the ocean changed in response to changing climate (glaciation and deglaciation), portions of the continental margin were alternately exposed and inundated.

Before reviewing recent sea-level investigations, it is useful to consider some important aspects of the methodology used to construct sea-level curves.

Inferences regarding higher and lower sea levels along portions of the Atlantic coast have been made for over a century. Early researchers emphasized the position of former beach deposits found above sea-level, or submerged forests (Lyons and Goldthwait 1934; Sears 1905), as evidence for different sea-level positions in the past, but until the discovery of radioisotope dating methods, few conclusions could be drawn regarding these changes over time.

Radiocarbon dating has added much information on sea level change since the Late Pleistocene. To date, there have been 2 basic approaches to determining earlier sea levels. The first approach used by investigators was to collect specific types of peat material. The choice of

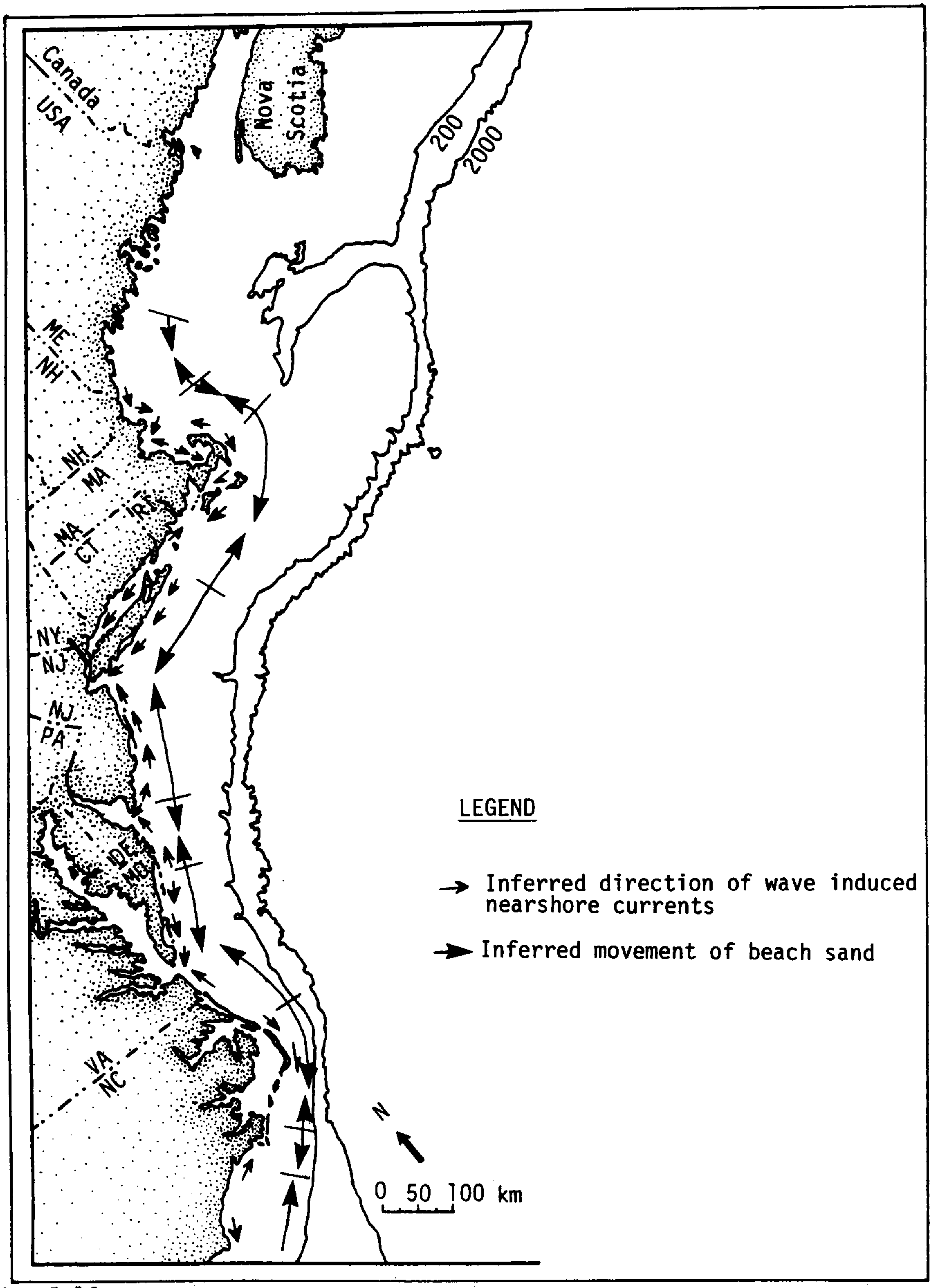


Fig. I-16 Direction of wave induced nearshore currents as deduced from shapes of beaches and movement of beach sand (after Emery and Uchupi 1972).

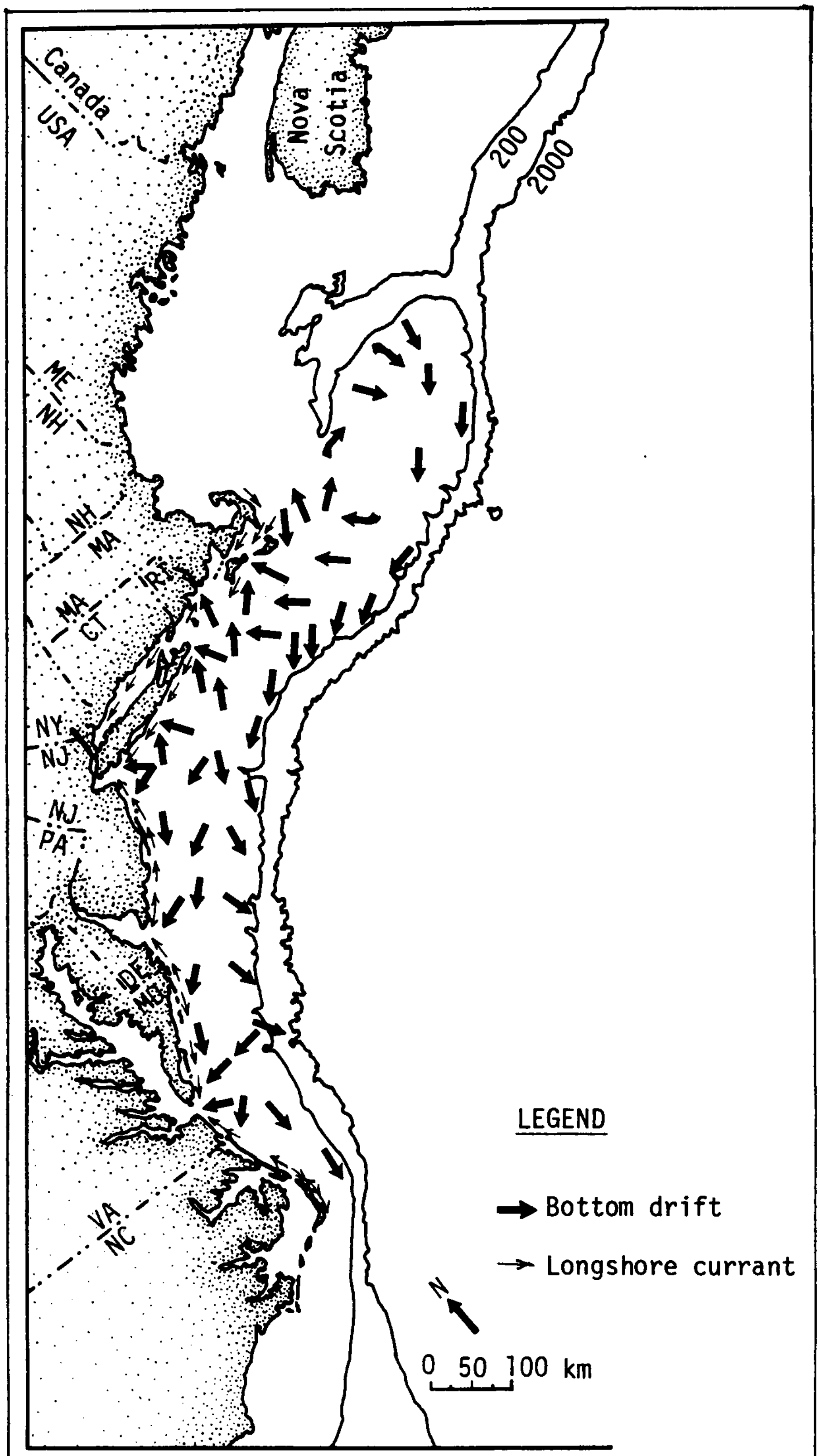


Fig. I-17
Direction of bottom drift on the shelf as derived from bottom drifters (after URI 1973).

organic material for dating is extremely critical. Redfield and Rubin (1962), for example, recognized that specific types of salt marsh grew within a very limited tidal range. The depth of similar peat materials below the present surface, plotted against the radiocarbon age of the peat, produces a sea-level curve.

Similar investigations of coastal deposits have been conducted in Virginia (Harrison and others 1965; Newman and Rusnak 1965), Delaware (Belknap and Kraft 1977; Kraft 1971; Kraft 1977; Sheridan and others 1974); New Jersey (Stuiver and Daddario 1963); Long Island (Newman 1977; Redfield 1967; Sanders and Kumar 1975a); Connecticut (Bloom and Stuiver 1963); Massachusetts (Kaye and Barghoorn 1964; McIntire and Morgan 1963; Refield 1965; Redfield and Rubin 1962); New Hampshire (Keene 1971); and Maine (BLoom 1963; Schnitker 1974; Stuiver and Borns 1975). The results of several of these studies are shown in Fig. I-18. As this figure illustrates, submergence curves between Virginia and northeastern Massachusetts do not differ by more than a few meters during the last 5,000 years. Unfortunately, submergence rates along all the Atlantic coast of North America do not form a consistent trend. Part of this may be due to the inclusion of erroneous data points. A large portion of the differences observed in this region during the last 5,000 years is probably due to local tectonics and subsistence (Dillon and Oldale 1978; Fairbridge and Newman 1968; Newman and March 1968).

It is evident from Fig. I-18 that nearshore salt marsh deposits rarely produce information on sea-level positions before 5,000 years ago. This is to be expected because sea levels rise progressively throughout time. Obviously, older salt marshes are now buried beneath marine sediments further offshore. These studies illustrate that the width of the lagoonal zone is a function of slope. The gentle slope of the coastal plain from New Jersey southward makes possible the development of extensive marshes, broad swamps, and broad estuaries along the coast. In most of New England, more steeply sloping uplands converge on the shoreline, creating numerous estuaries but restricting estuary width. There are local exceptions, of course, but these general differences tend to hold true when large sections of each region are compared.

Since today's coastal salt marsh deposits do not contain a full record of Holocene sea-level change, other approaches were developed in order to produce submergence curves for earlier time periods. Submerged peat deposits would offer one potential for extending sea-level curves back into the Late Pleistocene, if these deposits could be found at sufficient depth along the middle and outer Continental Shelves. To date, no systematic investigations have been conducted for the purpose of locating suitable submerged peat deposits, thus making possible the completion of Early Holocene-Late Pleistocene sea-level curves. Freshwater and salt-marsh peats have been recovered from the CS but usually as a result of trawling activities (Emery and others 1967). A substitute for marsh peat has been found in shell material, usually collected within the upper 100 cm, which can be used to construct sea-level curves. This method is undergoing considerable re-examination at the present (compare

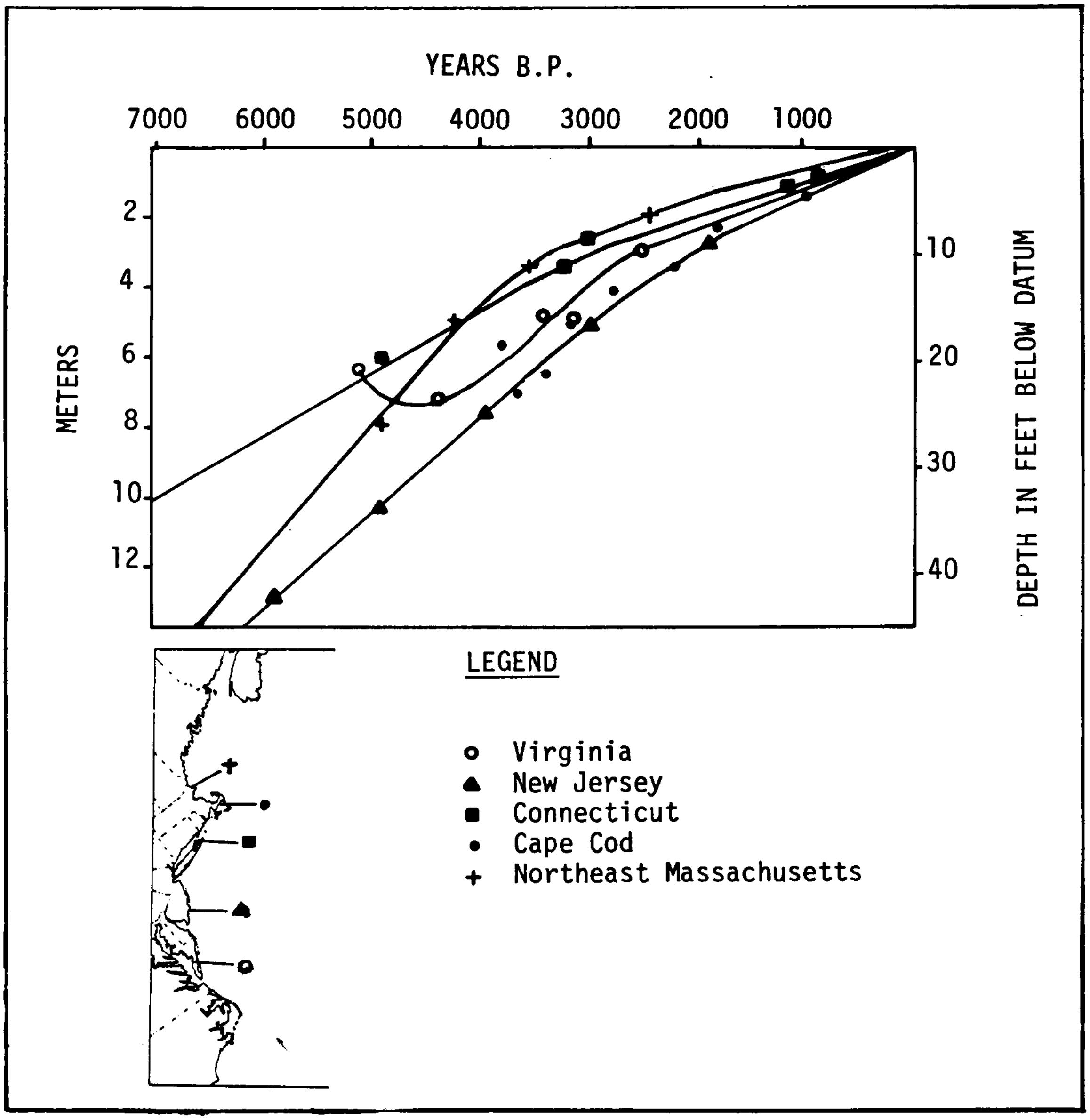


Fig. I-18
Submergence curves taken from marshes along the Atlantic Coast. Lines provide possible interpretation for post-sea level curves along the middle Atlantic Bight. Cape Cod region does not conform easily to a simple sea level curve.

Dillon and Oldale 1978; Emery and Merrill 1978; Macintyre and others 1978a, 1978b).

For over a decade, shells of specific lagoonal and intertidal species have been collected from the CS, dated using the radiocarbon method, and their depths plotted to construct sea-level curves (Emery and Garrison 1967; Macintyre and others 1978a; Merrill and others 1965; Milliman and Emery 1968). Most commonly used are shells of the oyster *Crassostrea virginica* (Gmelin).

The use of shell material from contexts within the "surficial sand sheet" of the CS to construct submergence curves has several drawbacks. First, it is most important that the shell fragment dated be from a species whose habitat is limited to a narrow range. This allows for a reasonably accurate reconstruction of the sea level *providing* the shell has not substantially moved from the original depth it inhabited. Unfortunately, few estuarine or intertidal shells have been found in undisturbed contexts. Considering nearshore dynamics and some aspects of the model presented earlier, lagoonal shells recovered in the surficial sand sheet may have been moved to depths 10 to 15 m lower (Sheridan and others 1974; Swift and others 1972) during the migration of the shoreline across the lagoon. Similar erosional transgression has been observed along sections of the coast between New Jersey and North Carolina.

On the shallow shelf just south of Cape Hatteras, recent work by Macintyre and others (1978a) has suggested that significant shoreward movement of shell material has occurred during the Holocene. Emery and Merrill (1978) agree with this position but suggest that such processes have had less effect on the deeper mid-shelf regions to the north.

The opinion that shells have been moved shoreward (and possibly upward) may hold true for the CS south of the Cape Hatteras as well. Fig. I-19 shows the distribution of pre-Holocene shells recovered from a portion of the Atlantic coast. If there is no sampling bias in the reported locations, the distribution of shell suggests that Holocene lagooonal and marsh sediments have been largely destroyed by erosion south of Cape Hatteras. The recovery of a considerable number of older (Late Pleistocene) shells indicates that the Holocene subaerial surface and some transgressive lagoonal-marsh deposits have been completely penetrated by erosion, which continued well into the underlying Late Pleistocene regressive deposits. North of Hatteras, few Late Pleistocene shells have been reported, which suggests that little erosion of the Late Pleistocene deposits has occurred. One factor that may be influencing this pattern is the difference in sediment load entering each region by river transport.

The use of shell material, especially *Crassostrea virginica*, for submergence curves would be much more helpful if a rigorous set of sampling conditions were met. In order to avoid collecting shells that have migrated significantly from their original position of growth, it would be most beneficial to use only shell material recovered below the

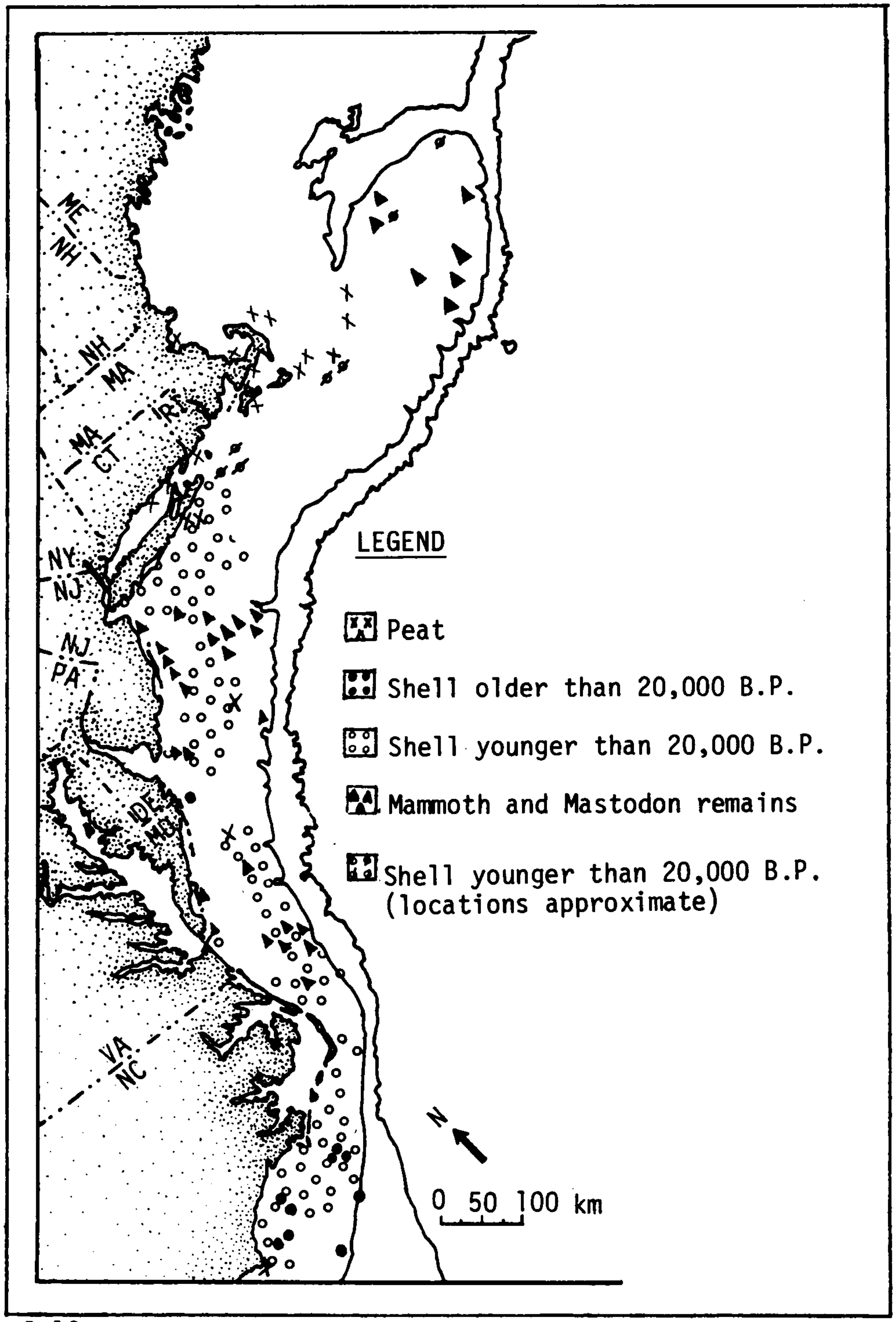


Fig. I-19

Distribution of some organic materials used to infer age-depth relationships along the Eastern Atlantic Coast. Distribution of organic materials taken from: Milliman and Emery (1968); Macintyre, et. al. (1978); Emery and Garrison (1967); Emery, Wigley, Bartlett, Rubin, and Barghoorn (1967); Whitmore, Emery, Cooke, and Swift (1967); Merrill, Emery, and Rubin (1965).

"Holocene surficial sand sheet." If shell material is only obtainable in the "Holocene surficial sands," then several shell-sampling programs should be undertaken when determining age-depth relationships. The earliest dates within a date cluster should probably be used to draw a submergence curve. If a wide scatter of dates is found, then all of the dates should be discarded. That is, if the range of shell dates obtained is greater than 2,000 years, then we should infer that shoreward transport or some other process has disrupted the age-depth relationship.

Besides shell, some investigators have used beachrock, total organic carboon, oolitic rock, and oolite to determine ancient sea levels (Milliman and Emery 1968). Once again, the critical issue rests upon both the depth of formation and possible post-growth movement which may significantly alter the interpreted sea-level values, The use of beachrock for radiocarbon dating for example, has received severe criticism from Macintyre and others (1975), and Allen and others (1969). Their research strongly suggests that northern varieties of beachrock lithify at sub-tidal depths. The research of Macintyre and others (1975) also shows that in their example the shells yielded dates substantially more recent than the associated cement, producing a meaningless "contaminated" date.

Several regional sea-level curves have been constructed for the eastern Atlantic shelf along the United States (Emery and Garrison 1967; Emery and Milliman 1971; Milliman and Emery 1968; Redfield 1967). Unfortunately, rather substantial depth-age differences occur throughout this region, and single curves, such as that proposed by Milliman and Emery (1968), have been shown to be inadequate on the basis of the work of Dillon and Oldale (1978) or Belknap and Kraft (1977). Isostatic readjustment and local uplift or subsidence have been used to explain the observed variability (Emery and Garrison 1967; Fairbridge and Newman 1968; Harrison and others 1965; Newman and March 1968). Other factos, such as the effect of water loading on the shape of the earth's geoid, also seem to play a part and give added complexity to the issue.

Eustatic curves for the Holocene are not particularly useful given the variability observed along the eastern coast, especially during the early Holocene. Relative sea-level curves and occasional dated material in good context are much more important to use than any single eustatic curve. Several dozen sources have been used to construct general sea-level curves along the project area. Fig. I-20 lists by state the sources used for making these curves. Chart I-1a shows the position of the shoreline at 3,000-year intervals during the Holocene. The shoreline positions shown on the map have been adjusted for transgressive shoreface erosion. This adjustment was necessary, since at least 10 m of nearshore deposits are eroded and releveled during landward translation of the shoreline, through the process of erosional shoreface retreat (Swift 1968). A 10 to 20 m scarp extends from the beach to a distance of several kilometers offshore today. This radical change in nearshore slope is a direct result of expended wave and nearshore current energy.

LEGEND

A. Gulf of Maine Region
Bloom 1963
Grant 1970
Harrison and Lyons 1963
Kaye and Barghoorn 1964
Keene 1971
McIntire and Morgan 1963
Schnitker 1974
Stuiver and Borns 1975

B. Cape Cod Region
Oldale and O'Hara 1979
Redfield 1967
Redfield and Rubin 1962

C. Long Island Region
Bloom and Stuiver 1963
Newman 1977
Redfield 1965

D. New Jersey Region
Meyerson 1972
Stuiver and Daddario 1963

E. Delmarva Region
Belknap and Kraft 1977
Harrison, Malloy, Rusnak and Terasmae 1965
Harrison and Rusnak 1962
Newman and Rusnak 1965
Newman and Munsart 1968

F. North Carolina Region
MacIntyre, Blackwelder, Land, and Stuckenrath 1975, 1978
Redfield 1967

Other general sea level studies for this section of the Continental Shelf
Dillon and Oldale 1978
Emery and Garrison 1967
Emery, Wigley, Bartlett, Rubin, and Barghoorn 1965
Merrill, Emery and Rubin 1965
Milliman and Emery 1968
Newman and March 1968

Fig. I-20

Listed by regions shown in the figure (right), these sources of data were used to construct local sea-level curves for sections of the project area.

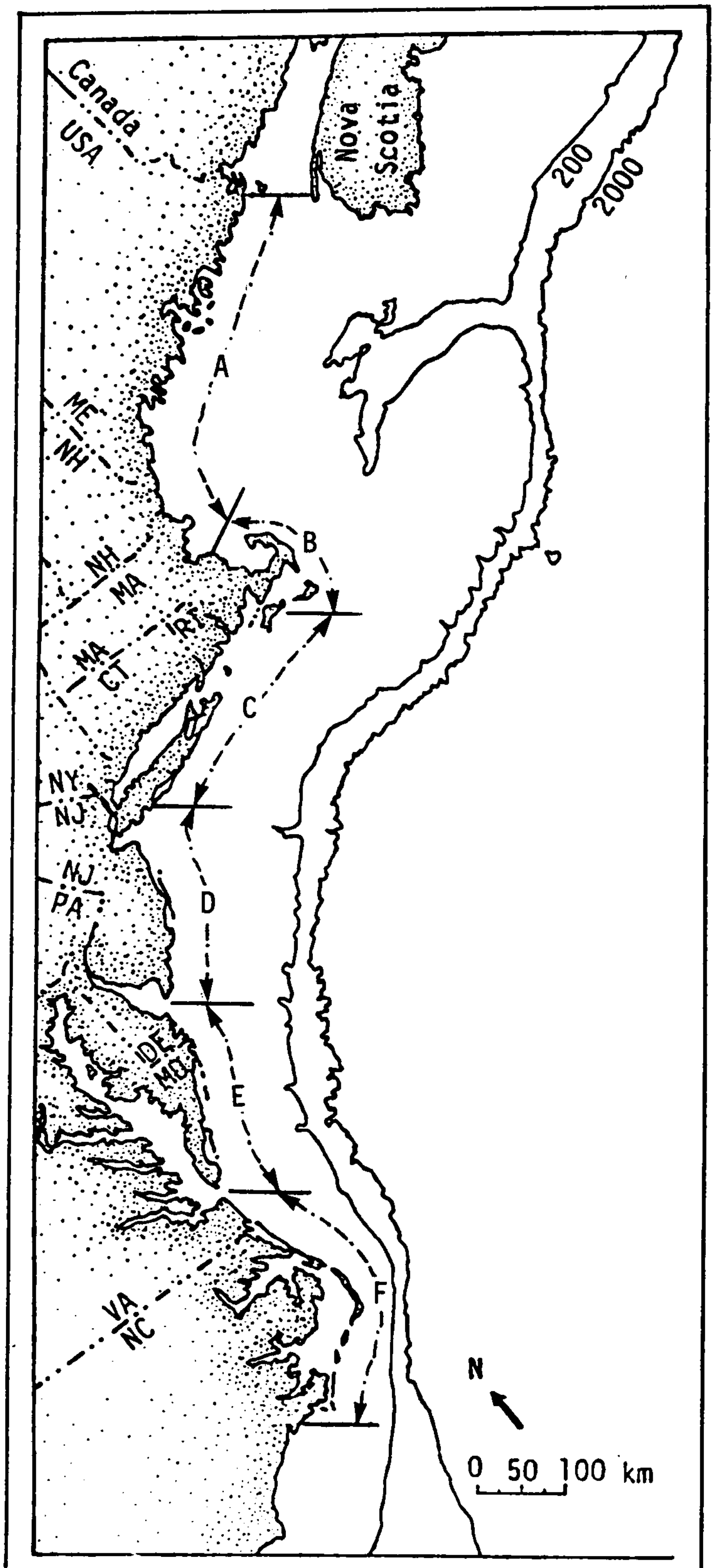


Fig. I-20 (continued)
Sources offering information on Holocene sea-level positions. (Data from some of these studies were used to construct Chart I-1a.)

Many problems exist in constructing ancient shoreline positions during the Holocene and Late Pleistocene. Little information is available for sea-level postions before 10,000 years ago, as pointed out by Emery and Merrill (1978) and Macintyre and others (1978a, 1978b).

The recovery of shell material 7,000-8,000 years old on North Carolina beaches (Macintyre and others 1978a, 1978b) has been used to support the landward transportation of shell material by waves and currents. An alternative explanation for these "fossil" shells on modern beaches is that they may be the result of nearshore scour of older deposits. Tidal inlet scour is quite capable of eroding through older lagoonal deposits. The relative sea-level curve for North Carolina (south of Cape Hatteras) indicates that former lagoon deposits should be 15 to 25 m below present sea level if lowland areas used to exist in this region. Tidal inlet scour to this depth is reasonable (Swift 1968; Kumar and Sanders 1975) and may also account for older shells found nearshore without recourse to significant landward migration of shell material. Scour 2 to 3 times the depth of the adjacent lagoon or shoreface is quite common for tidal inlets (see for example Kumar and Sanders 1975). If these shells are from locally eroded lagoon deposits rather than from mid-shelf features, they would indicate that rather large lagoons once extended inland for at least 15 to 25 km given the location of the shoreline between 7000 and 8000 B.P. shown on Chart I-1a. The lagoons in this area of North Carolina (Cape Lookout) today range from 5 to 10 km in width. The existence of larger lagoons is not improbable but would have required gentler slopes offshore and the formation of substantial barrier islands. The thinness of the "Holocene surficial sand sheet" found throughout this area may be indicative of barrier island migration by means of storm washover in a manner similar to that described by Dillon (1970) but most researchers advocate a different evolution for this section of the coast during the Holocene (see for example Swift and Sears 1974).

Many more systematic investigations need to be conducted before the reconstruction of shoreline positions shown on Chart I-1a can be regarded as absolutely accurate. Absent from these reconstructions are the shoreline locations for large bays and estuaries. Dozens of rather closely spaced transects would be necessary if we are to determine the extent of these estuaries, and that only if suitable deposits can be easily located beneath the "Holocene sand sheet."

5.0 GENERAL APPROACH TO THE RECONSTRUCTION OF MAJOR CONTINENTAL SHELF FEATURES SINCE THE LATE PLEISTOCENE

Before attempting to reconstruct the major landforms that once existed when portions of the CS were subaerial, it is important to recognize several aspects of shelf physiography. The most significant factor to consider is that the surface expression found on the CS today does not closely correlate with its pre-transgressive topography (Knebel and Spiker 1978; Kraft 1971; Sheridan and others 1974; Stubblefield and Swift 1976; Swift 1975b; Swift and others 1970; Swift and others 1974).

Almost all of the shelf topography has been formed by submarine hydraulic processes and not by subaerial processes during the Holocene. Sand ridges and sand waves are good examples of medium- to large-scale topographic features, ranging up to a dozen kilometers in length, that have been formed by nearshore and mid-shelf submarine processes. Sand waves and sand ridges are not always properly distinguished in the literature (see for example Emery and Uchupi 1972; Uchupi 1968, 1970). In this report, sand waves are defined as morphologic submarine features which develop transverse to the direction of flow. Sand ridges, on the other hand, are features which develop parallel or sub-parallel to the direction of flow. The distribution of sand ridges is given in Fig. I-21 (Uchupi 1968). Sand waves were found to be less frequent submarine features than sand ridges (USGS 1978).

The development of sand ridges is not adequately understood (Swift 1975a; Swift and others 1977), although they are no longer viewed as relict subaerial features. Research done by Stubblefield and others (1975), and Stubblefield and Swift (1976) indicates that even deeply submerged sand ridges still actively respond to certain hydraulic processes.

Since sand waves and ridges are formed by submarine processes, these features do not necessarily reflect the pre-transgressive subaerial topography. Their relief is great enough to obscure most pre-transgressive subaerial drainage patterns.

Other processes which rework the subaerial surface have been discussed in detail in the section describing our model on marine transgression. Extensive releveling and beveling of headland areas further complicate attempts to reconstruct the pre-transgressive subaerial surface. Longshore transport and estuarine sedimentation help to fill in many areas of low relief with the material eroded from coastal and submerged nearshore headlands. Submerged river valleys have formed effective sediment sinks as estuaries have retreated up their axes. The major coastal compartments in the project area, as outlined earlier, are separated by large estuary-retreat paths. In general, upland areas beside these estuaries have been eroded, and the material deposited in the estuaries

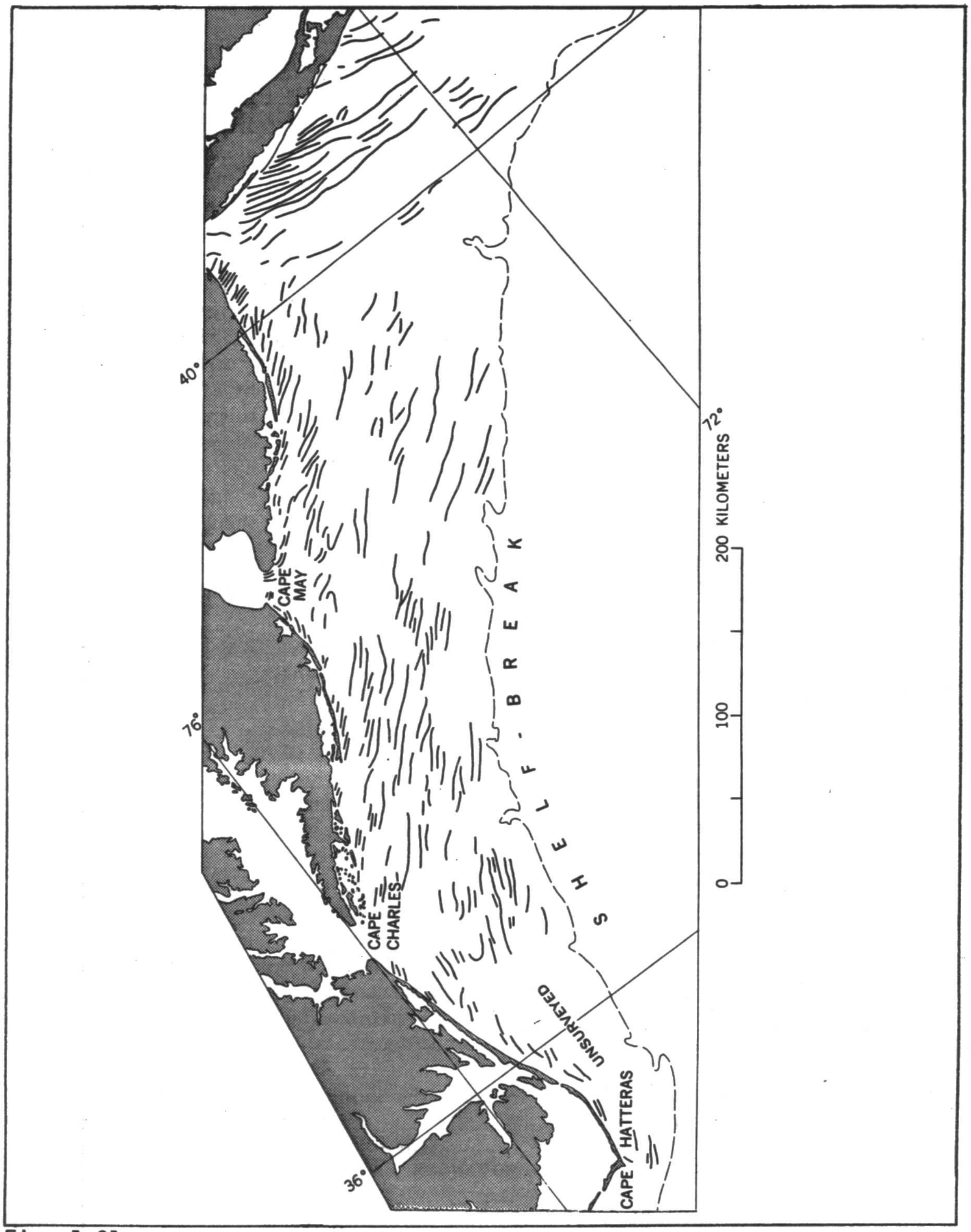


Fig. I-21 Distribution of sand swells on the continental shelf between Long Island and Cape Hatteras, North Carolina. Sand swells are formed by submarine hydraulic processes and do not provide information on the past subaerial environment of the continental shelf during periods of lower sea level. The age of these features on the inner shelf is much younger than those along the outer shelf. After Uchupi (1968).

and their associated shoal-retreat massifs or in the "surficial sand sheet" (Swift and others 1972).

Shoal-retreat massifs are important features developed during marine transgression (Swift 1975b; Swift and others 1972). Massifs are formed as the result of longshore currents interacting with estuarine currents. The disruption of these currents and their combined interaction produces a depositional center for sediment. As the shore migrates landward in response to rising sea level, major depositional centers also move landward, creating the 3-dimensional features designated by Swift (1973) and Swift and others (1978) as shoal-retreat massifs. Shoal-retreat massifs represent yet another example in which modern shelf physiography does not represent the pre-transgressive subaerial topography.

The cumulative effect of erosional shoreface retreat, headland beveling, lowland infilling, and the development of shoal-retreat massifs, estuary retreat paths, sand ridges, and sand waves has substantially altered the pre-transgressive subaerial surface. Evidence of their cumulative effect comes from detailed seismic profiles and cores. The locations of buried pre-Late Wisconsin channels, flood plains and river valley deposits of the Delaware and Great Egg Rivers diverge substantially from their submerged shelf valley thalwegs (Kraft 1977; Kraft and others 1978; Sheridan and others 1974; Stubblefield and Swift 1976; Swift 1973, 1975a; Swift and others 1977; Swift and Sears 1974). These problems notwithstanding, the following pages review research done to date on important aspects of the gology of the Continental Shelf. As mentioned previoussly, only those aspects of CS geology which aid in the identification and interpretation of the distribution of archaeological resources are covered. The discussion is organized by coastal compartments and their seaward counterparts. This type of organization was selected as one way to group the data collected for each portion of the project area. It may be advantageous for this particular project in that drainage systems and major drainage divides often correspond to similar boundary zones used by archaeologists to separate different "culture areas."

6.0 NORTHERN NORTH CAROLINA - SOUTHEASTERN VIRGINIA CONTINENTAL SHELF

The boundaries for the northern North Carolina- southeastern Virginia Continental Shelf have been somewhat arbitrarily selected. Cape Hatteras forms the southern boundary and from it a line has been projected seaward to meet the continental slope. The northern boundary for this compartment is formed by the approximate position of a drainage divide which separated the ancestral James River from the ancestral Susquehanna River.

Some bathymetry of this portion of the CS is shown on Fig. I-22. In general, relief of the Shelf is gentle and some sections of the Shelf contain linear ridges up to 10 m in height. These ridges result from nearshore and mid-shelf submarine hydraulic processes as discussed by Swift (1975b), and Swift and others (1978).

The major morphological features (Fig. I-22) on the Continental Shelf between Cape Hatteras, NC and Cape Henry, VA are 2 shelf-valley complexes (Swift and others 1978) and the cuspate foreland at Cape Hatteras (Shideler and Swift 1972; Swift 1973; Swift and Sears 1974; Swift and others 1978). The present coast consists of a major barrier chain. The chain is separated from the mainland by a prominent lagoonal system made up of Pamlico and Albemarle Sounds (Fig. I-22). Oregon Inlet connects these sounds to the Atlantic Ocean.

The evolution of this section of the Continental Shelf during the last 20,000 years is not well known, although some research on the subject has been attempted (Macintyre and others 1975; Pierce and Colquhoun 1970; Shideler and Swift 1972; Swift and others 1977; Swift and others 1978; White 1978). These investigations point out the complex evolution that this area experienced during the Holocene and Late Pleistocene.

The probability of intact Holocene-Late Pleistocene subaerial surfaces being preserved on the CS between Cape Hatteras and Cape Henry is lower than it would be in most sections of the Atlantic Continental Shelf to the north. Shoreface retreat during the last marine transgression apparently has eroded and redistributed a considerable amount of soil from the subaerial surface that once covered the uplands and interfluves between Cape Hatteras and Cape Henry. Widespread loss of the pre-transgressive subaerial surface may also be due to the more intense wave climate found along this portion of the Atlantic coast. Last of all, entrenchment of the valleys in this area during the Pleistocene has inhibited the formation of extensive undissected sections of coastal plain (White 1978). Unlike the flat coastal plain south of Cape Fear, NC, the Cape Hatteras - Cape Charles coastal plain has undergone significant fluvial erosion from tributaries of the Susquehanna River, an important Pleistocene meltwater river system (White 1978).

The near-surface stratigraphy of the northern North Carolina - southeastern Virginia Shelf has been studied by Shideler and Swift (1972). In general, post-Miocene sediments in this area average 47 m in thickness and represent several periods of deposition. The deepest post-Miocene materials are thought to represent coastal and marine deposits laid down during a pre-Wisconsin glacial cycle. These units together average 20 m in thickness and make up most of the post-Miocene strata found in this region.

Overlying these pre-Wisconsin deposits are Late Wisconsin sediments which range from 0 to 8 m in thickness and average about 3.2 m. These sediments display relatively uniform stratification and thin out in a westerly direction. The deposits are interpreted as representing widespread Late Wisconsin deltaic and nearshore marine environments which developed during the marine regression following the mid-Wisconsin interstadial. Shell materials recovered from some of these layers have been dated to between 20,000 and 24,000 B.P. The sediment associated with the shells is obscurely mottled massive grayish-green mud. The deposits are discontinuous and are absent from some cores obtained from this area. Mollusc species recovered are generally small forms with large surface area per weight and which are therefore adapted to living in very soft muds. The total molluscan assemblage is euryhaline, possibly representing estuarine, lagoonal, or deltaic environments. The presence of zones of homogeneous watery mud, interbedded mud and fine sands, and shallow euryhaline fauna, likewise suggests that deposits along sections of this compartment represent restricted environments such as deltas or lagoons.

The youngest strata in this region are Holocene in age and are composed of coarse to medium sands. They represent the Holocene "transgressive sand sheet" and were deposited sometime during landward migration of the shoreline and are subsequently shaped by marine processes. The "sand sheet" ranges in thickness from 0 to 9 m and averages 3 m. Coarse sands and some gravels appear in the troughs between ridges and probably represent "lag deposits." Some fauna recovered from the sand sheet were reworked from the underlying Late Wisconsin strata, implying erosion of that surface either during or after transgression.

Of the above 3 units (i.e. pre-Wisconsin, Late Wisconsin, and Holocene), the Late Wisconsin unit is most important to archaeological studies. The unit was laid down by streams and lagoons during Late Wisconsin times. On this unit, subaerial surfaces developed between 36,000 and 18,000 B.P. depending upon elevation and distance from the continental slope. Subaerial surfaces were formed once sea-level lowering had exposed a section of the Shelf to subaerial processes of weathering (for example physical, biological and chemical processes leading to soil formation; see Basile 1971). The upper section of the Late Wisconsin unit, when found intact, has the greatest likelihood of containing pre-transgressive subaerial surface. Unfortunately, it seems, on the basis of the few available radiocarb on dates, that much of this unit has been eroded or reworked subsequent to transgression (Shideler and Swift 1972; Shideler and others 1972; Swift and Sears 1974).

LEGEND Major Late Pleistocene-Holocene shelf features

••••• Buried valley on the shelf identified by seismic profiling and containing flood plain, river channel, river terrace and estuarine deposits

▬ Surface valley on shelf indicating that buried estuarine and possibly river valley deposits probably exist within the general vicinity

oooo Inferred buried valley approximately located and containing flood plain, river channel, river terrace and estuarine deposits

Major shelf delta identified by bathymetry, morphology, and sometimes seismic data when available

Shoal retreat massif (shelf highs of constructional origin due to littoral drift deposition centers)

Buried river valley deposits with location questionable

Scarp

Fig. I-22
Major Late Pleistocene-Holocene features on the northern North Carolina-southeastern Virginia shelf.

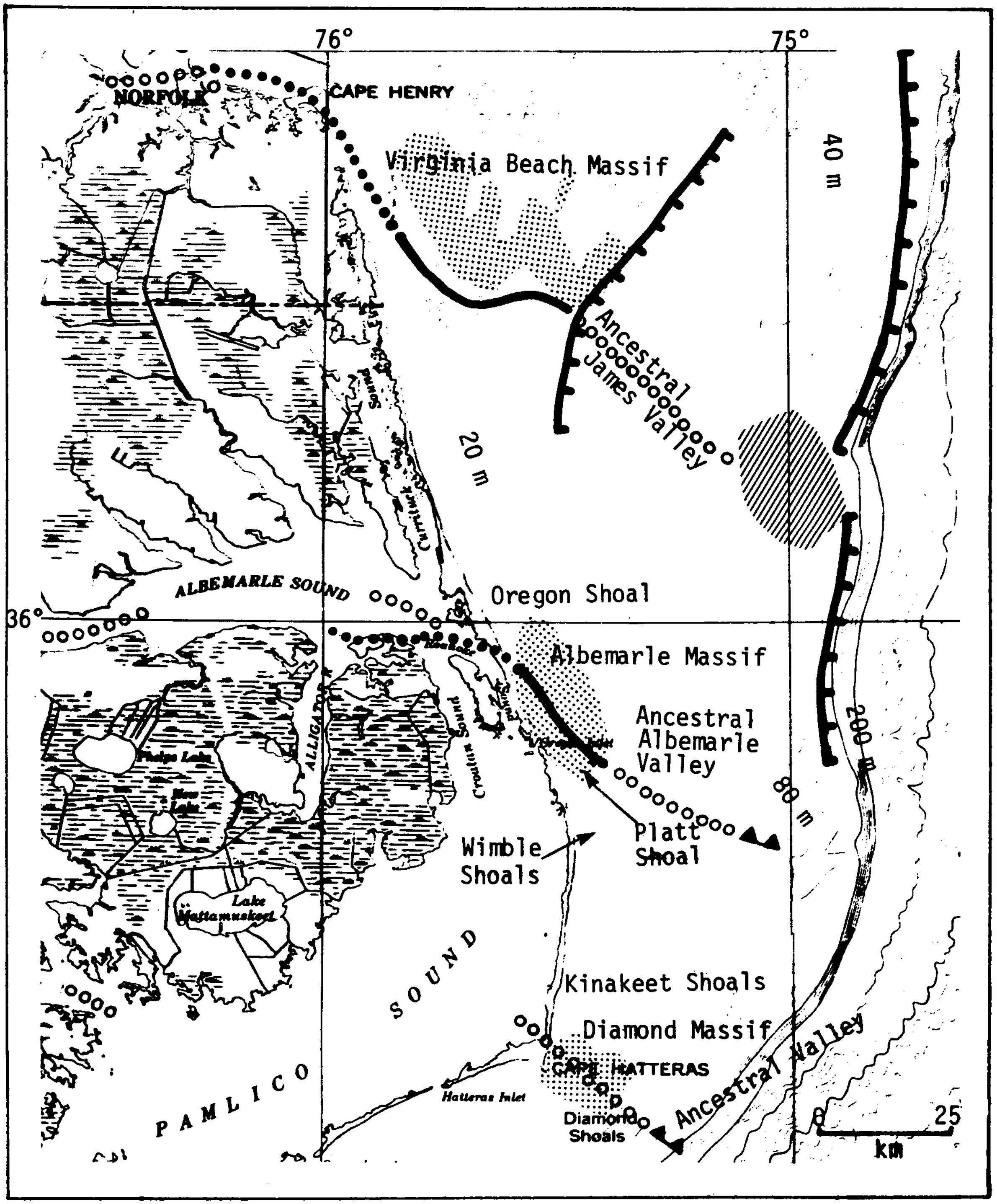


Fig. I-22

Major Late Pleistocene-Holocene features on the northern North Carolina-southeastern Virginia shelf. Features compiled from Goldsmith (1974); O'Connor and others (1972); Pierce and Colquhoun (1970); Shideler and others (1972, 1973); Shideler and Swift (1972); Swift (1975a); Swift and Sears (1974); Swift and others (1972); Swift and others (1977); Swift and others (1978).

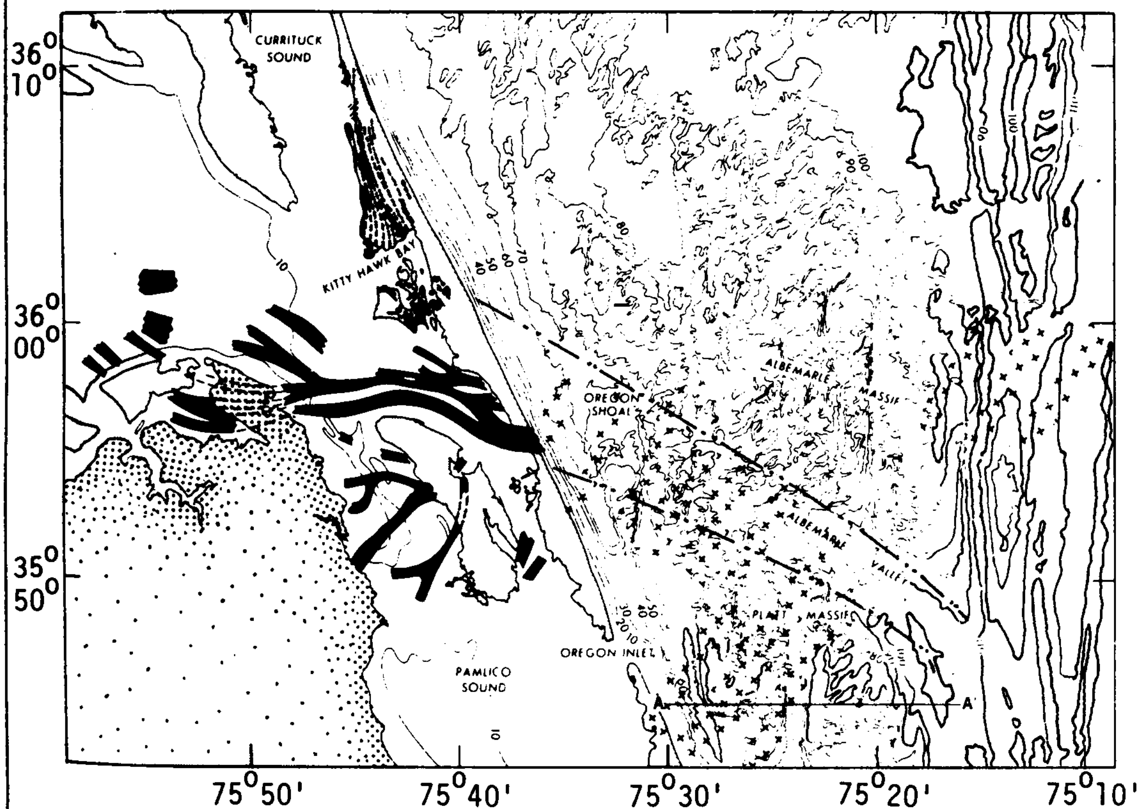
More recently Swift and others (1978) have reported finding lagoonal deposits in the nearshore region between Pamlico and Albemarle Sounds, adjacent to Oregon Inlet. These deposits date around 10,000 B.P. and can be associated with the most recent transgression. They are found in conjunction with complex deposits forming the Platt Shoals retreat massif on the south side of the ancestral Albemarle Valley (Fig. I-23). The landward extension of this valley has been traced into Albemarle Sound by seismic profiling (Swift and Sears 1974). The area of Platt Shoals has been interpreted as an ancient estuary mouth dating between 10,000 and 5000 B.P. (Swift and Sears 1974). At that time, the area may have been similar morphologically to the estuaries of the present-day Georgia coast, consisting of lobate delta configurations bordered by marshes.

At the southern end of the northern North Carolina - southeastern Virginia Shelf compartment is the cuspate foreland forming Cape Hatteras. Two somewhat different origins for Cape Hatteras have been suggested. Some researchers (Shideler and Swift 1972; Shideler and others 1973; Swift 1975; Swift and others 1978) consider Cape Hatteras to have evolved during the submergence stages of a delta probably belonging to the ancestral Pamlico River (submerged delta hypothesis). Fig. I-24 illustrates this view and shows several stages in the evolution of a cuspate foreland. Similar processes of mainland-beach detachment are considered to have operated along portions of the coast south of Cape Hatteras during the Early and Middle Holocene (Swift and Sears 1974).

A slightly different viewpoint regarding the formation of Cape Hatteras is given by Pierce and Colquhoun (1970). They suggest that Cape Hatteras evolved by coastwise spit progradation from an eroding headland area just north of it as illustrated in Fig. I-25 (spit progradation hypothesis).

At present there are not enough data available to determine which hypothesis is more accurate, although the modern littoral drift pattern contradicts the spit progradation hypothesis (Swift personal communication; see Swift 1975a). The 2 processes may have worked somewhat together to produce the landforms seen today. However, what is of importance to archaeological studies is the question of whether intact Late Quaternary subaerial surfaces are preserved in this region. According to both reconstructions, uplands would have existed north of Cape Hatteras, while the area around the Cape would have consisted of lowlands.

Under the submerged delta hypothesis, uplands would form a normal constituent of the topography flanking the ancestral Pamlico River valley. Uplands are also an important part of the spit progradion hypothesis. The ancestral Albemarle Valley north of this area (Fig. I-22) forms the next adjacent drainage system. Lying somewhere between the ancestral Pamlico and ancestral Albemarle Valleys would have been an upland region. Within this upland region there would have existed a poorly defined drainage divide separating the two valleys. Given the topography found along the coastal plain today, a broad, relatively flat upland surface probably once existed and subsequently evolved into broad marshlands as



LEGEND

Corridor for Albemarle Shelf Valley

Subsurface channels

Location of prominent Holocene-Pleistocene contact observed on seismic profiles (Shideler and Swift 1972).

Fig. I-23

Buried channels and inferred buried valley in the vicinity of Albemarle Sound, North Carolina (adapted from Swift 1975A). Location of prominent Holocene-Pleistocene contact based on Shideler and Swift (1972). Contours in feet.

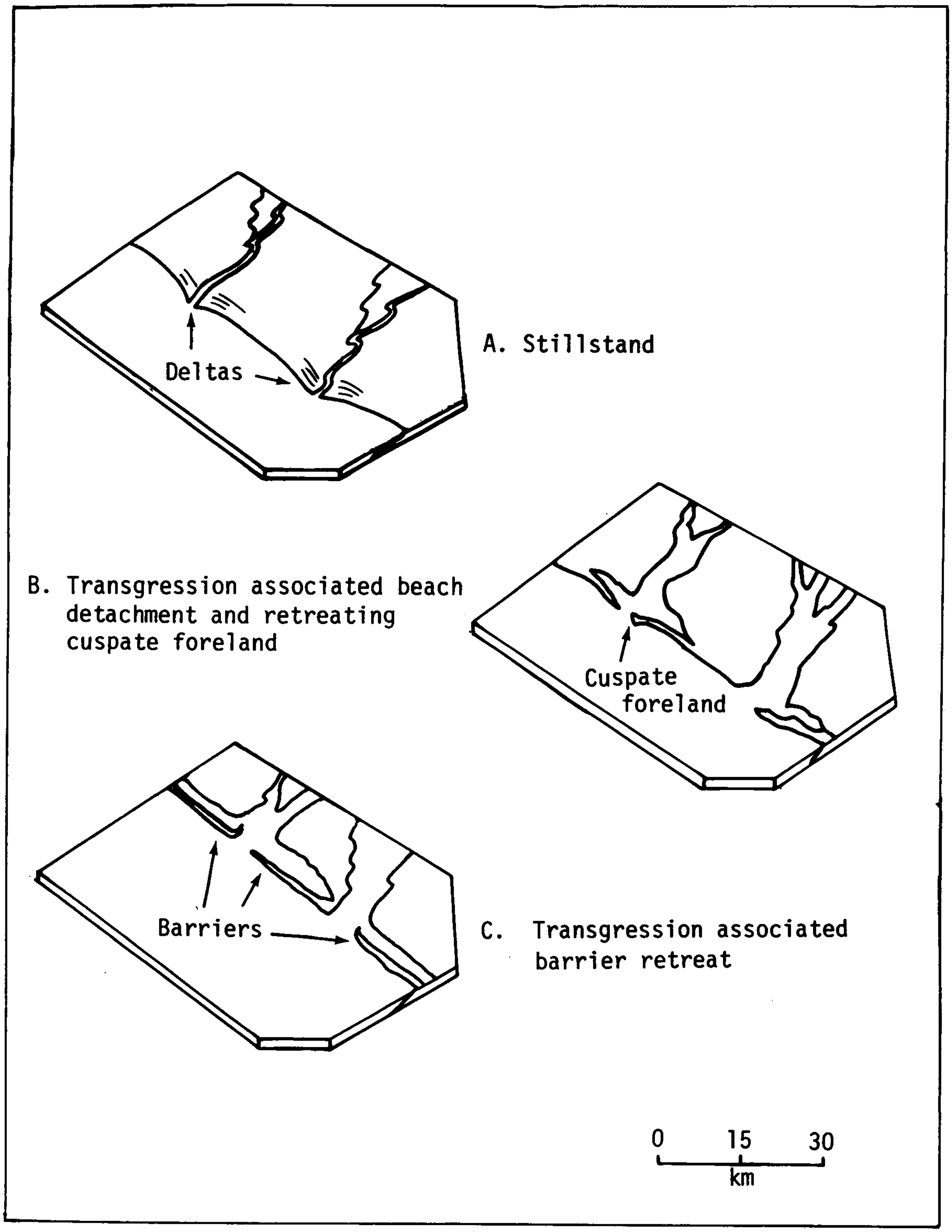


Fig. I-24

Model for the development of a cuspate foreland from a still-stand delta along a coast with low relief (adapted from Swift 1975a). Evolution of the shoreface of a low relief coastal region as it passes from a stillstand (A.) to a retreating coastline due to marine transgression (B. and C.).

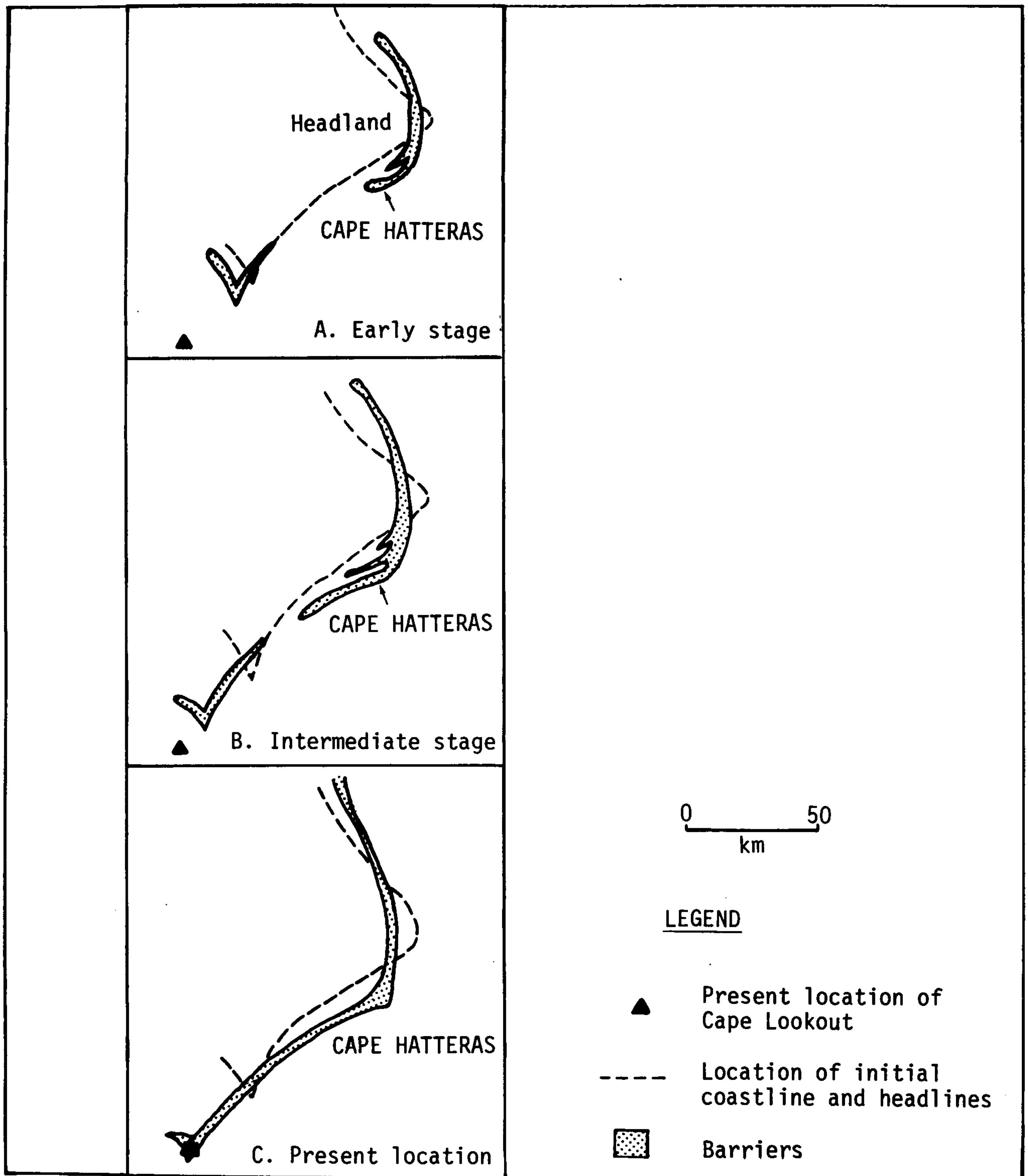


Fig. I-25 Inferred evolution of Cape Hatteras through coastwise spit progradation from eroding headlands (adapted from Pierce and Colquhoun 1970). Schematic of one hypothesis for the development of Cape Hatteras. Barriers form from headland erosion in early stage (A.) Coastwise spit progradation occurs during the intermediate stage (B.) finally reaching the present coastal configuration (C.).

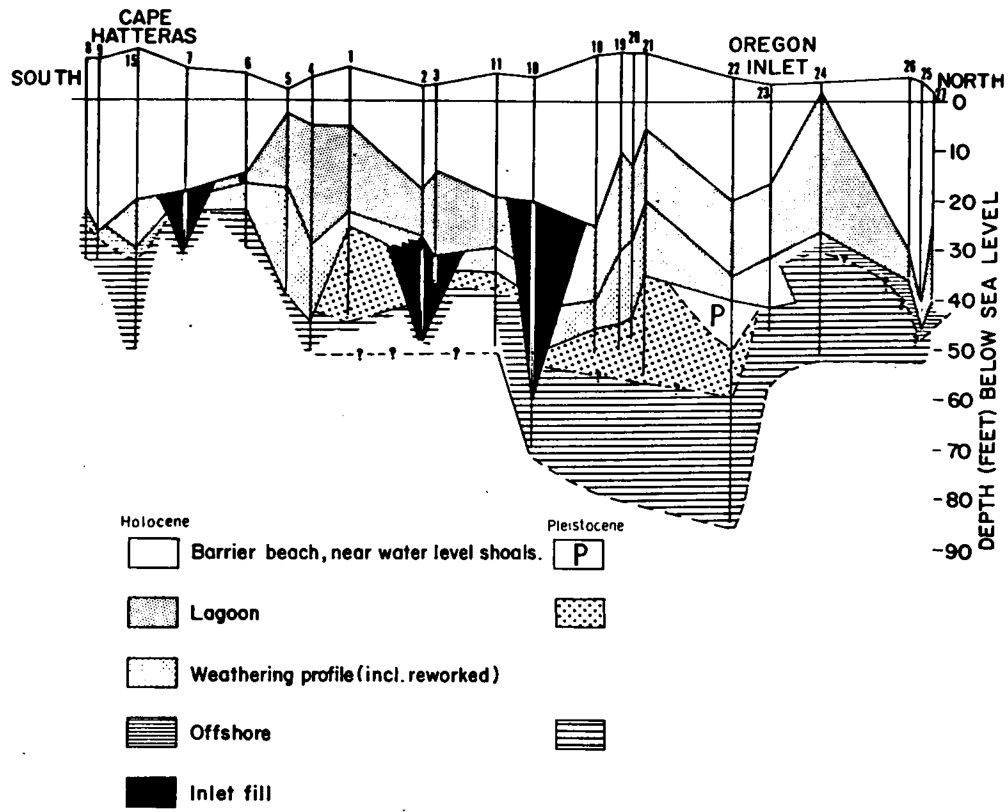
sea level rose. Since this area was by definition higher than adjacent regions, marsh and lagoon sediments would not have had the opportunity to become as thick as in the adjacent valleys. As marine transgression and landward migration of barrier beaches occurred, shallowly buried subaerial surfaces did not have as great an opportunity for preservation. Evidence in support of this can be found in the work done by Pierce and Colquhoun (1970), Shideler and Swift (1972), Swift and Sears (1974), Swift and others (1977) and Swift and others (1978).

Erosional shoreface retreat has probably eroded and redistributed much of the Late Quaternary subaerial surface and many of the lagoonal sequences which used to lie beneath Hatteras Island. Pierce and Colquhoun (1970) found what they interpreted as a substantial soil horizon underneath barrier and lagoon sequences of Hatteras Island and Currituck Spit. A profile made from their borings is shown in Fig. I-26.

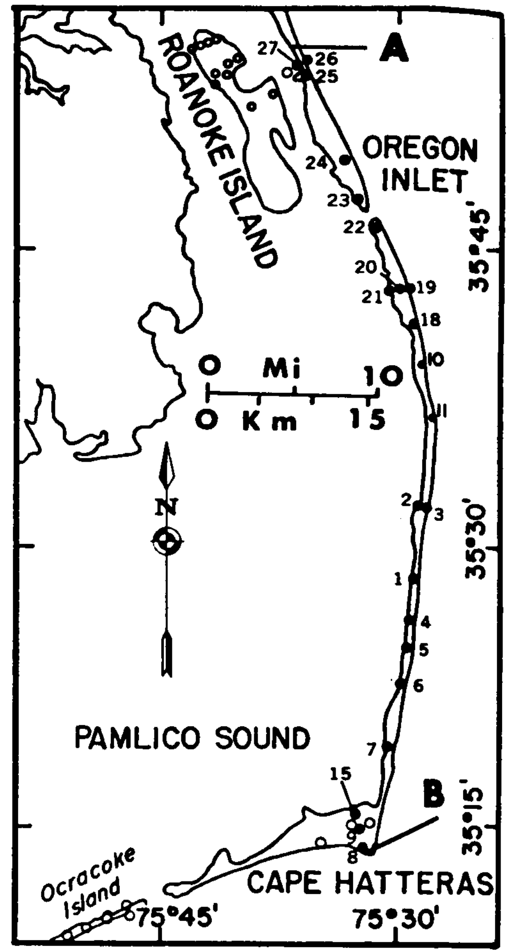
Investigations conducted by Shideler and Swift (1972) nearshore and into deeper water along this portion of the coast usually found Holocene transgressive submarine and nearshore deposits lying unconformably over Late Wisconsin and older sequences as outlined earlier. The absence of lagoonal sequences underlying much of the transgressive marine sands in this area has been recognized by several other investigations (Shideler and others 1972 and 1973; Swift and others 1977 and 1978). Apparently, shoreface retreat has eliminated most of the pre-transgressive subaerial surface with the exception of that lying within major estuary retreat paths (for example, ancestral Albemarle and James Valleys).

Fig I-27 shows the location of the seismic profiles discussed by Shideler and Swift (1972). This figure also shows those locations where the Holocene-Pleistocene contact was most visible in the seismic profiles. In the remaining areas, it was either absent (previously eroded), coincident with the present sea floor, or obscured by seismic noise in the upper few meters of each profile. The distribution of this reflector seems to cluster in those areas considered to be former shelf valleys (such as the ancestral Pamlico, ancestral Albemarle, and ancestral James). Its absence in the areas between these ancestral river valleys may be due in part to complete or near-complete erosion of the pre-transgressive subaerial surface. The model described at the beginning of this report calls for a substantial sediment covering over the subaerial surface (at least 5 or 10 m of marsh, lagoonal, estuarine, or flood plain sediments) to preserve it from transgression and subsequent submarine hydraulic processes. The areas mentioned above have a low potential for intact pre-transgressive subaerial surfaces since they would have been upland regions adjacent to major drainage systems. Several field checks are needed to verify this interpretation.

Additional investigations in the northern North Carolina - southeastern Virginia compartments have focused on mapping portions of the ancestral Albemarle and James River Valleys. Fig. I-23 shows the location of subsurface channels found by O'Connor and others (1972) within the buried Albemarle Valley west of Currituck Spit. Swift (1975a) shows the



Cross section along the barrier chain. Hole locations shown on adjacent figure. Vertical exaggeration: 2000X



Location of bore holes on the barrier chain. Black dots, holes shown on the cross section in adjacent figure; open circles, additional holes drilled, but not included on any cross section

Fig. I-26 Location and results of field investigation conducted by Pierce and Colquhoun (after Pierce and Colquhoun 1970).

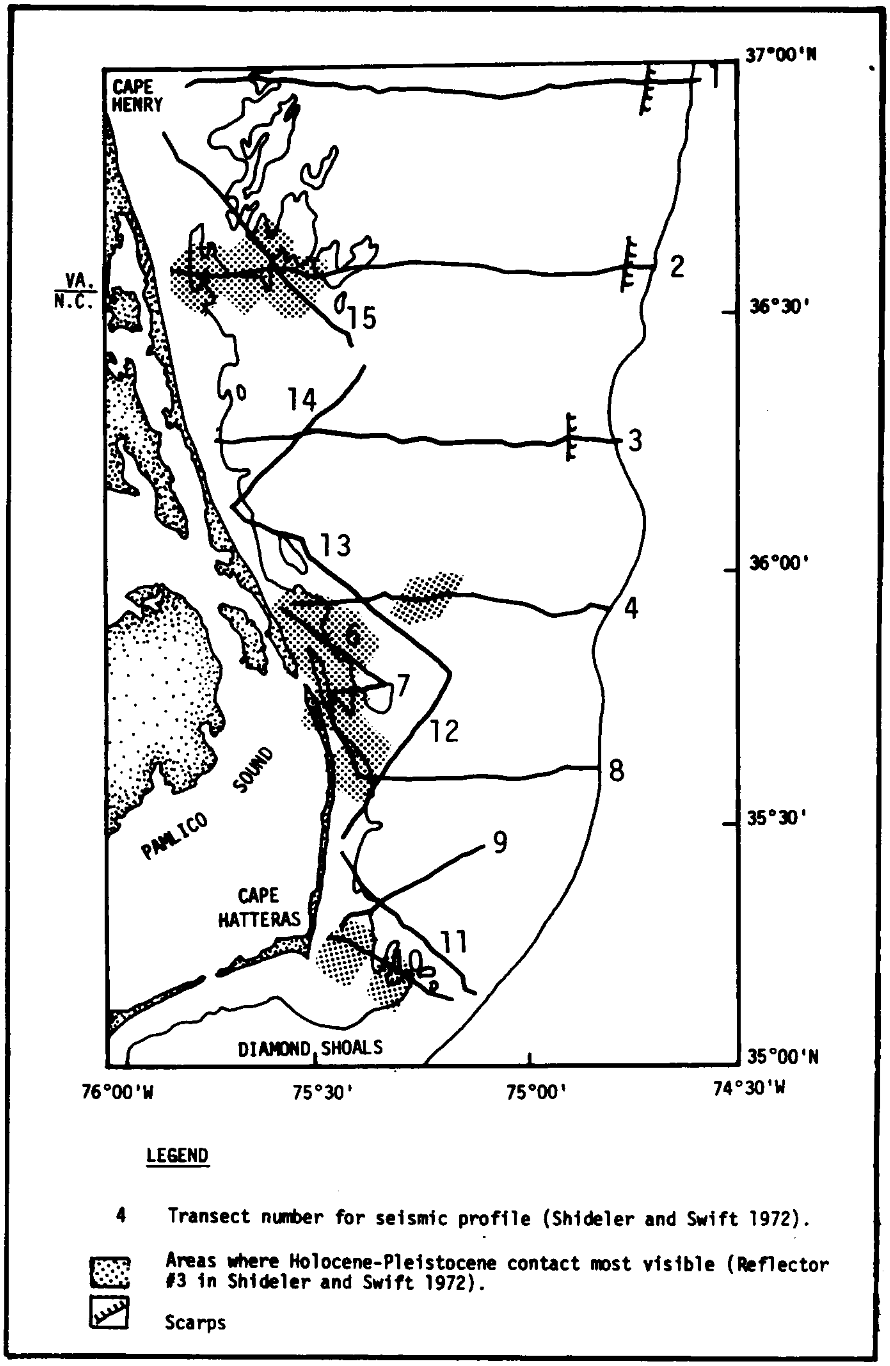


Fig. I-27

Location of seismic profiles discussed by Shideler and Swift (1972). Areas where Holocene-Pleistocene contact most evident shown by dot screen (adapted from Shideler and Swift 1972).

projected position of the central portion of the buried Albemarle River Valley on the CS (Fig. I-22 and I-23) and the shelf valley which is impressed into the estuarine sediments filling the old subaerial valley. As mentioned previously, Shideler and Swift (1972) observed a strong Holocene-Pleistocene reflector in the vicinity of the Albemarle Shelf Valley (Fig. I-26).

The next major drainage system north of the Albemarle Valley during the Early to Middle Holocene was the ancestral James River. Sometime after the Early Holocene, the James River Valley was captured by the larger Chesapeake Bay system, as each estuary increased proportionately in size with rising sea level. Before this event, a drainage divide separated the ancestral Susquehanna and James River Valleys. This divide probably consisted of a narrow but somewhat flat zone dissected by smaller rivers and streams which had become entrenched during the Late Wisconsin lowstand.

The position of the central portion of the ancestral James Valley on the Shelf has been mapped by Shideler and others (1973), Swift and Sears (1974) and Swift and others (1977). The ancestral James Valley is also referred to in the literature as the Virginia Beach Valley (see for example Shideler and others 1973; Swift and others 1977). Fig I-28 shows the location of the ancestral James Valley as identified on the Inner Shelf from seismic profiles. Swift (1975a) and Swift and others (1978) have extended this valley along the middle and outer Shelf on the basis of negative topographic relief (Fig. I-22).

The ancestral James Valley passes beneath Cape Henry, VA as illustrated in Fig. I-29 (Shideler and others 1973). The position of Cape Henry has not remained fixed during the Holocene but has slowly migrated as the mouth of Chesapeake Bay moved inland with rising sea level. The Cape's configuration has also evolved over time and it may have enclosed much larger bays during some periods (Kraft and others 1978). Fig. I-30 illustrates the shape of the spit today and shows the location of the contact between the Late Holocene dune and spit sands and the much older uplands of Late Pleistocene origin (about 32,000-40,000 B.P.).

Several scarps have been tentatively identified within this compartment by Swift and Sears (1974). Upon closer inspection of the bathymetry in this area (Fig. I-28a) the "scarp" north of the ancestral James Valley (Fig. I-22) may actually represent part of an incised stream valley flowing southwest and later captured by the ancestral James River. Chart I-la shows the position of old shorelines within this region of the Shelf, following the recent research of Dillon and Oldale (1978). Their investigations identified scarps at depths of about 42 m (Atlantis or Middle Shelf Shore) and 90 m (Franklin Shore). The Atlantis Shore dates sometime between 9000 and 13,000 B.P. The Franklin Shore may be as old as 15,000 B.P. (see Macintyre and others 1978a and b, and compare with Milliman and Emery 1968 and Dillon and Oldale 1978).

Investigations concerned with Late Quaternary sea-level change along the North Carolina - Virginia coast have provided little information

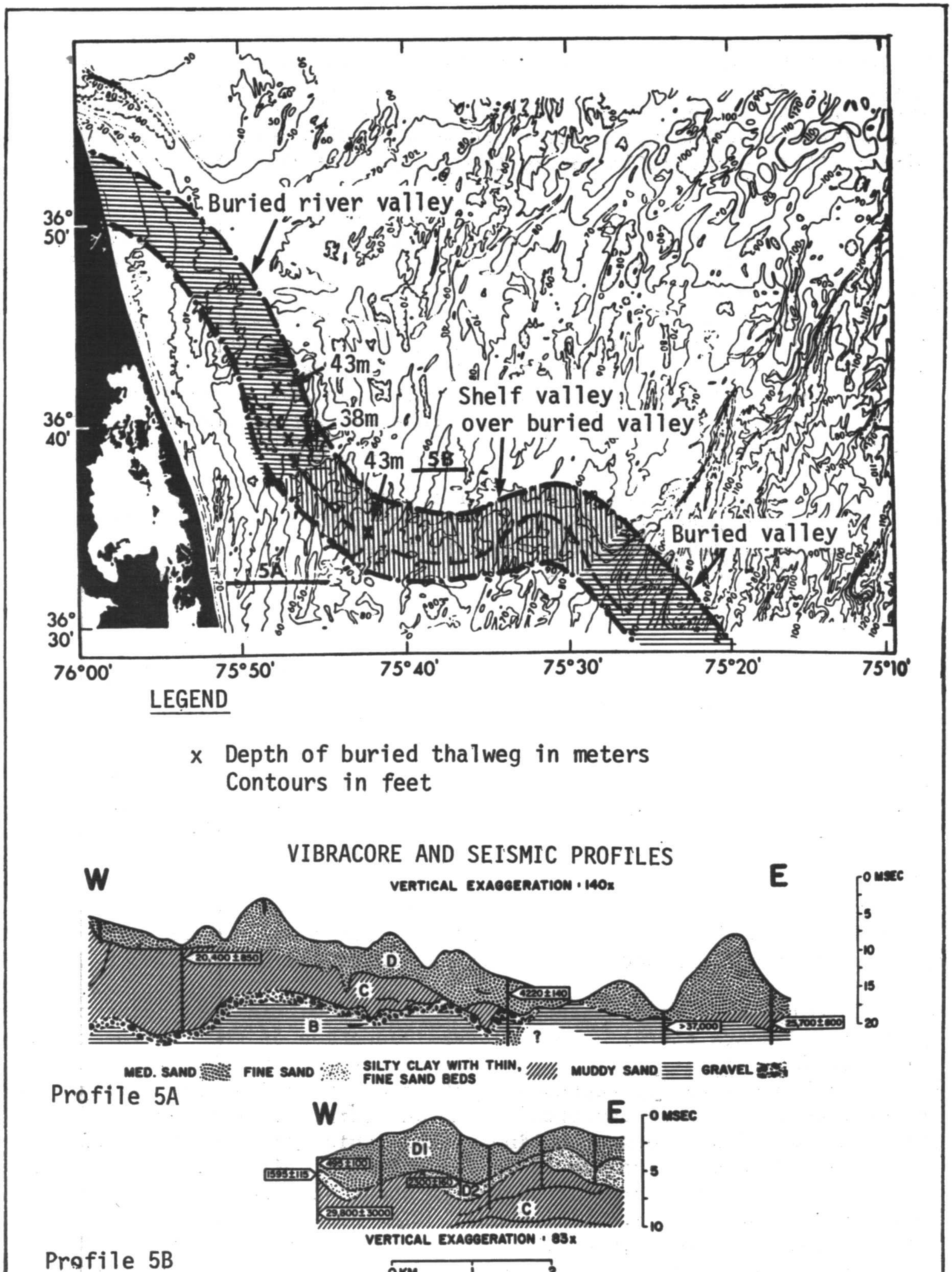


Fig. I-28 Approximate path of the partially buried James River Valley and two vibracore/seismic profiles in this vicinity (after Swift and others 1977).

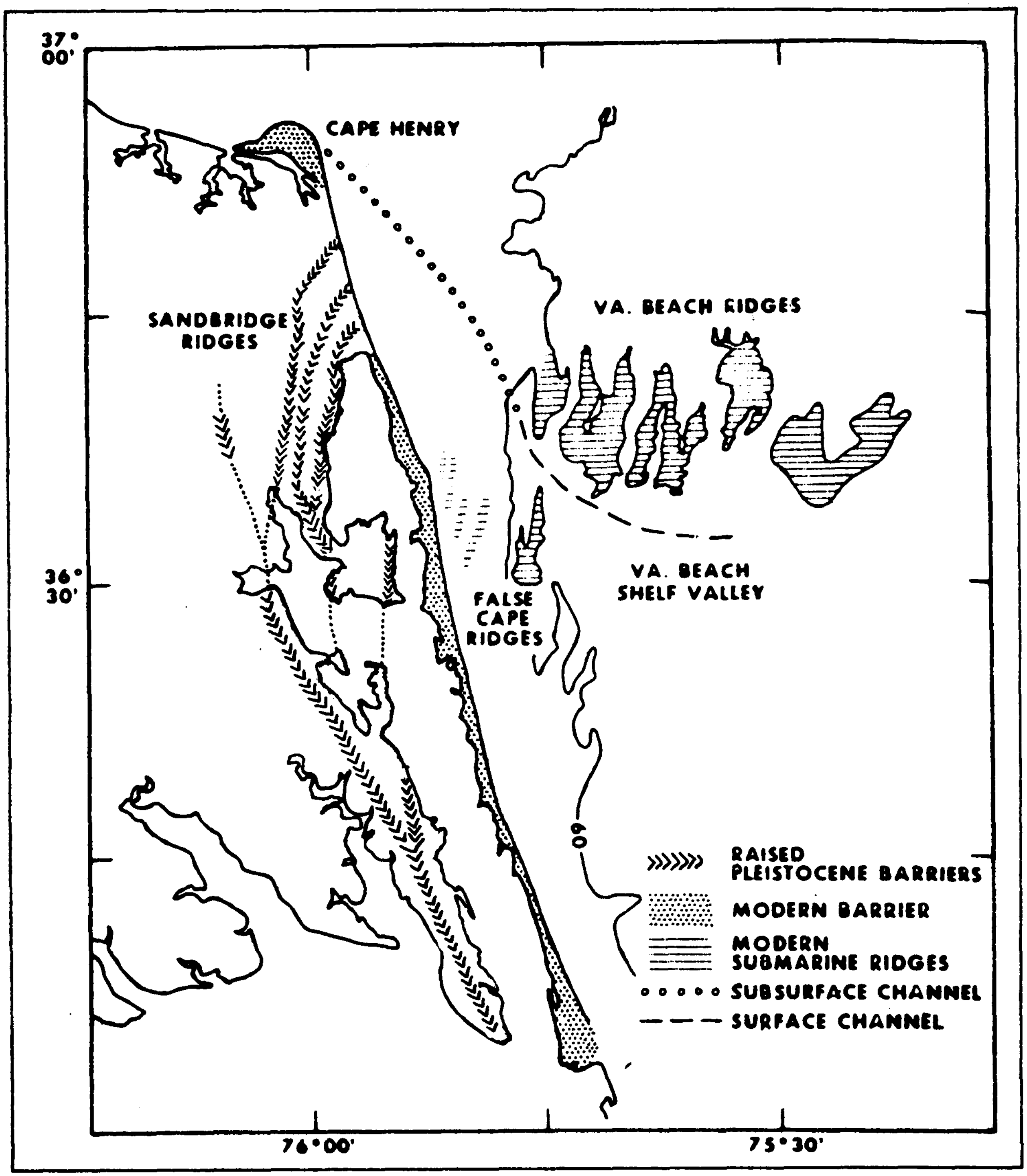


Fig. I-29
Major morphological elements of the Virginia coastal plain and shelf southeast of Cape Henry (after Shideler and others 1973).

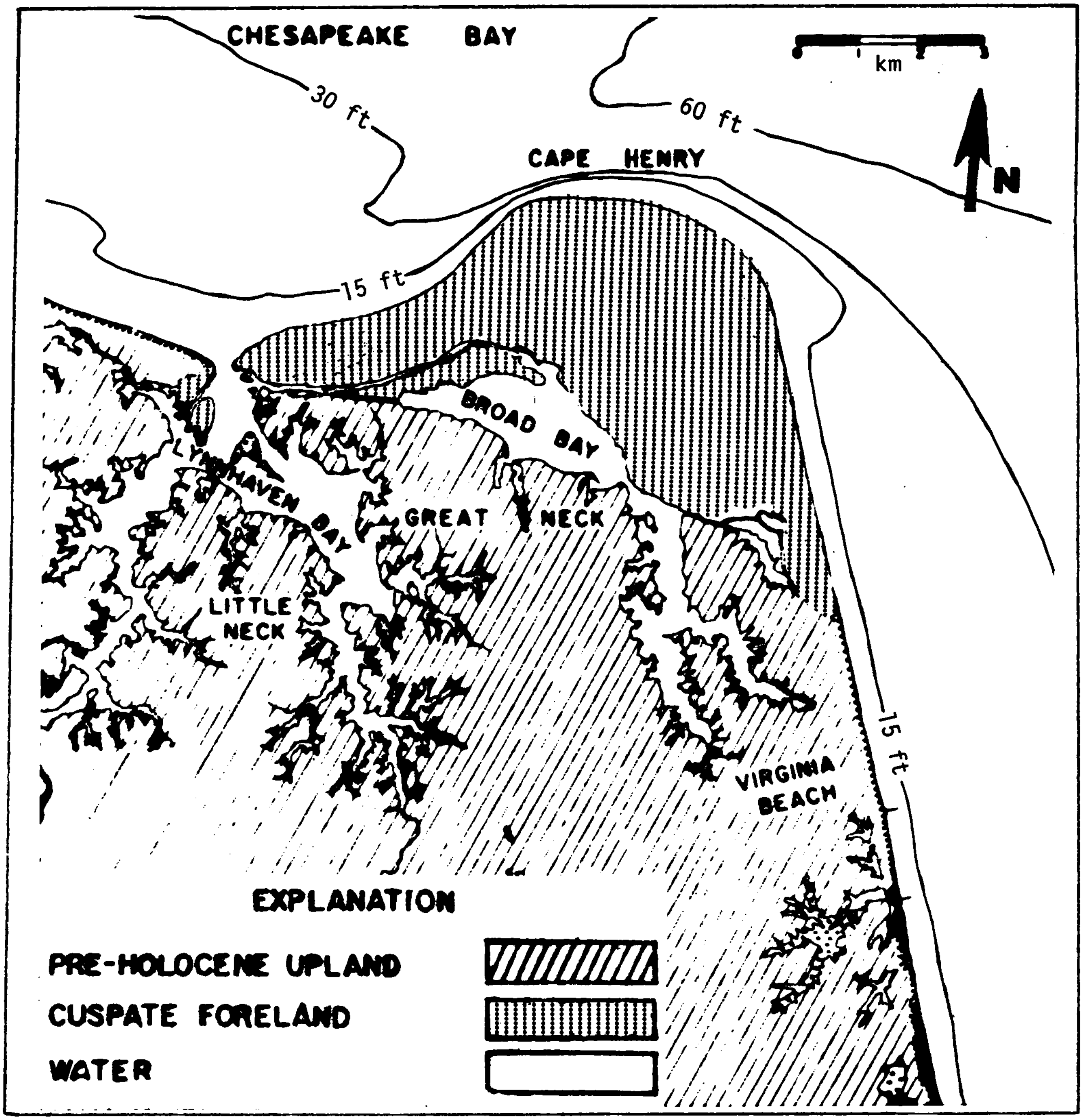


Fig. I-30
Cape Henry, Virginia. A cuspate foreland type spit comparable to Cape Henlopen 200 years B.P. (after Kraft and others 1978).

on positions older than 10,000 B.P. (Macintyre and others 1978a and b; Redfield 1967). Table I-2 lists past sea-level positions at 3,000-year intervals. Additional studies focusing on in situ materials are needed before an accurate understanding of Late Quaternary sea-level positions can be derived.

Table I-2: Approximate sea-level positions at 3,000-year intervals along the coast of North Carolina and southeastern Virginia.

	RANGE	BEST ESTIMATE	SOURCE
3,000 B.P.	-2.2 to 3.7 m	2.8 m	1,2,4
6,000 B.P.	-13.5 m	13.5 m	2
9,000 B.P.	-21 to -22.5 m	21.8 m	2,3
12,000 B.P.	-25 to -60 m?	40.0 m?	2,3
15,000 B.P.	-26 to -110 m	70 m?	2,3
18,000 B.P.	?	100 m?	

Sources

1. Redfield (1967)
2. Harrison and others (1965)
3. Macintyre and others (1978a)
4. Newman and Rusnak (1965)

7.0 DELMARVA SHELF

The next shelf compartment north of the northern North Carolina - southeastern Virginia compartment is the Delmarva shelf compartment (an acronym derived from Delaware, Maryland and Virginia). The boundaries for this compartment have also been selected somewhat arbitrarily, mainly for the convenience of this study. The southern boundary consists of the general position of an Early Holocene drainage divide which once separated the Susquehanna and James River Valleys. Holocene sea-level rise eventually submerged this divide and fused the two estuary systems, resulting in the Chesapeake Bay as found today. The northern boundary is formed by the axis of the Holocene ancestral Delaware Valley.

The amount and type of research done within this compartment, especially at its northern end, is exceptional in light of the information needed for archaeological studies. Because this is one of the few areas where data are available to support our model concerning transgression, the northern portion of this compartment is discussed in detail. The coastal section of this compartment follows the same general plan as other coastal compartments along the Middle Atlantic Bight (Swift 1970). The coast consists, from south to north, of a barrier island chain and a mainland beach flanked by barrier spits. This compartment differs from its southern neighbor in that no major valleys sub-divided it during the Holocene. It is, however, flanked on either end by major valleys belonging to the ancestral Susquehanna and ancestral Delaware Rivers. In contrast to the northern North Carolina - southeastern Virginia shelf compartment, the CS along this portion of the Atlantic coast is underlain by extensive lagoonal deposits as illustrated by the work of Field and Duane (1976), Kraft (1971, 1977), Kraft and others (1978), and Sheridan and others (1974).

The bathymetry of the Delmarva continental shelf is shown on Fig. I-31 along with the distribution of important shelf features and place names (Swift 1975a; 1976b). Chesapeake Bay joins this shelf compartment at its southern end and will be discussed first. Fig. I-32 shows Chesapeake Bay and its major tributaries.

Although considerable research has been done on the geology of Chesapeake Bay, only a very preliminary understanding is available regarding its evolution since the Late Pleisotcene. Chesapeake Bay is the largest estuary along the eastern Atlantic Coast. It represents a complex system of estuary retreat paths and modified Pleistocene river valleys. Except possibly for the James River system, the Late Pleistocene basin forming Chesapeake Bay consists of the ancestral Susquehanna River and its tributaries. Our knowledge of the ancestral Susquehanna River system is limited since detailed seismic profiling is obstructed in some areas by the presence of trapped gas, thus depriving us of some important stratigraphic information.

LEGEND   Major Late Pleistocene-Holocene shelf features

Buried valley on the shelf identified by seismic profiling and containing flood plain, river channel, river terrace and estuarine deposits

Surface valley on shelf indicating that buried estuarine and possibly river valley deposits probably exist within the general vicinity

Buried valley probably older than 35,000 B.P.

Shoal retreat massif (shelf highs of constructional origin due to littoral drift deposition centers)

Major shelf delta identified by bathymetry, morphology, and sometimes seismic data when available

Scarp

Fig. I-31

Major Late Pleistocene-Holocene features on the Delmarva Continental Shelf.

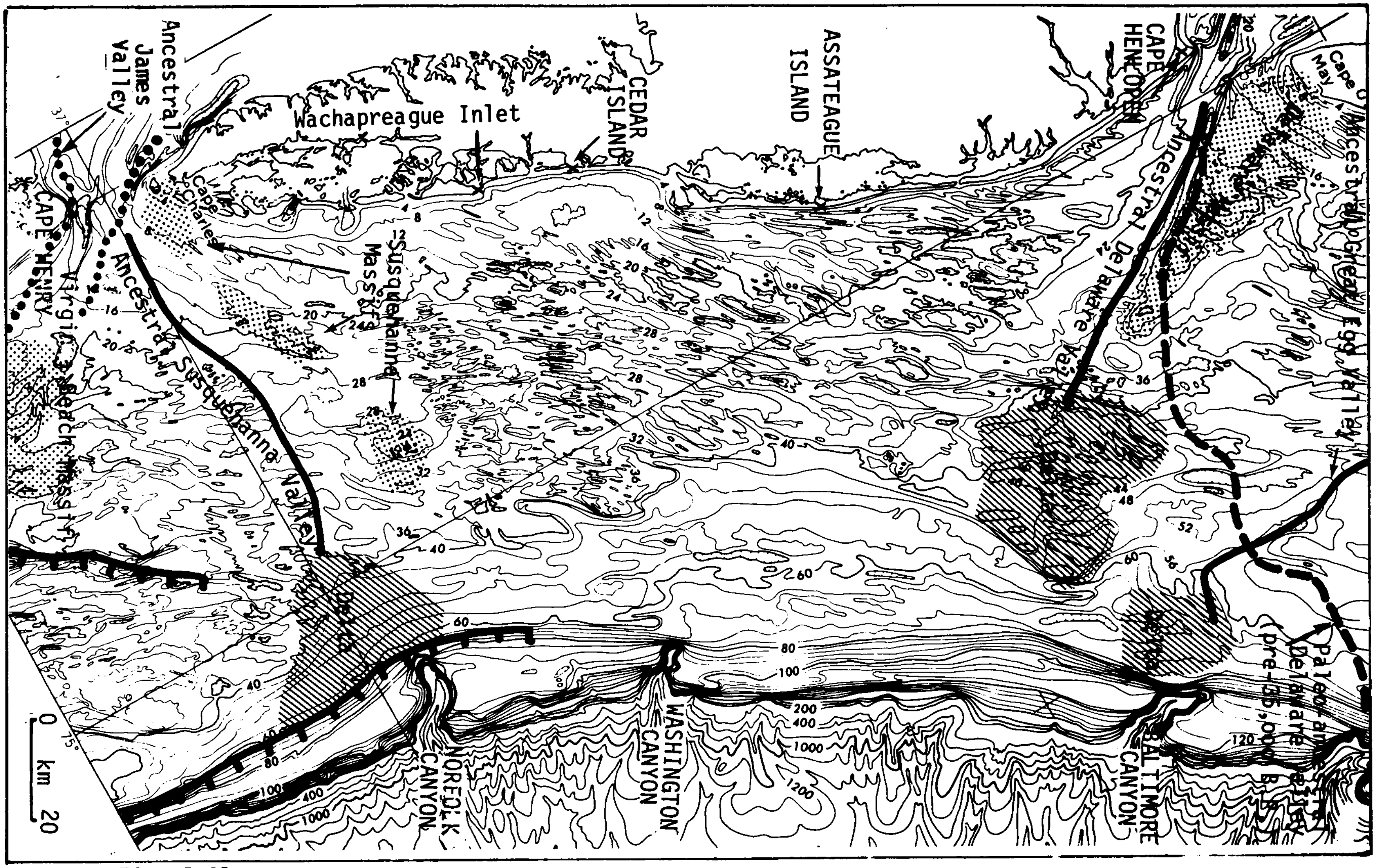


Fig. I-31

Major Late Pleistocene-Holocene features on the Delmarva Continental Shelf. Shelf features compiled from Belknap and Kraft (1977); Dillon and Oldale (1978); Duane and others (1972); Field and Duane (1976); Harrison and others (1965); Kraft (1971,1974,1977); Kraft and others (1978); Knott and Hoskins (1968); Meisberger (1972); Sanders and Kumar (1975); Sheridan and others (1975,1977); Swift (1975a,(1976b); Swift and others (1972,1978); Swift and Sears (1974); Twichell and others (1977).

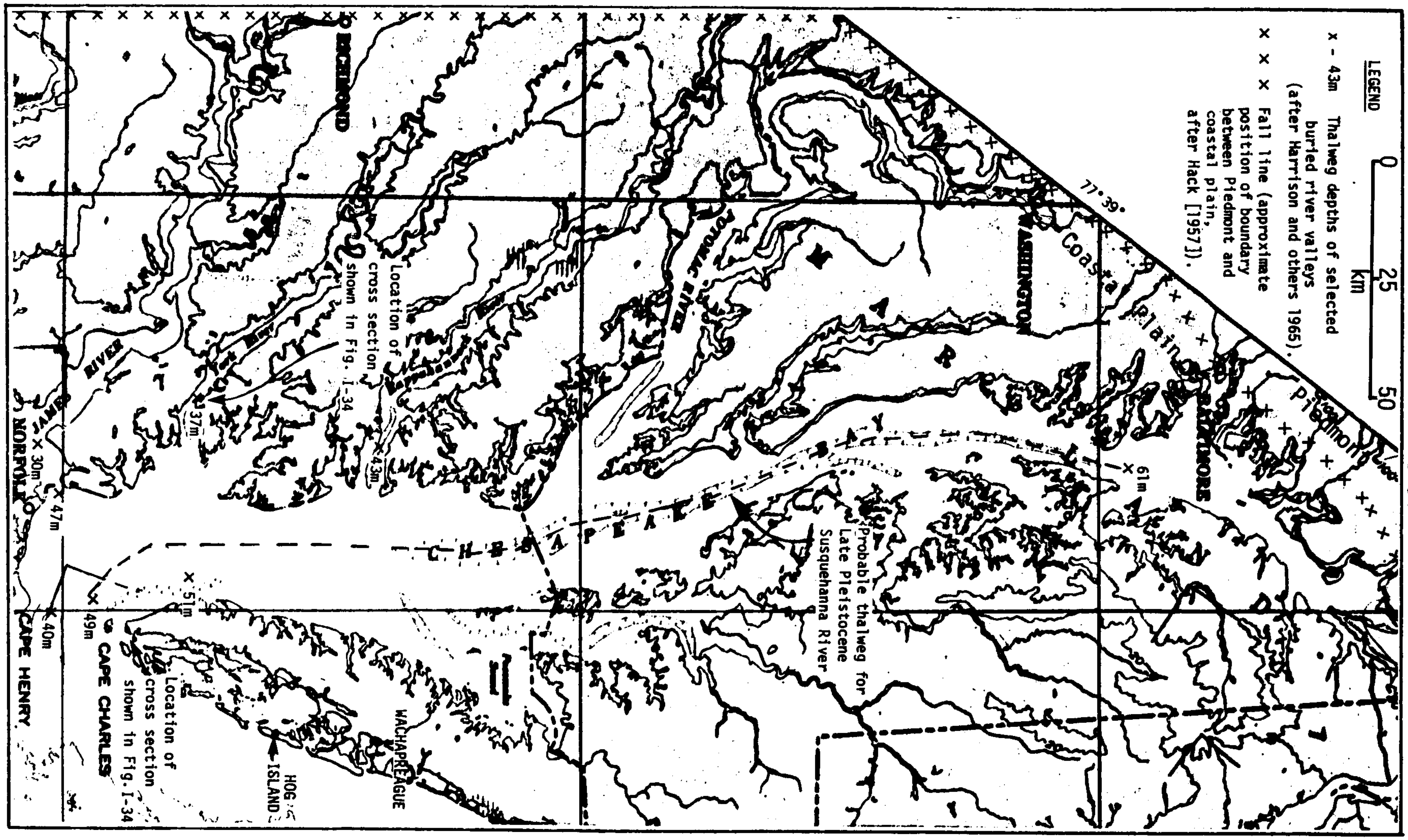


Fig. I-32

Chesapeake Bay and vicinity showing important buried valley depths and thalweg locations. Information from Harrison and others (1965). Location of profiles shown in Fig. I-34 are indicated at the mouth of Chesapeake Bay and also at the mouth of York River (Hack 1957).

Chesapeake Bay lies within the Atlantic coastal plain. Numerous rivers flow into the Bay from several directions. Large rivers draining piedmont regions enter along its western side (Fig. I-32) while smaller rivers flow in from the eastern peninsula (Delmarva) and coastal plain regions.

The fall line occurs along the zone of contact between the piedmont and coastal plain. During the Holocene, this fall line zone may have extended slightly further to the east when bedrock sills were exposed by rivers as valleys became entrenched during the last marine regression and lowstand.

The mixing of fresh and salt water in Chesapeake Bay accounts for some of the important geological processes that have changed the basin over time. The interaction of fresh and salt water can be typified by a circulation pattern where fresher (less dense) surface waters flow seaward and salty (more dense) bottom waters flow landward. Although this circulation pattern is mediated by tidal action, it represents the overall trend or direction of water flowing in the bay. At any particular phase of the tide, currents at any depth within the bay may be flowing landward or seaward, but taken on the average over longer periods, the net motion of the lower water layer is landward. This action traps most of the sediment inside the Bay rather than allowing it to reach the ocean.

Circulation in Chesapeake Bay plays an important role in the transportation and deposition of sediment entering the basin. The fate of the subaerial surface is also dependent on sedimentation patterns, circulation, and the source of the sediment moving into the basin. For convenience, Chesapeake Bay can be divided into the following two subregions, one north and another south of the Potomac River. North of the mouth of the Potomac, Chesapeake Bay is primarily controlled by the Susquehanna River. Over 90% of the fresh water entering the northern Chesapeake Bay (that portion north of the Potomac) comes from the susquehanna. From the Potomac River southward, Chesapeake Bay is jointly influenced by several major rivers as well as the net flow coming from its northern half.

Sedimentation rates are similar in the two subregions except in the northernmost section of the Bay. The central portion of Chesapeake Bay south of latitude 39° 27' N, receives about .1 cm of sediment per year (Schubel and Carter 1977; Schubel and Hirschberg 1978). The extreme northern portion of Chesapeake Bay receives an annual sediment input 6 to 10 times larger (Schubel 1974; Zabawa and Schubel 1974).

The source of sediment differs between the southern and northern sections of Chespeake Bay. The extreme northern head of Chesapeake Bay receives sediment transported by the Susquehanna River. Major floods may deposit as much as 0.2 cm in a single event (Zabawa and Schubel 1974). Over 0.5 m has been deposited in this area since 1900 (Schubel 1974, Zabawa and Schubel 1974). South of latitude 39° 27' N, most of the sediment comes from shoreline erosion (Schubel and Carter 1977). These facts have an important effect on the preservation of the subaerial surface.

The shorelines along Chesapeake Bay undergo erosional shoreface retreat similar to that experienced along the ocean. Wave action associated with tides and storms erodes the shoreline. Erosion rates vary considerably, but in general marsh regions experience a lower rate of shoreface retreat than sandy, unprotected shorelines or cliffs (Rosen 1976, Schubel 1968, Singewald and Slaughter 1949).

Slope and tidal fluctuation play an important role in determining shoreline erosion. Many flat to gently sloping areas have experienced extensive marsh development before erosional shoreface retreat has passed over them. Marsh development first follows the small stream valleys and low areas preceding sea-level rise. More steeply sloping interfluves and headlands experience less marsh development. If the local slope is steep enough for a cliff face to be cut, then total loss of the subaerial surface will occur.

The rate of shoreline erosion in the lower Chesapeake Bay has been examined in detail (Rosen 1976). Schubel (1968) and Singewald and Slaughter (1949) have considered the processes affecting shore erosion rates in the northern Chesapeake Bay. In the Virginia portion of the Bay, the rate of shore erosion of all types of shorelines (permeable, impermeable, and marsh barrier beaches) has been found to be inversely proportional to the tidal range. Since there is an almost regularly progressive decrease in the tidal range from south to north in the estuary, the mean erosion rates for all shoreline environments tend to increase in the northern estuary. The increasing erosion with decreasing tidal range is related to several effects. First it is noted that most shore erosion takes place during large storms. During storms the water level is raised by storm surges and large amounts of erosion occur in low-lying areas. Since the height of a beach above mean high tide (the supra-tidal elevation) is to a large extent proportional to the tidal range, those areas with larger tidal ranges will tend to become less heavily flooded than areas of lesser tidal range. Also, since a storm may occur at any stage of the tide, and a tidal surge makes up a greater proportion of the tidal rise in regions of small tidal range, flooding is less severe in regions of larger tidal range.

Secondly, in areas of large tidal range the energy of the wave crests is dissipated over a larger region of the shoreline than in areas of small tidal range. This dissipation of energy may account in part for the lower erosion rates in regions of larger tidal range. The character of the shoreline also has important effects on the rate of shore erosion. In general marshes, both barrier-beach protected and directly exposed, have the lowest rate of shoreline erosion. In the Virginia portion of Chesapeake Bay the average erosion rate for all marshes is about 0.57 m per year. The average rate of erosion of beaches is almost twice the rate for marshes or about 1.1 m/y. Beaches make up 80.31%, marsh barrier beaches 18.3% and exposed marsh only 1.3% of the total shoreline length within the Virginia portion of Chesapeake Bay. Marshes, of course, are located more frequently at the heads of estuaries along Chesapeake Bay, offsetting erosion in low-lying areas.

Schubel (1968) has considered the effects of shore erosion in the extreme northern Chesapeake Bay. The character of the shoreline in this region is variable. Some of the coast is bordered by low sandy beaches, other parts are bounded by sand cliffs up to 10 m in height. The rate of shore erosion is also variable, and does not seem to correlate simply with the shoreline character. The rate of shore erosion varies from 0.12 m to over 0.6 m per year. The sand derived from shore erosion in this area is deposited near the littoral zone because of water circulation patterns. The net result of shoreface erosion is the progressive straightening of the shoreline. Most of this erosion is of bank material. As the cliffs are undermined by the waves, they collapse and destroy the overlying soil horizons. It is unlikely that there would be any preservation of the old land surface under conditions of cliff retreat, since any archaeological information would lose its intra-site provenience.

Because of the higher resistance of salt marshes to erosional shoreface retreat and their tendency to cover and protect subaerial deposits from wave attack, salt marshes are important features to delimit. Essential requirements for the formation of salt marshes include protection from high-energy waves, an adequate supply of sediment to a shallow water area, and salt-tolerant plant species (Chapman 1960). In the Chesapeake Bay area these conditions are most frequently obtained near the heads of the various tributary estuaries. The extensive salt marshes in the Joppa Town area of northern Maryland are an example. These small, shallow bodies of water often have a significant sediment input, and are protected from large wave energy. It would seem likely that during the last sea-level rise, the heads of all the proto-estuaries were lined with marshes.

From the above discussion, it is possible to generalize about the preservation of the pre-transgressive subaerial surface in Chesapeake Bay. In the center of Chesapeak Bay and along each major river entering the Bay, extensive flood plain deposits were developed between 30,000 and 18,000 B.P. (see Dillon and Oldale 1978 for sea level). The central corridor of the Chesapeake basin consisted at the end of the Pleistocene of low relief meander plains and river terraces. Sea-level rise between 18,000 and about 12,000 B.P. probably drove the head of the newly forming Chesapeake estuary into the lower section of the basin. From 12,000 to about 6000 B.P. the Chesapeake estuary system expanded along many of the adjacent river systems, probably reaching to about four-fifths of its present size. For the last 6,000 years slower sea-level rise has caused the estuaries along Chesapeake Bay to increase to their present size. Infilling of all the estuaries along the Chesapeake Bay system will soon overtake sea-level rise. Eventually the head of each estuary will begin to migrate toward the ocean as infilling allows progradation of the shorelines.

Hack (1957), Harrison and others (1965), and Schubel and Zabawa (1972) have studied buried Pleistocene river channels within Chesapeake Bay. Hack (1957) combined the information obtained from borings, well logs, and sub-bottom profiles to make inferences regarding the depth of post-Miocene river channels. Harrison and others (1965) and Harrison and Rusnak (1962) conducted similar studies and proposed quite different gradients than those suggested by Hack (1957). Fig. I-33 illustrates the differences between the results of Harrison and others (1965) and those of earlier investigations.

Along the eastern shore of northern Chesapeake Bay, a buried river channel was investigated by Schubel and Zabawa (1972). Unfortunately this feature is probably Illinoian in age and consequently is of no direct relevance to archaeological studies. However, the detailed investigations conducted illustrate that flood-plain deposits do indeed become preserved beneath transgressing estuarine sediments (Schubel and Zabawa 1972).

The post-Miocene channels of several major rivers (Fig. I-34) were found to be less than 50 m below sea level (for example, the Elizabeth River: -30 m; the James River: -47 m; the York River: -37 m; the Susquehanna River: -49 m; Harrison and others 1965). In comparison to sea-level curves for the last 18,000 years, Harrison and others (1965) interpreted these results along with other data to indicate substantial Holocene uplift of most of Chesapeake Bay. Although they offered additional supporting evidence, their arguments seem to lack conclusive data. The deeper thalweg in existence near Annapolis, for example, may represent the effect of less sedimentation combined with active Holocene (tidal?) scour as water moves between the "narrows." Furthermore, the recent work of Dillon and Oldale (1978) offers a warning against using any "eustatic curves" for determining local uplift. Late Pleistocene valleys along the ancestral Delaware show no evidence of uplift (Belknap and Kraft 1977) and also do not penetrate deeper than 30 to 50 m below sea level (Kraft 1977; Sheridan and others 1974).

Other evidence in support of regional uplift of Chesapeake Bay needs further clarification and critical reexamination. Peats found between 25 and 27 m below sea level and dating 10,000 to 15,000 B.P. have not been positively identified as salt water species and consequently should not be used to construct sea-level curves. More information is badly needed about the shell bed on Hog Island (Harrison and others 1965) in order to determine whether its origin is natural. Last of all, the ages assigned to the buried river channels should be rigorously reexamined before regrouping them as contemporaneous drainage systems.

Critical review of the available data for the Chesapeake Bay drainage network leaves many questions unanswered. In comparison to the well-developed sea-level curves obtained for the Dalaware Bay area (Belknap and Kraft 1977), the inferences drawn by Harrison and others (1962, 1965) seem conspicuously anomalous. The inappropriate use of unrelated and inconsistent data may be responsible for their conclusions, which still remain unsubstantiated. Fig. I-34 gives two profiles showing the inferred buried Pleistocene valleys for the ancestral Susquehanna and York Rivers. The Susquehanna Channel is being buried by the southward

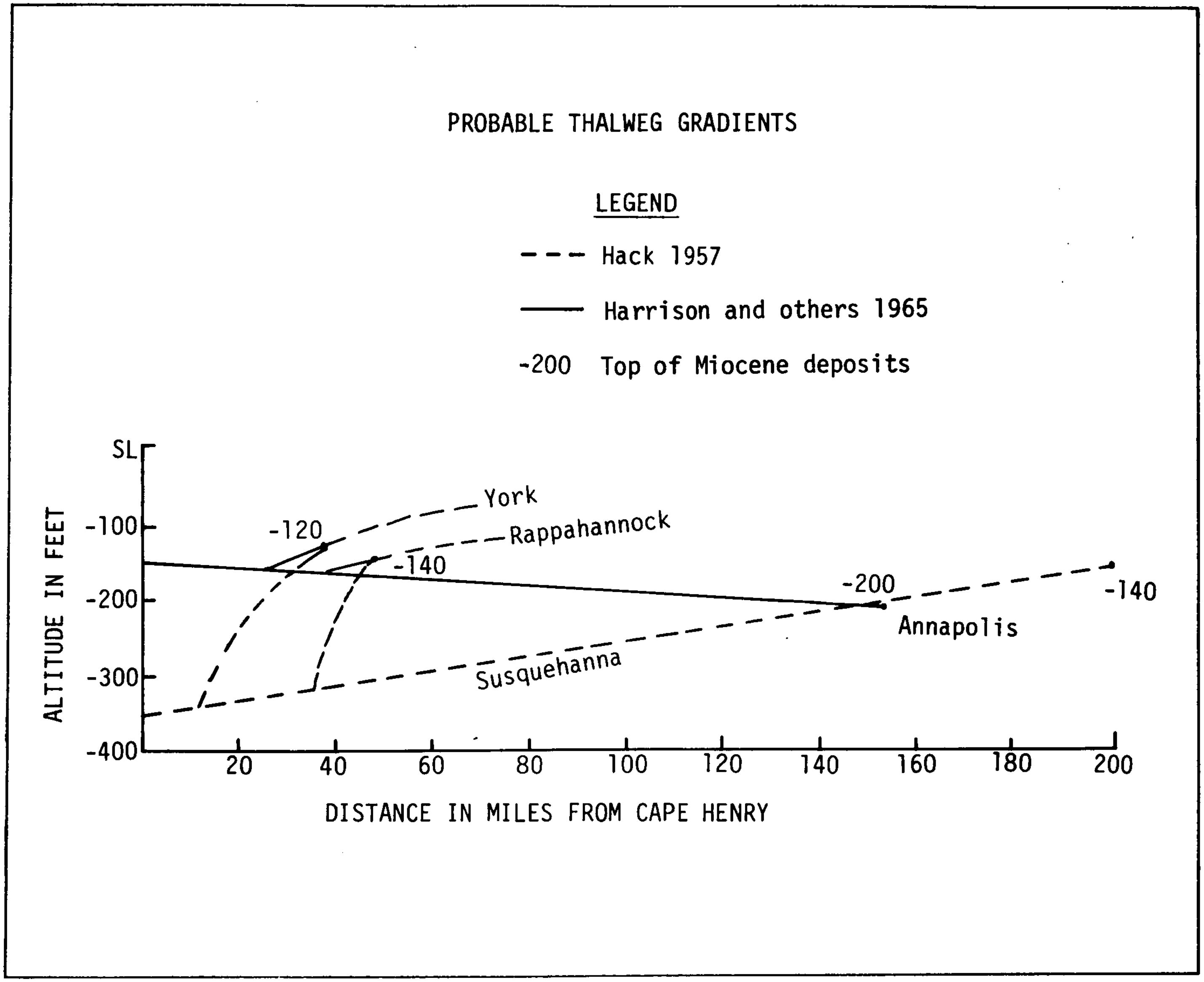


Fig. I-33

Interpretations of the thalweg gradients for several buried valleys in lower Chesapeake Bay. Harrison and others (1965) interpretation requires substantial uplift of the lower section of Chesapeake Bay and has not been substantiated to date.

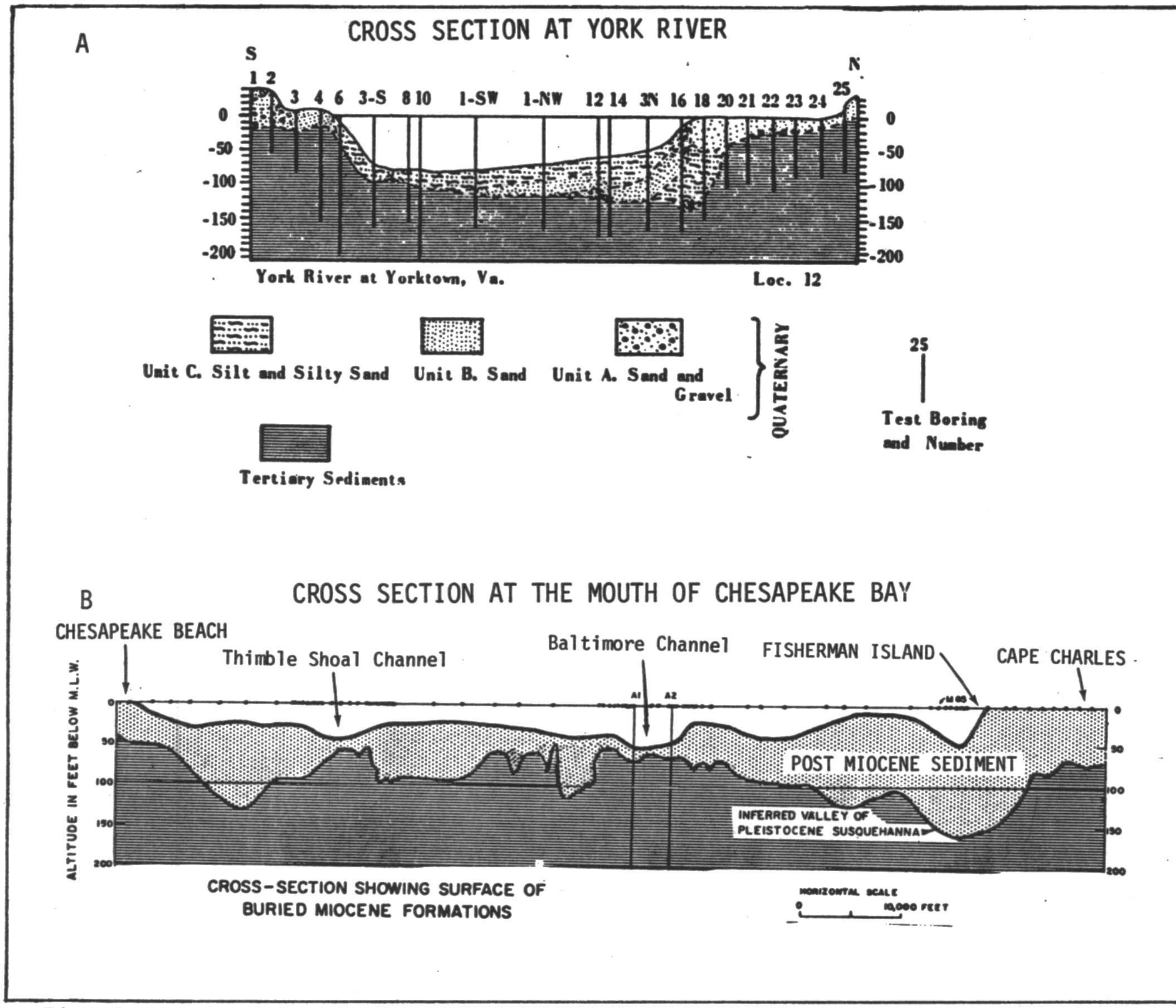


Fig. I-34 Two profiles from the southern portion of Chesapeake Bay (after Harrison et. al. 1965). Top cross section (A) taken at York River. Bottom cross section (B) taken at the mouth of Chesapeake Bay. The location of each cross section is shown on Fig. I-32.

growth of the Cape Charles platform shoals (Field and Duane 1976; Meisburger 1972). This demonstrates that littoral drift (Fig. I-35) is capable of greatly altering the pre-Holocene drainage system by partially obscuring older valleys.

The path of the ancestral Susquehanna River on the shelf east of the mouth of the Chesapeake Bay has been inferred from sub-bottom profiles and existing bathymetry (Meisburger 1972; Swift and Sears 1974; Swift and others 1972; and Swift and others 1978). The general valley corridor is shown on Fig. I-31. Also illustrated on this figure is a buried channel (Meisburger 1972) slightly south of the ancestral Susquehanna Valley. The origin of this valley is unclear, but it probably is the product of earlier fluvial erosion predating the Late Pleistocene.

North of the ancestral Susquehanna Valley corridor are several submerged "plateau-like" features (Fig. I-31). They have been designated (Swift 1976a; Swift and others 1978) as shoal retreat massifs. Shoal massifs represent areas which have received substantial sediment deposition and subsequent erosion and redistribution during the Holocene. They initially were formed on valley margins and possibly some low-lying uplands which were substantially beveled during transgression. Erosion and redistribution of massif sediment subsequent to its formation is usually accompanied by movement toward the valley center. Consequently, along the outer edge of massifs (away from the valley center), erosion and sediment redistribution would threaten to destroy the pre-transgressive subaerial surface. On the other hand, the migration of features toward valley centers would help to preserve portions of the valley floor. The center portion of a valley has the highest probability for containing extensively preserved portions of the subaerial surface beneath estuarine deposits. Flood-plain deposits should still exist beneath estuarine sediments where tidal scouring has not penetrated too deeply.

The area between the massifs and the valley corridor would probably display patchy preservation related to transgressive topographical factors. North of the massifs is an area where barrier islands have developed and migrated for at least the last 6,000 years (Field and Duane 1976; Newman and Rusnak 1965; Sheridan and others 1974). Extensive lagoonal deposits partially beveled and buried by migrating barrier islands have been identified along the Atlantic coast of Delaware and dated as early as 8000 B.P. (Sheridan and others 1974). Scour by tidal inlets and tidal creeks, however, has destroyed a portion of these deposits, leaving discontinuous preservation of the subaerial surface in some localities.

The southern portion of the Delmarva coast (Fig. I-36) has been investigated in detail in the vicinity of Wachapreague Inlet (Harrison 1972, 1975; Morton and Donaldson 1973; Newman and Munsart 1968; Newman and Rusnak 1965). On the basis of these investigations, it is reasonable for us to extend the existence of the barrier islands back for the last 6,000 years. Marsh sequences buried by lagoon sediments have given radiocarbon dates ranging from the Middle Holocene (6000 B.P.) to the present (Newman and Munsart 1968; Newman and Rusnak 1965). These sequences can range up to 11 m in thickness and cover Late Pleistocene compact silty sands which formed the subaerial surface prior to being covered by rising fresh- and brackish-water marsh. Harrison (1975) reports sequences ranging from 5 m up to 30 m deep, the latter in rare instances.

Three major physiographic subdivisions were recognized in this area by Newman and Munsart (1968). These were barrier islands, lagoon complexes (marshes, tidal flats, tidal channels, and shallow bays), and upland regions forming the western boundary. An important interpretation presented by Newman and Munsart (1968) is that before about 1000 B.P. extensive tidal flats and open bays existed along this portion of the Delmarva coast. Since then, extensive tidal marshes have covered these areas as a result of decreased submergence rates and increased sedimentation. The observed shift in the extent of two of the physiographic environments (marshes vs. tidal flats) may have had important repercussions for some Late Woodland settlement patterns.

Morton and Donaldson's (1973) investigations of inlets in the Wachapreague region indicated that these features generally occupied former Pleistocene stream valleys and remained in these corridors during much of the Holocene. Barrier islands were found to occupy the drainage divides between stream systems. Harrison (1975) also noted that these channels did not migrate.

A sea-level curve for this portion of the Delmarva coast was constructed from the work done by Newman and Rusnak (1965) and is shown in Fig. I-18. Newman and Munsart (1968) point out that the idea of a single absolute curve for the eastern Atlantic Coast is unrealistic. Local events make it difficult to correlate widely separated regions. The indication of such a hypsithermal "high stand" as shown by this curve has not been observed elsewhere (for example, Belknap and Kraft 1977) and may be in error. Several of the critical radiocarbon dates use the total humic fraction which does not always produce accurate results in comparison to other organic materials (Belknap and Kraft 1977). The "uplift" or relative "high stand" suggested by Newman and Rusnak (1965) is not supported in the extensive research reported by Belknap and Kraft (1977) for the entire Delaware coast as discussed further along in this subsection.

North of the Wachapreague region and south of Maryland, little evidence concerning the Holocene evolution of the Delmarva coast is available. Some investigations (for example, Swift 1976a) of ridge development have taken place but offer little insight into the problem of the preservation of a subaerial surface beneath the transgressive "sand sheet." Investigations near Chincoteague Shoals, for example, illustrate the active nature of nearshore ridges (Fig. I-37) and their southward migration during a 50-year period (Duane and others 1972; Swift 1976a). Fig. I-38 shows the distribution of ridges near Ocean City, MD and their hypothetical evolution during transgression (Swift 1976a; Swift and others

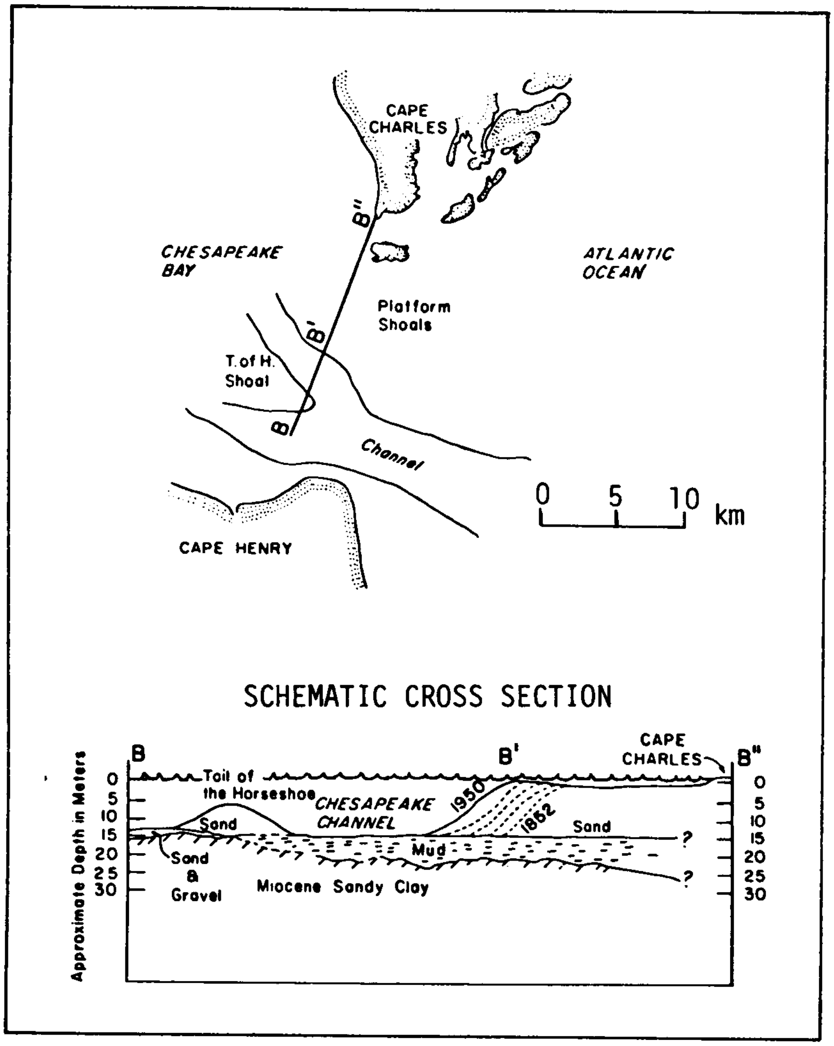


Fig. I-35

Schematic cross section of the mouth of Chesapeake Bay depicting burial of the pleistocene Susquehanna River Valley by Cape Charles platform shoals. After Field and Duane (1976). Dates indicate net accretion of shoals over the past century.

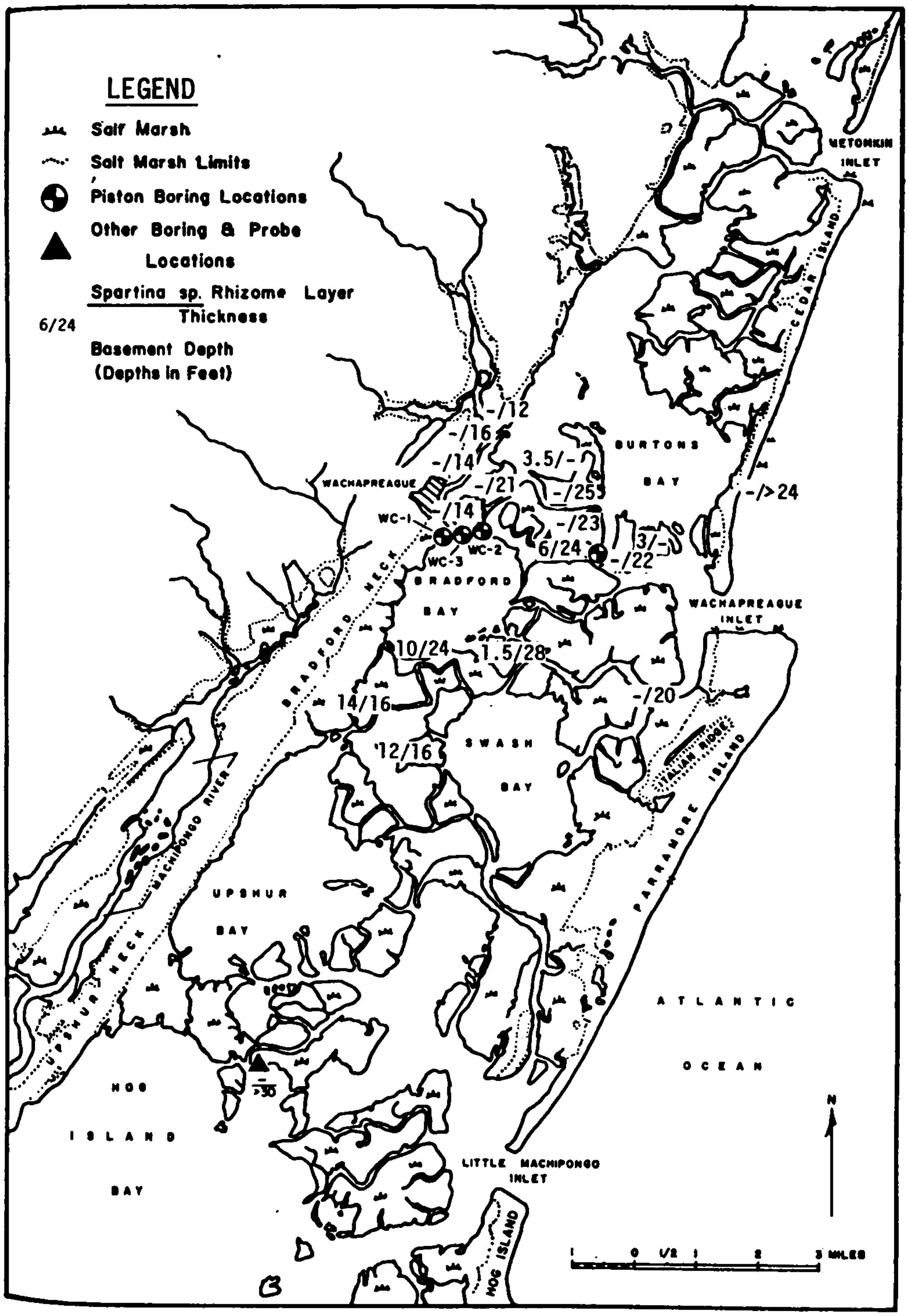


Fig. I-36

Area studied by Newman and Munsart (1968) in the vicinity of Wachapreague Inlet. Thickness of peat, boring locations, and depth to basement indicated (after Newman and Munsart 1968).

1974). The southward migration of these ridges during historic times has been sufficient to bury historic material under several meters of sand. The currents responsible for redistributing sand along these ridges usually are not strong enough to transport gravel and larger pieces of material. Consequently, shipwrecks in this area would either become buried under the southwestward-migrating sand ridges or would be added to the lag deposits exposed in the scour feature (Fig. I-37). Erosion at their northern ends may also expose previously buried material possibly including some subaerial deposits.

Field and Duane (1976) report encountering extensive Holocene back-barrier and lagoon deposits along the Maryland Inner Continental Shelf. They offer a generalized vertical section for this portion of the Middle Atlantic Bight (Fig. I-39a), which suggests a high potential for encountering an intact pre-transgressive subaerial surface. Fig. I-39b gives a representative profile for the Ocean City, MD section of the Delmarva Inner Shelf. On the basis of data from 75 cores and 700 km of seismic profiling, Field and Duane (1976) offer substantial evidence that these deposits may be an intact buried subaerial surface although they do not address this particular issue. The preservation of these deposits along the Maryland Inner Shelf may be due to the original gentle slope of the pre-transgressive topography, allowing substantial lagoons to form behind barrier islands and spits.

North of the area investigated by Field and Duane (1976) is a fairly large region which has received considerable examination (Kraft 1971, 1977; Kraft and others 1978; Sheridan and others 1977). For those concerned with the pre-transgressive subaerial surfaces, this is the best documented region between Cape Hatteras and the Gulf of Mexico. Extensive use has been made of vibracores and borings, high-resolution seismic profiling, and radiocarbon dating. Systematic sampling and overlapping traverses have given investigators in this area greater understanding of the Inner Shelf and nearshore deposits. Sophisticated geo-biological research has added important ancillary information to some of these studies, making it possible to draw inferences about the paleo-environment.

Investigations along the Atlantic coast of Delaware and along Delaware Bay have revealed the complexity of the Holocene transgressive deposits. Fig. I-40 gives the location of those areas discussed below.

The Holocene evolution of the coastal and nearshore environments of the Atlantic coast of Delaware has been investigated by Kraft (1971, 1977); Kraft and others (1978); and Sheridan and others (1974). These investigations have revealed that extensive Holocene lagoonal deposits exist beneath the transgressive "sand sheet" and have been truncated during the landward migration of the shoreline. These Holocene lagoonal deposits are thickest over pre-Holocene depressions and drainage corridors. Marsh peat as old as 7500 B.P. has been encountered beneath the lagoonal sediments, suggesting that estuarine or lagoonal conditions existed at that early date (Sheridan and others 1974). Older marsh

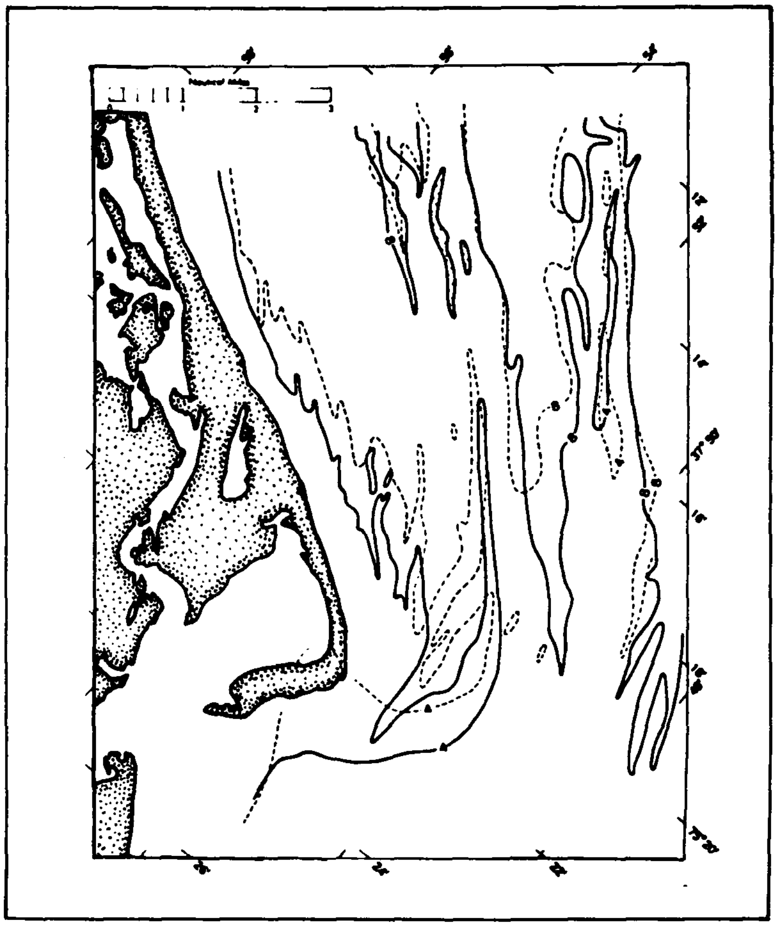


Fig. I-37

Mobility of the nearshore surficial sand sheet near Chincoteague Shoals, Virginia. Sand ridges have migrated slightly offshore and to the south between 1881 (dashed line) and 1934 (solid line). After Swift (1976a).

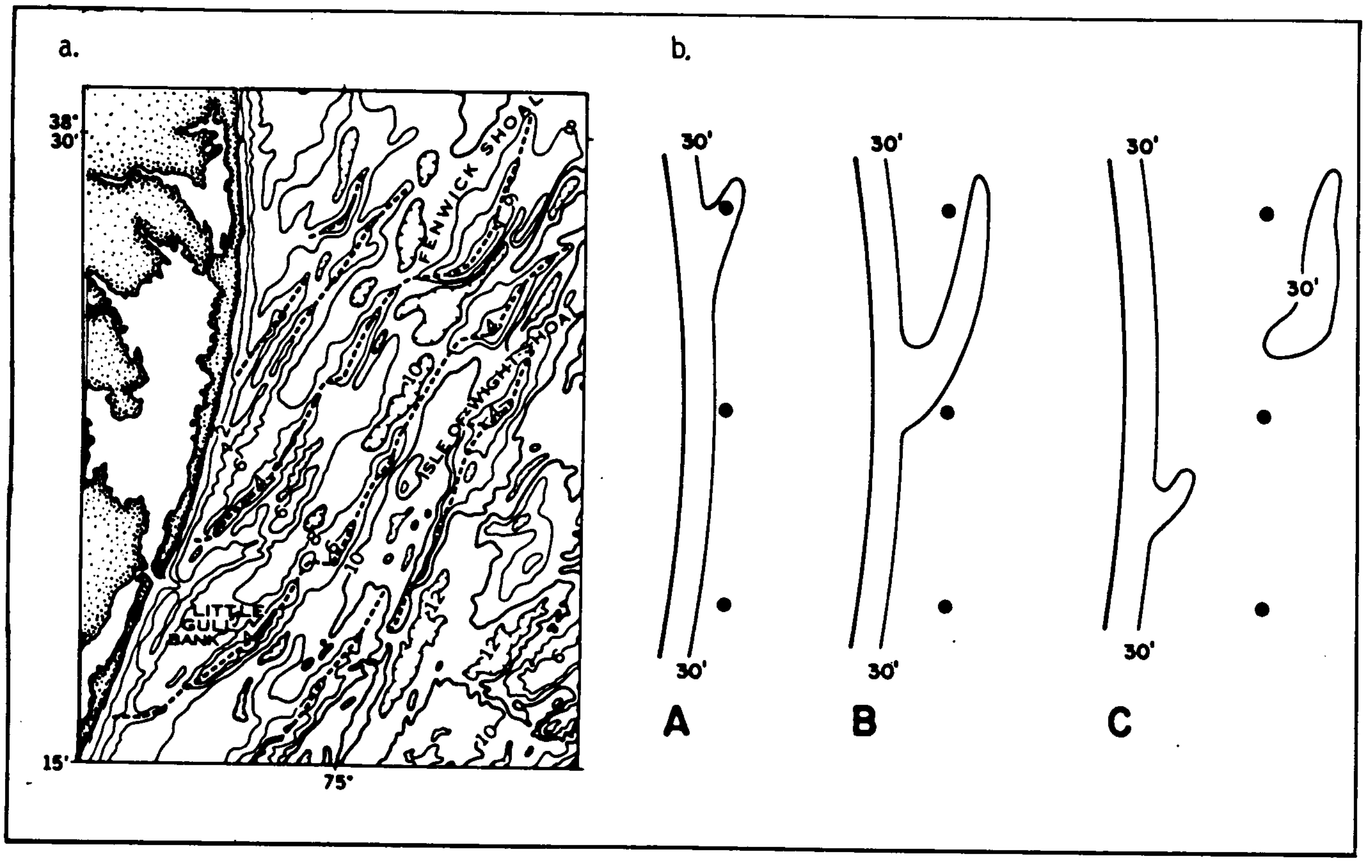


Fig. I-38

Inferred evolution of nearshore ridges along a portion of the Maryland coast (after Swift 1976a). (a) Shoreface-connected ridges of the Maryland inner coast, contoured at 2 fathom intervals. (b) Schematic diagram showing sequence of ridge detachment as inferred from (a). Dots depict hypothetically fixed points during period of erosional shoreface retreat.

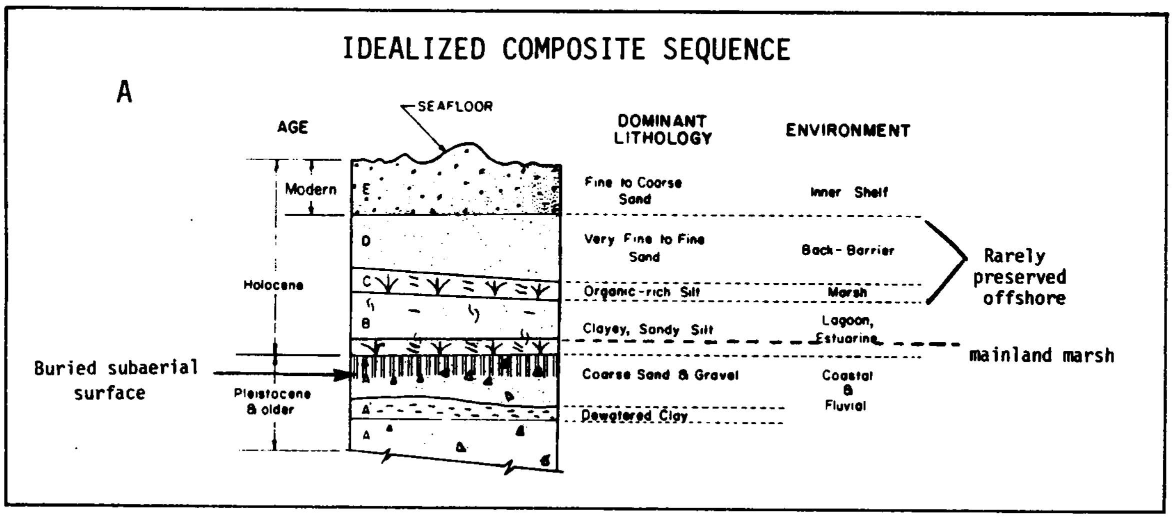
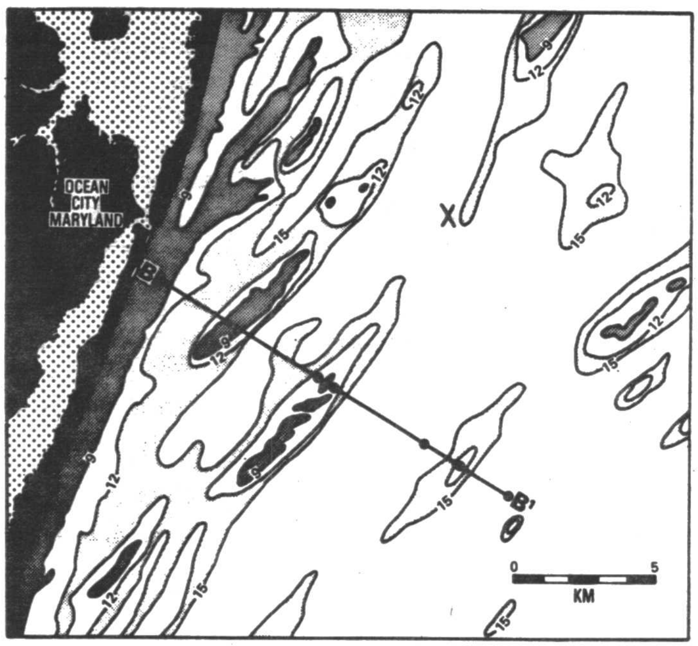


Fig. I-39

Types of subsurface deposits found nearshore along the Maryland coast. (A) Generalized section for sequences found along the middle Atlantic shelf off barrier island - spit complexes (adapted from Field and Duane 1976). Unit A is always present although its upper section may have been truncated. Position of the buried subaerial surface has been deduced hypothtically. The marsh unit directly over the subaerial surface is commonly present although not always very thick. Units B and E are usually present. Unit D is commonly present landward of the barrier front but absent offshore. Unit C is rarely present.

B



C REPRESENTATIVE NEARSHORE CROSS SECTION FOR MARYLAND

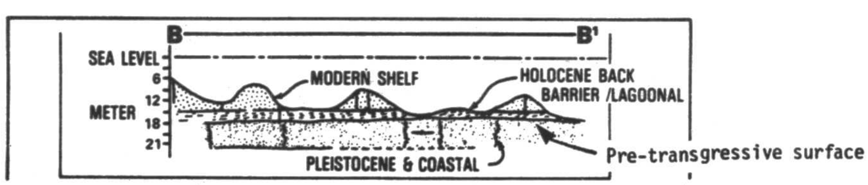


Fig. I-39 (continued)

peat has been encountered in a buried valley near Ocean Bay Inlet (Fig. I-41) and may indicate that estuary conditions in some of the deepest valleys began as early as 11,124 B.P. If the material is in original context, it would represent an estuary or lagoon system extending over 50 km inland from the shoreline suggested for 10,800 B.P. The present-day Delaware Bay system includes estuarine environments extending inland a similar distance from the Atlantic coastline at its mouth.

Along the Inner Continental Shelf and coastal regions of Delaware sedimentary units encountered landward of depths of 15 to 20 m frequently include Holocene lagoonal and marsh sequences covering Late Quaternary surfaces, most probably representing the subaerial surface of interest to archaeological studies. Kraft (1971) and Sheridan and others (1974) mention oxidation zones or soil horizons sometimes recognizable at the unconformity between the Holocene transgressive deposits and the earlier Pleistocene "surface." Kraft (1971) offers some criteria (Table I-3) for identifying the pre-transgressive erosional surface which in many cases would consist of the subaerial surface. As mentioned previously, the use of the term pre-Holocene to identify the pre-transgressive subaerial surface is incorrect for many areas, especially when dealing with Inner Shelf regions. Some of these surfaces would have been active throughout the Early and Middle Holocene up to the time when marshes began to encroach upon them.

Between Bethany Beach and Cape Henlopen, DE, numerous vibracores, drillings, and high-resolution refraction profiles have been taken. Fig. I-42 shows the location of profiles A through J taken along this portion of the Atlantic coast. Sheridan and others (1974) drew the configuration of the partially truncated pre-transgressive subaerial surface from some of these profiles (pre-Holocene erosional drainage surface). Although there appears to be some room for modifying their interpretation (for example, Fig. I-41), the general location of upland areas and drainage corridors (valleys) would remain essentially in the same regions. Fig. I-43 shows their results.

Several important points can be observed from the results of the above-mentioned work by Sheridan and others (1974). First, the relief of the pre-transgressive surface (Fig. I-43) does not correspond with the configuration of the bottom (Fig. I-42) as it exists today. Interfluves between the valleys were truncated during transgression and the valleys themselves were filled considerably. The end result was a fairly level surface whose present relief is a result of storm currents (or tidal currents as one approaches the mouth of Delaware Bay).

Another important point illustrated by Fig. I-43 is the type of pre-transgressive topography to be found adjacent to a mjaor river system (that is, the Delaware River). In general, allowing for the removal during transgression of 10 to 15 m from major interfluves and headlands, the Early Holocene topography would have been quite different in this area. Where many broad shallow bays exist today, incised narrow streams

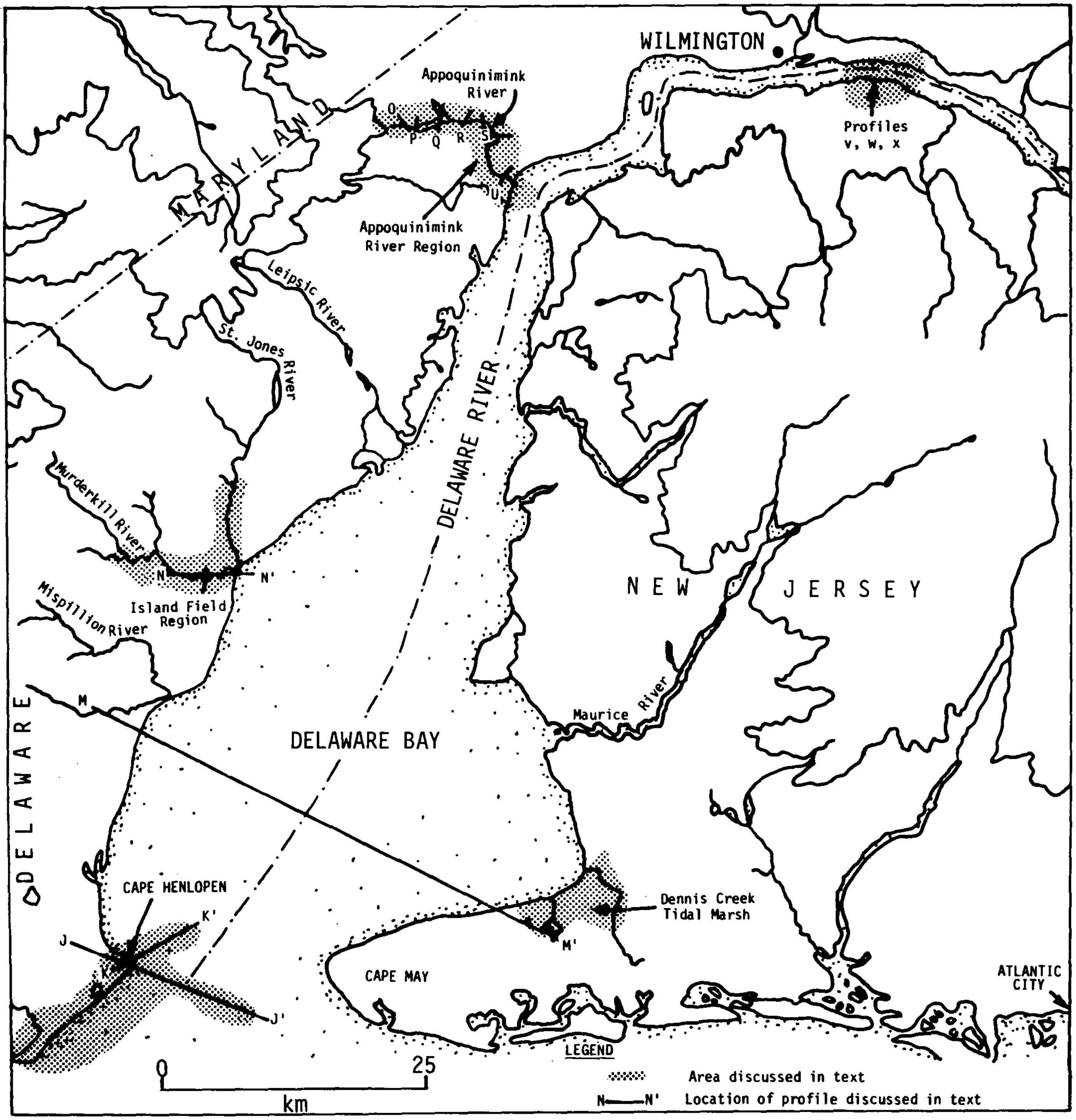


Fig. I-40
Index map of Delaware Bay showing areas and profiles discussed in text.

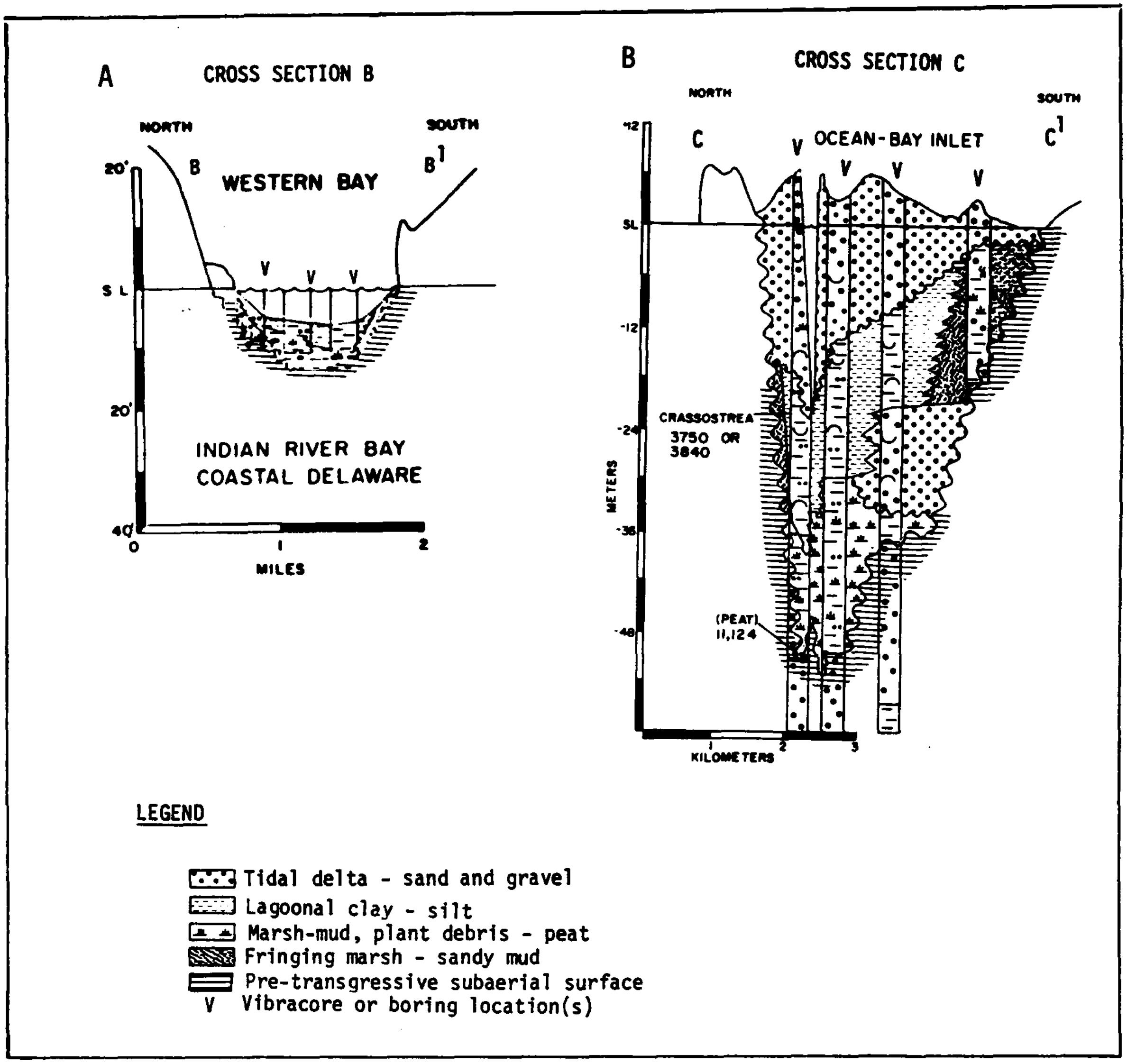


Fig. I-41

Inland (A) and coastal (B) cross sections along Indian River, Delaware. Cross section B shows considerable filling in of the valley during Holocene transgression. As shoreface erosion continues, the upper 12-15 m of the sequence (Fig. B) will be truncated over the next several millenia. Cross section B adapted from Kraft (1971) and C adapted from Kraft (1977). Position of the buried subaerial surface has been deduced hypothetically. Location of these cross sections given in Fig. I-42.

Table I-3: Some criteria for recognizing the pre-transgressive subaerial surface (after Kraft 1971:2133 Table 1).

1. Change in sediment characters such as mottling and oxidation of borings, plant debris, and other sediment features.

2. The more compact nature of the muds and their varicolored nature (gray and dark green overlying unconformity surface; brighter green, orange, tan, yellow, and gray under unconformity surface).

3. The lack or low quantity of decaying organic materials such as marsh grass and wood fragments under the surface.

4. Direct correlation with areal distribution of surface Holocene and Pleistocene sediment types and patterns.

5. Radiocarbon age determination of organic matter from the Holocene sediments and from a limited number of the Pleistocene sediments.

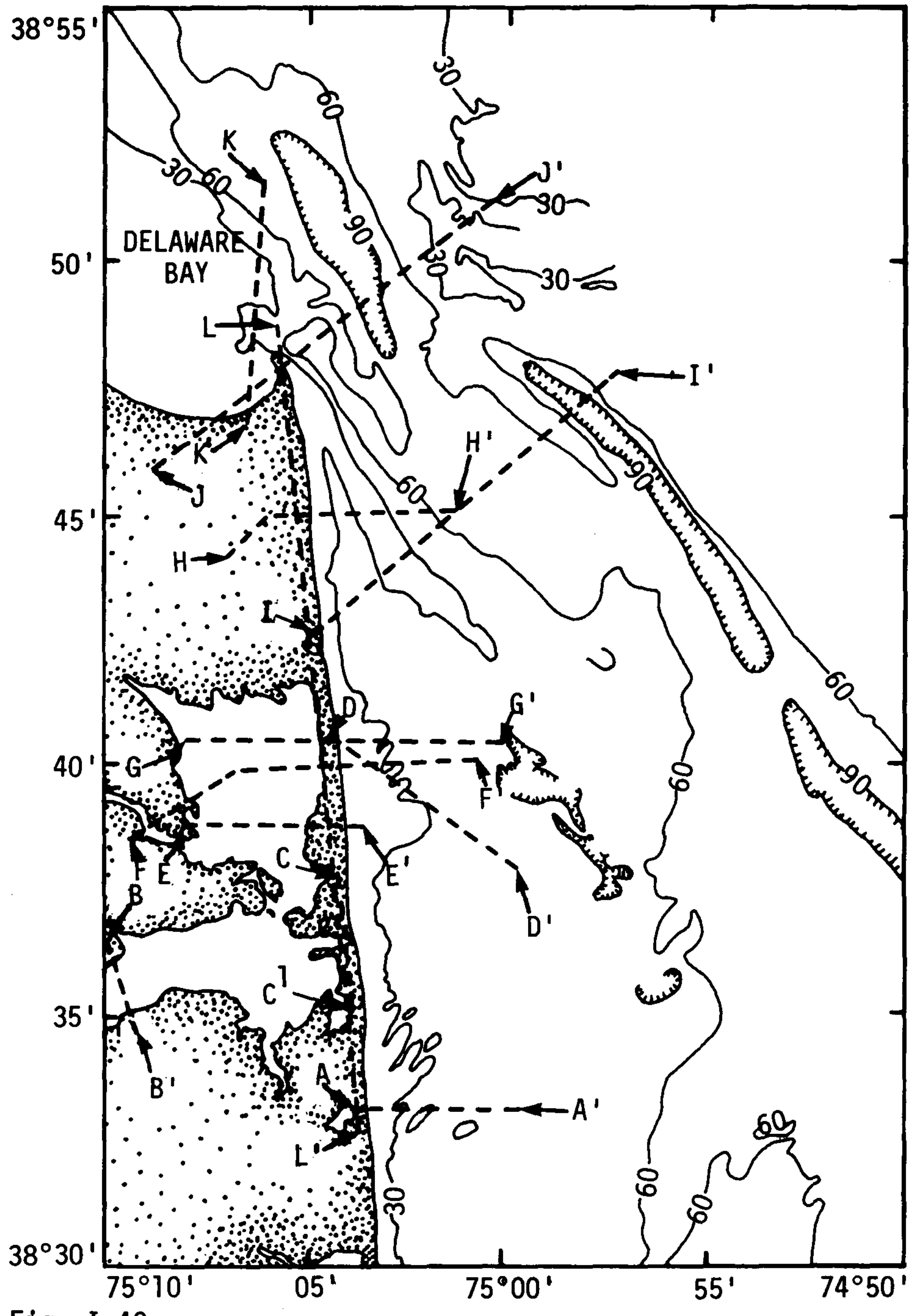


Fig. I-42. Place names and locations of profiles discussed in text. Adapted from Sheridan, Dill, and Kraft (1974).

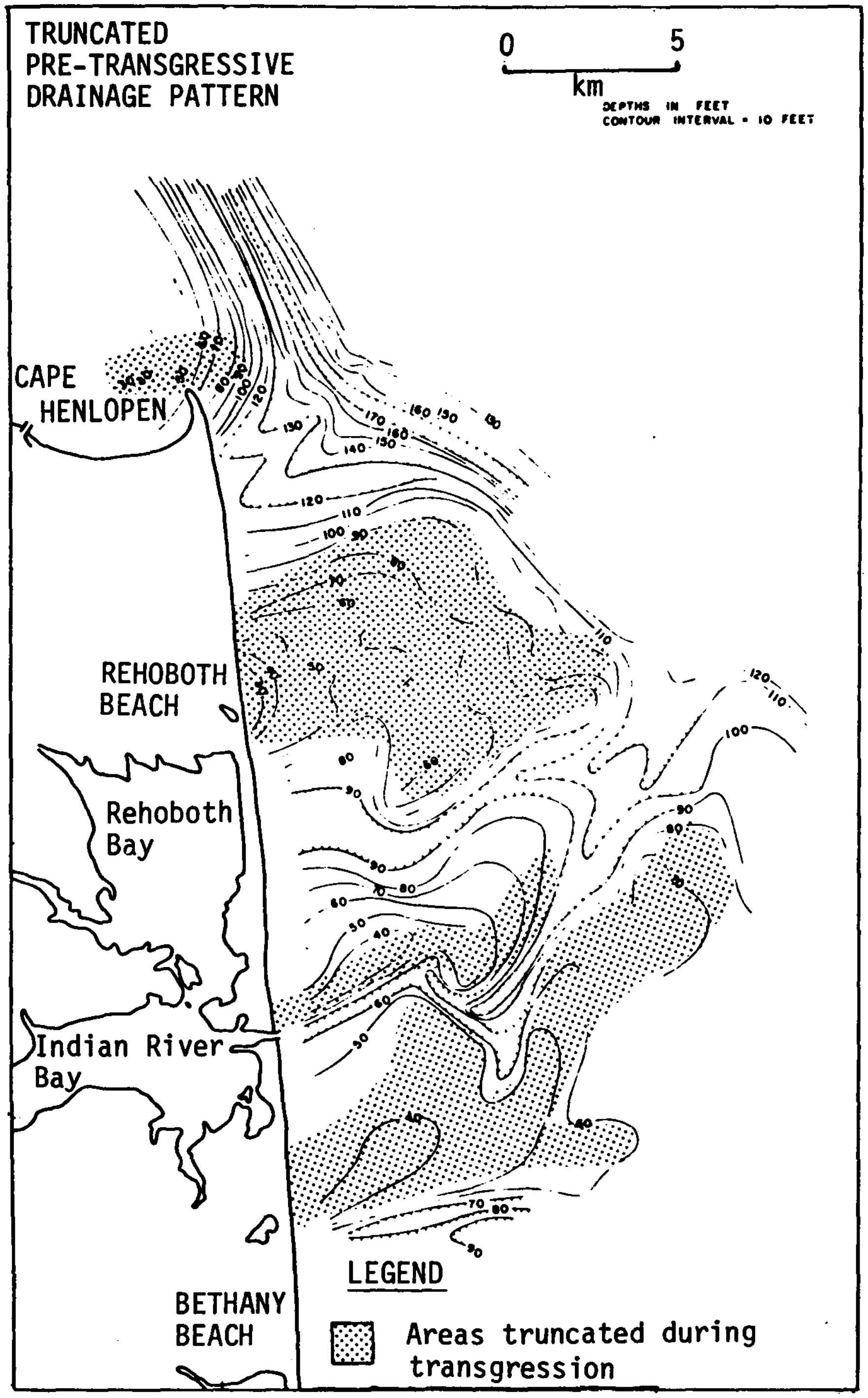


Fig. I-43

Configuration of the truncated pre-transgressive subaerial surface showing preservation of major drainages. Shaded areas have been truncated during erosional shoreface retreat destroying large sections of the pre-transgressive subaerial surface. Adapted from Sheridan et. al. (1974). Depth in feet. Surface shows up as a prominent seismic reflector (Sheridan et. al. 1974).

and valleys once existed. In areas where valleys were sufficiently far apart, broad plateau-like interfluves unprotected by barriers would have existed 35 to 50 ft higher than those shown in Fig. I-43. East (offshore) of Bethany and Rehoboth Beaches, the remains of two such upland areas can be seen in Fig. I-43. In these areas, erosion of the cliff face would have destroyed the subaerial surface during transgression in a manner similar to that occurring along mainland beaches (that is, headlands unprotected by barriers) today. Erosion which proceeds for several kilometers offshore results in the significant beveling of major interfluves and the removal and reworking of up to 15 m of unconsolidated sediment. If the interfluve was unprotected by barriers during transgression and instead confronted the ocean with a cliff, then there is little preservation of the subaerial surface during erosional shoreface retreat. This situation reduces the prehistoric archaeological potential of an area to zero but does not change its historic potential.

Landward of the bay-mouth barrier more extensive portions of the pre-transgressive subaerial surface are preserved beneath lagoonal deposits. As the shoreline migrates across these lagoons, only the more-deeply-buried subaerial surfaces will escape transgressive erosion and remain intact beneath truncated lagoonal sequences.

Although each profile (Figs. I-44 through I-55) is self-explanatory, it is useful to point out the following. Each cross section has been modified to show where the pre-transgressive subaerial surfaces may exist reasonably intact. Regions where the pre-transgressive surface has been substantially eroded are also emphasized on each cross section. Beneath the unconformity shown for major river valleys, there would quite probably exist preserved flood-plain and terrace deposits. These varieties of intact subaerial surfaces would also be of interest to archaeologists. There is a possibility that flood-plain deposits may exist above the "Holocene-Pleistocene" unconformity since they may not always be easily separable from other deposits.

Cross section $A\text{-}A^1$ (Fig. I-44b) is located in a headland area undergoing erosion at present. The pre-transgressive subaerial surface has been removed for the first 1.5 km of this cross section. The lagoonal sediments shown in the vicinity of drill hole 8-DH-70 (near A) are Middle Wisconsin in age. The deeper sequences further seaward of these are interpreted on the basis of their acoustical properties, since vibracores were unable to reach these depths (Sheridan and others 1974). Cross section $A\text{-}A^1$ also illustrates the fact that while the modern submarine bathymetry does not correlate directly with the relief of the pre-transgressive surface, in this particular case a negative correlation exists. Time studies show that even this type of correlation is not predictable, as Fig. I-44a illustrates. Fig. I-41 shows the evolution of sedimentary environments along the ancestral Indian River. Unless there exists a bedrock sill between these two cross sections (B and C), the shallow depth of the filled ancestral Indian River Valley in the vicinity of Western Bay (cross section B) is questionable. The picture is further in question since Kraft (1971 and 1977) shows the

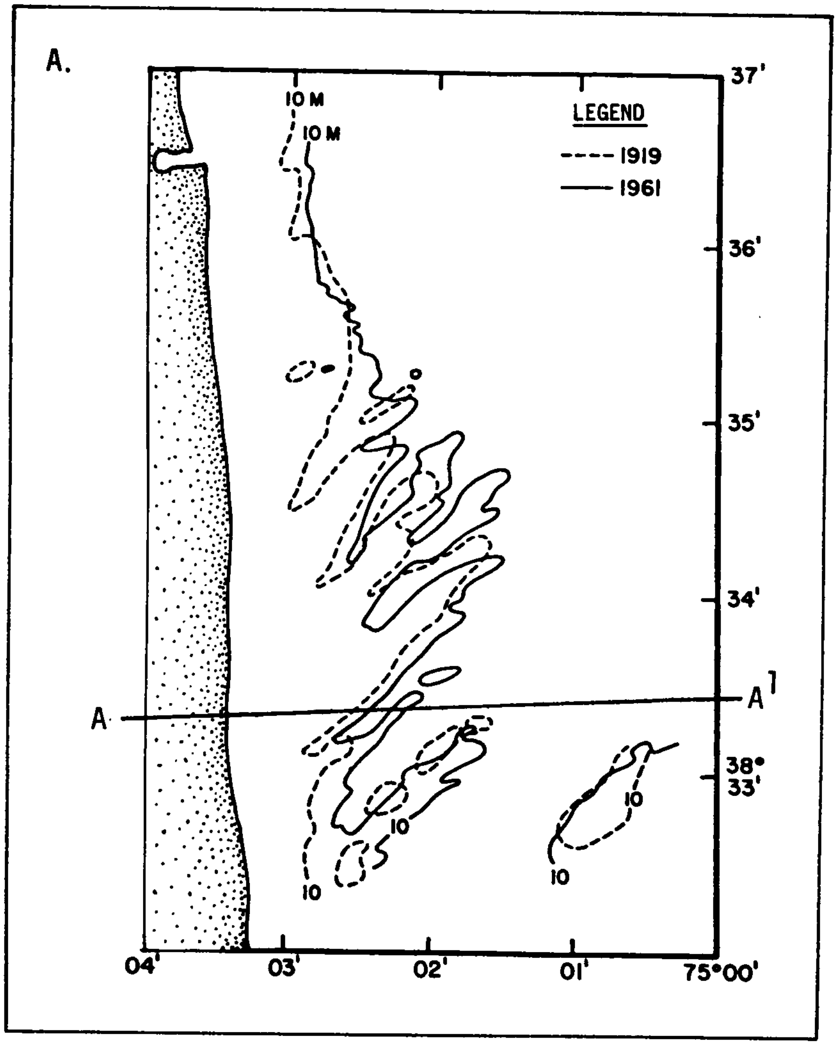


Fig. I-44a

Plan view of the Bethany Beach ridge field. The 10 m contour shows net accretion/erosion between 1919 and 1961 on the above figure. Adapted from Swift 1976a.

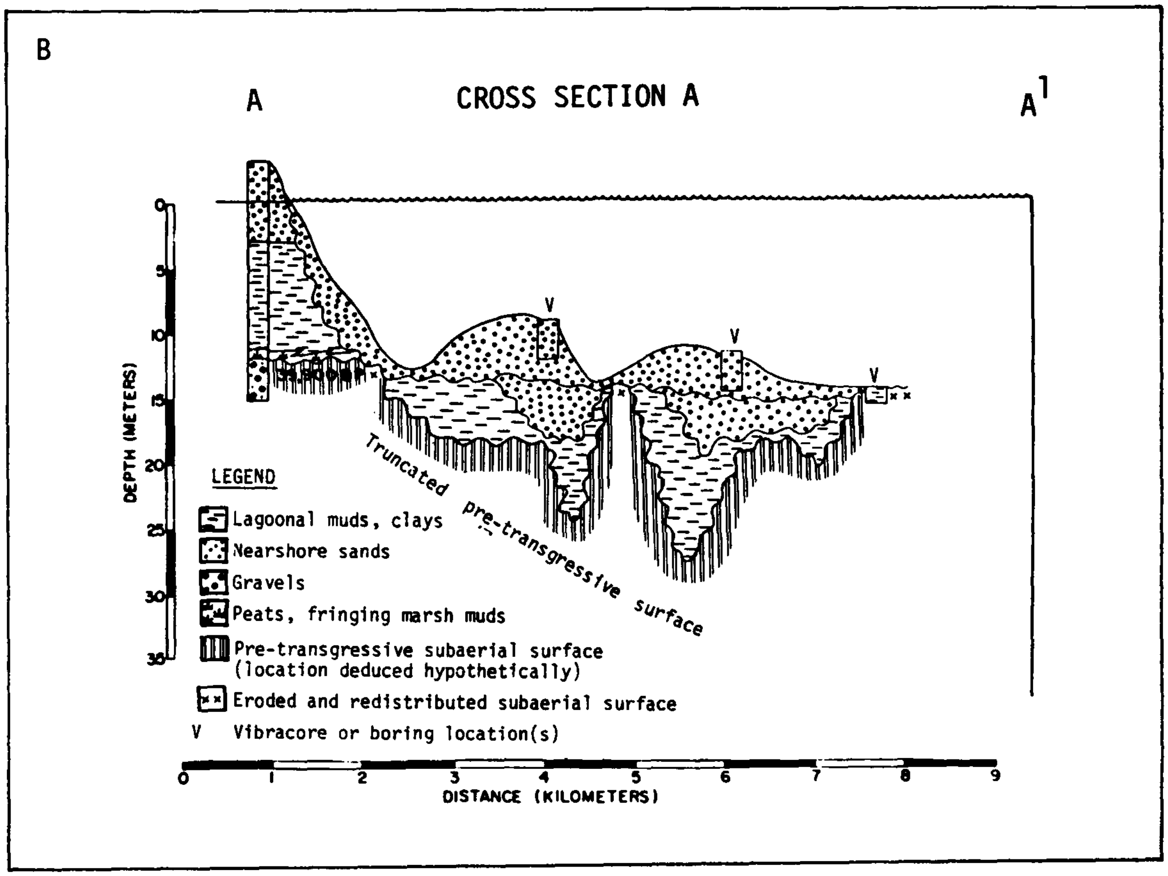


Fig. I-44b
Cross section of the Bethany Beach ridge field. Extensive sections of the pre-transgressive subaerial surface may exist up to 8 km from the beach. Location of this cross section shown on Fig. I-42. Adapted from Sheridan and others 1974.

same profile with depth differences amounting to 21 m for the location of the dated marsh peat. Depending upon which scale is correct, the gradient of the ancestral Indian River may have been somewhere between 70 and 160 ft for a 6-mi section (21 to 50 m for about 9.6 km). If a bedrock sill is responsible for restricting entrenchment along the ancestral Indian River, then a significantly different settlement pattern may have existed in this area reflecting the fact that prehistoric peoples may have gathered there for the purpose of harvesting migrating fish.

The Rehoboth Bay cross sections (Figs. I-45 through I-49) can be subdivided for discussion. Cross section E and the landward portions of F and G illustrate the sedimentary sequences before the bay-mouth barriers have migrated over the area. Cross section D and the seaward portions of sections G and F show truncated lagoon sequences beneath the transgressive "sand sheet." The seawardmost limit of each of these cross sections (sections D, E, F, G) intersects an ancient pre-transgressive headland. Erosion of the headlands has removed Holocene and Late Pleistocene soil zones leaving behind oxidized quartz-pebble sand and gravel (Sheridan and others 1973:1324). Basal peats were encountered along several of these transects, providing samples for establishing local sea-level curves. Belknap and Kraft (1977) have made extensive use of these materials to construct on the most complete "local" sea-level curves from the East Coast.

Fig. I-49 also shows the sedimentation rates for the central portion of Rehoboth Bay. The substantially lower sedimentation rates between about 5500 and 2800 B.P. probably represent the period when sea-level rise made this section fall well within a central bay environment.

Fig. I-50 is a generalized cross section of Rehoboth Bay and may be representative of other tributaries along the Delaware River system. The two other cross sections illustrate the bay shoreface environments.

Figs. I-51 and I-52 are cross sections north of Rehoboth Bay in the vicinity of a mainland beach (Rehoboth Beach). Cross section I (Fig. I-52) shows that erosion of a headland region during transgression has removed the subaerial surface for a distance of 8 km from the shore.

In the vicinity of cross section H (Fig. I-51) the shoreline has prograded in response to sea-level rise and the northwestward migration of the ancestral Cape Henlopen (Kraft and others 1978).

The vicinity of Cape Henlopen has been extensively investigated (Belknap and Kraft 1977; Kraft 1971; Kraft and others 1978). Fig. I-53 gives a cross section through Cape Henlopen and the mouth of Delaware Bay. Tidal currents have disrupted some of the Holocene and pre-Holocene sequences in this area. Moody and Van Reenan (1967) were able to delineate the ancestral Delaware River channel on their seismic profiles by its cut-and-fill sequences (Fig. I-54). Northwest of cross section J, Sheridan and others (1974) found another beveled headland. Fig. I-55 is

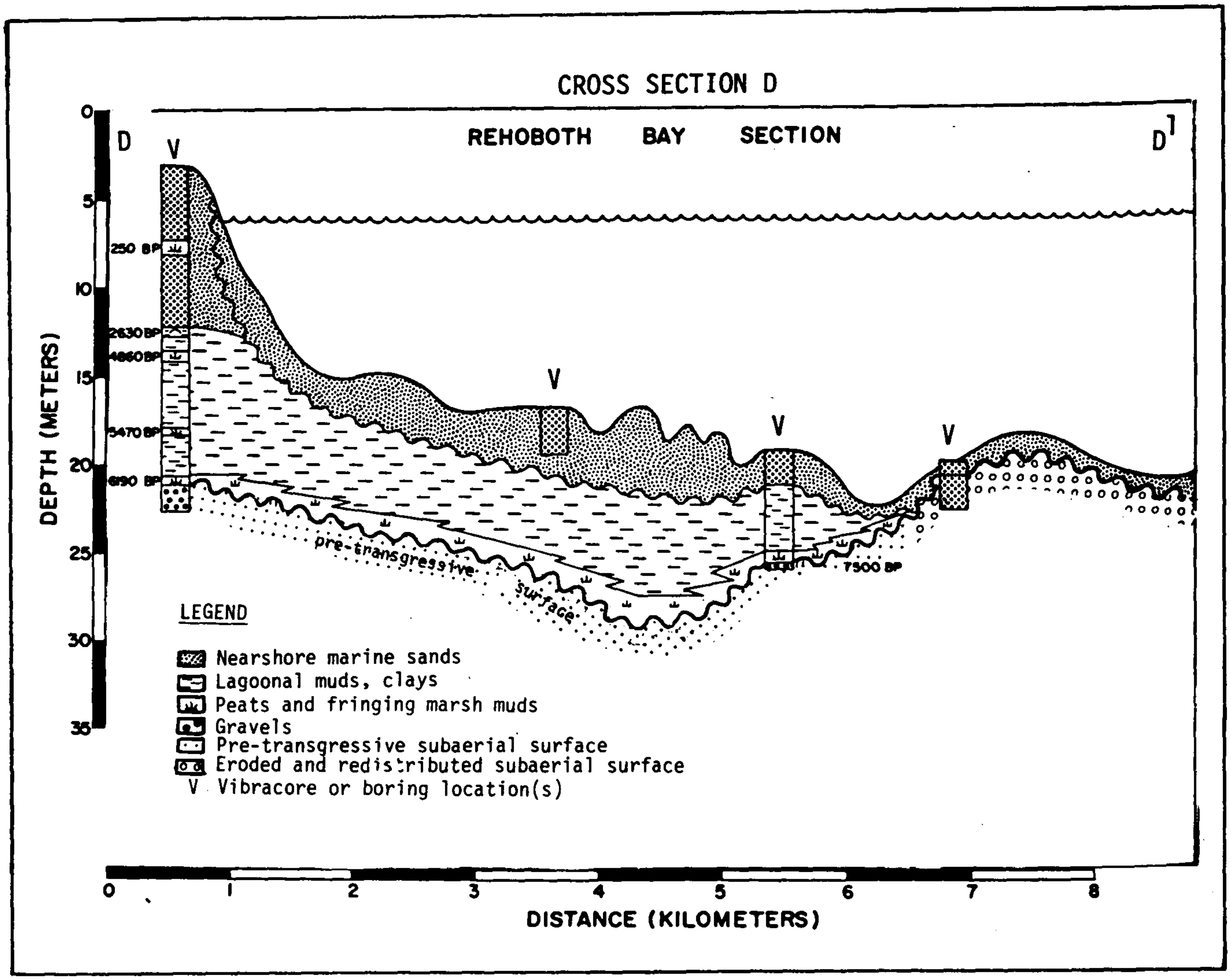


Fig. I-45

Cross section from Rehoboth Bay (Delaware) seaward. Vibracore data suggests that a 6 km wide strip of the pre-transgressive subaerial surface has been preserved beneath transgressive marsh and lagoonal muds. Location of this cross section given in Fig. I-42. Adapted from Sheridan et. al. (1974).

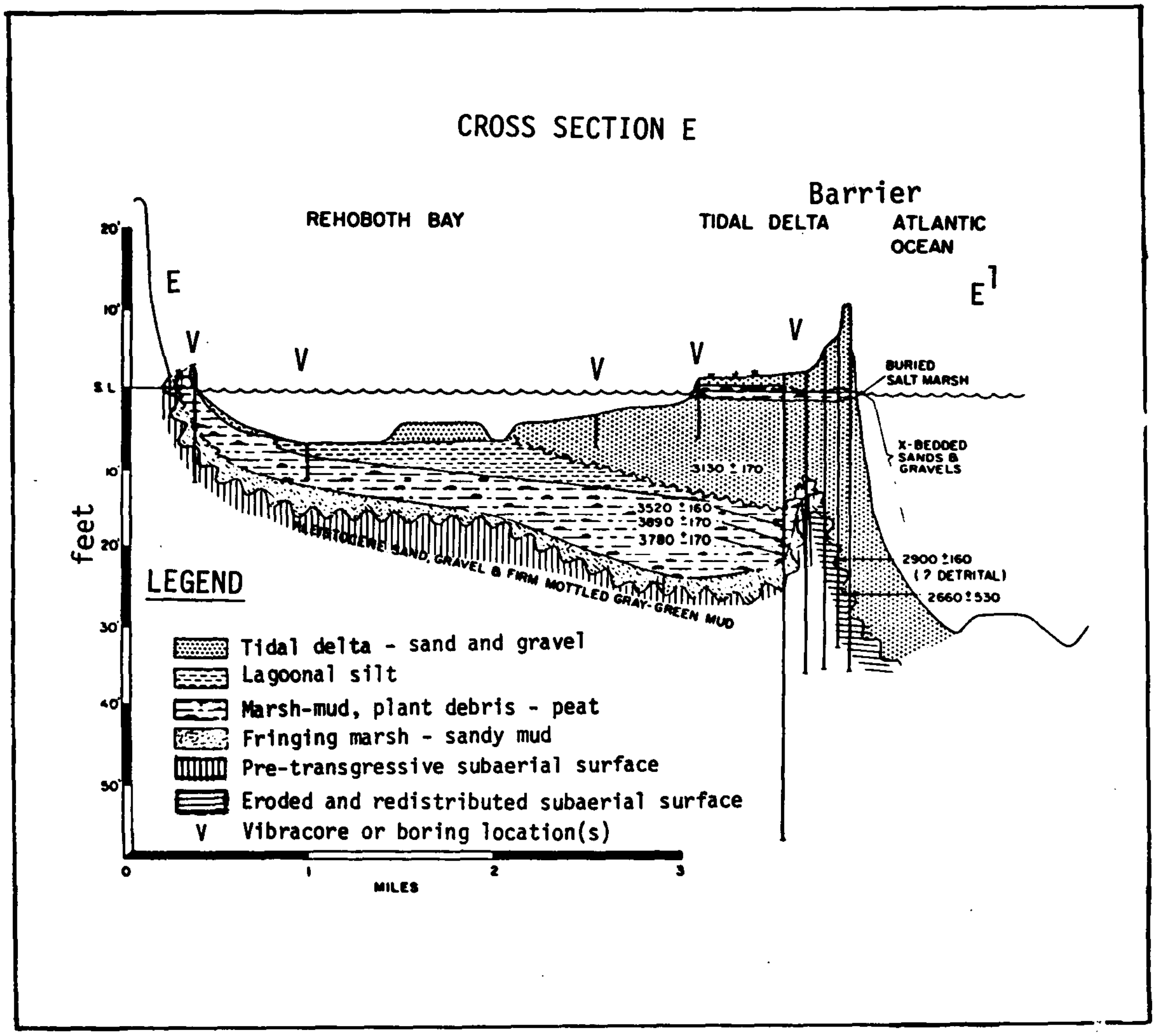


Fig. I-46

Cross section through Rehoboth Bay, Delaware depicting erosion of the transgressive record seaward of the barrier. Depth of erosion probably marked by occurrence of detrital peat "balls" on surface of eroded Late Quaternary sediments. Adapted from Kraft (1971). Position of the buried subaerial surface has been deduced hypothetically. Location of this cross section shown on Fig. I-42.

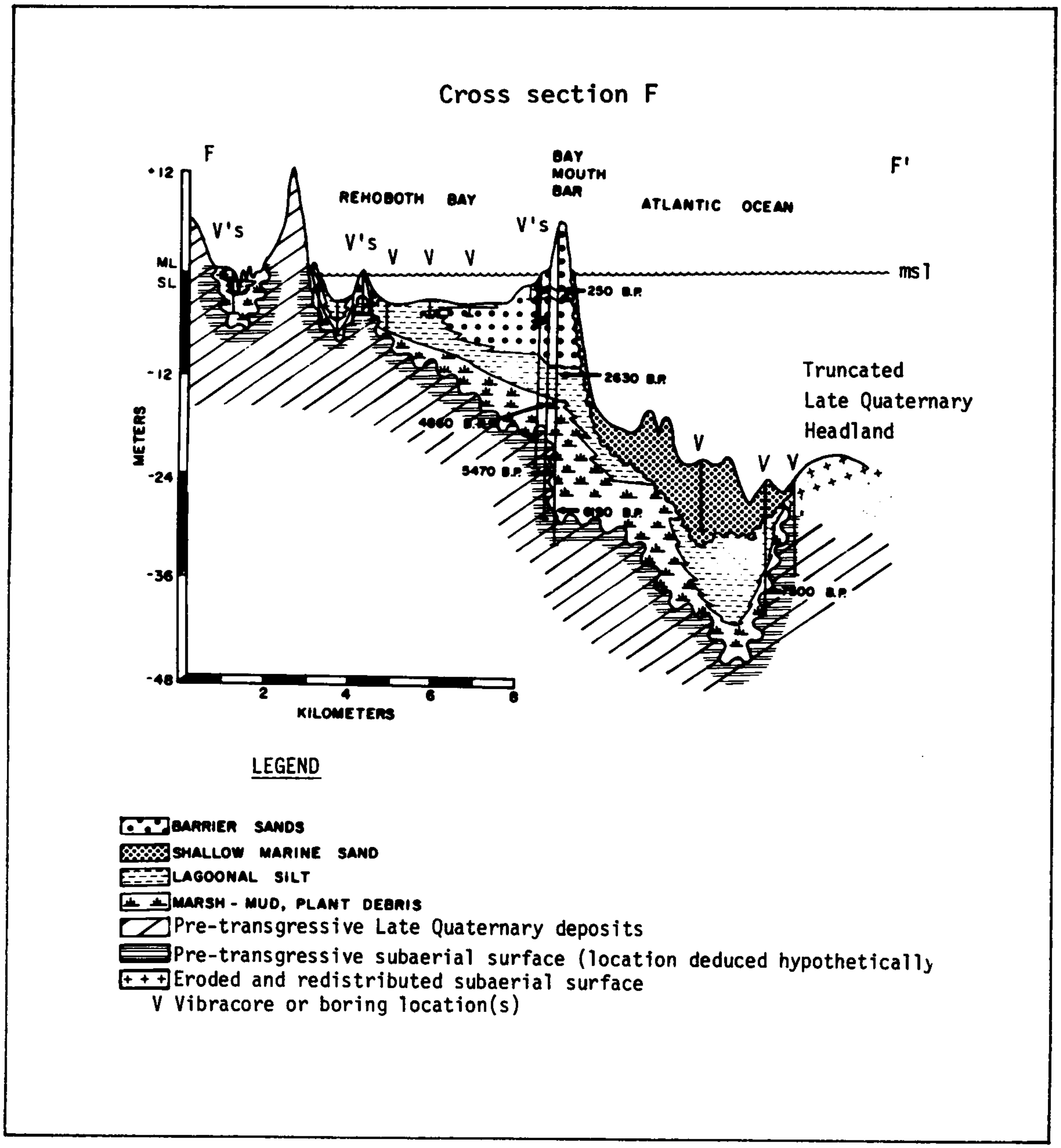


Fig. I-47

Cross section through Rehoboth Bay, Delaware, depicting the units that generally accompany transgression. The location of this cross section is shown on Fig. I-42 . Position of the subaerial surface is hypothetical. Adapted from Kraft (1977).

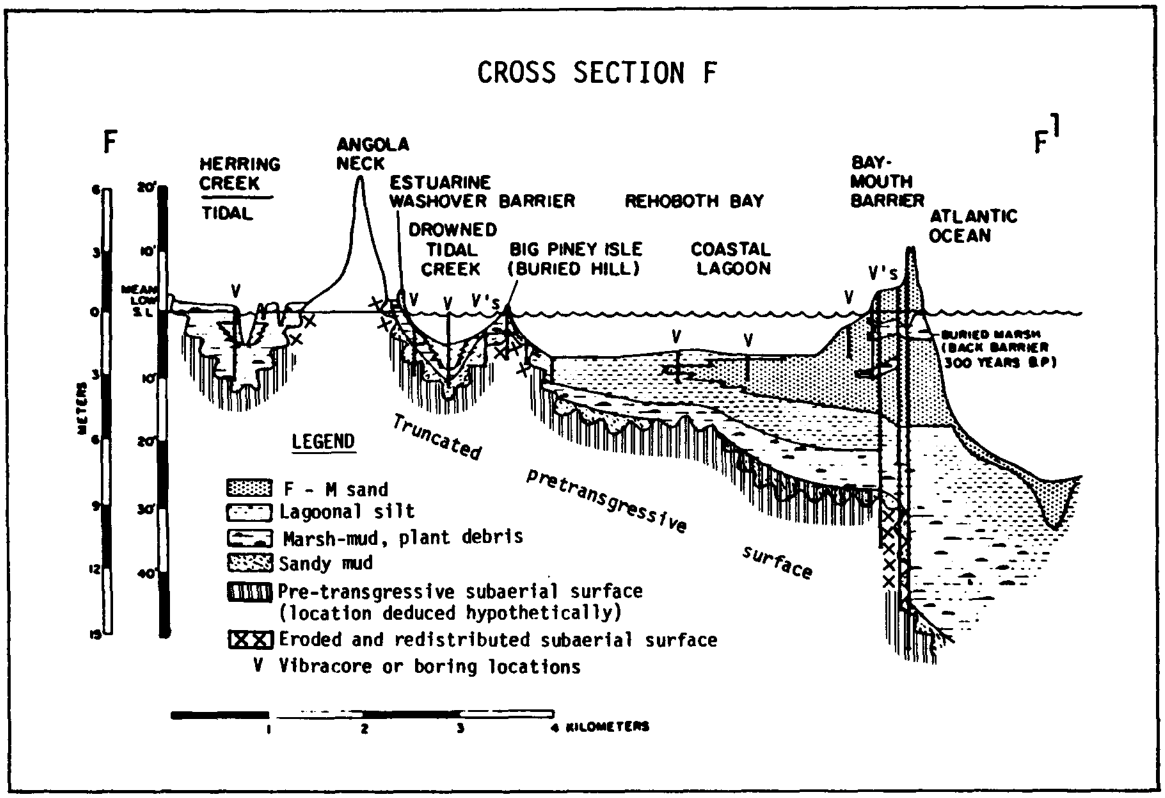


Fig. I-48
Cross section through Rehoboth Bay, Delaware (adapted from Kraft 1971). Preservation of the subaerial surface deduced hypothetically from adjacent units. Location of this cross section given in Fig. I-42.

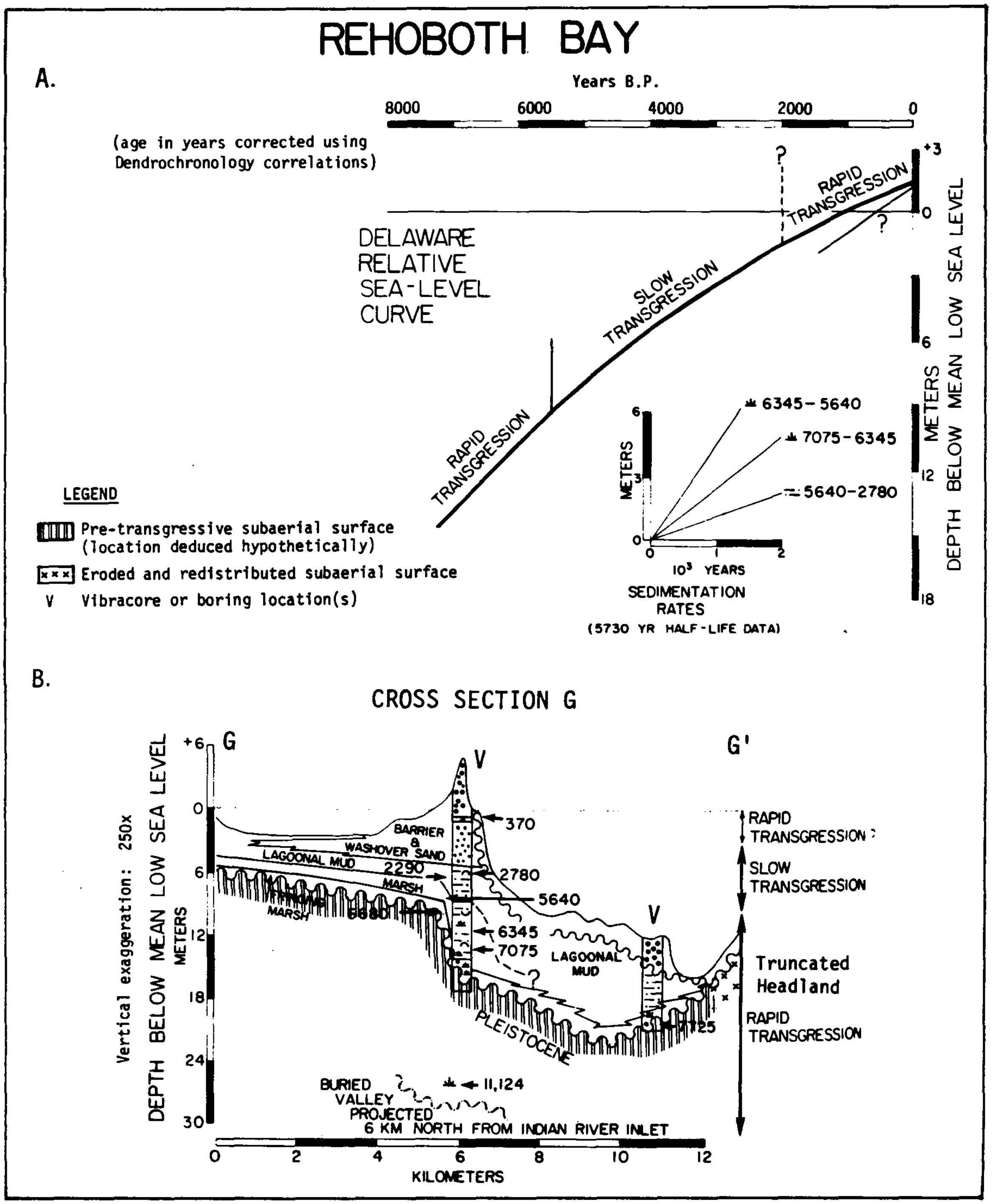


Fig. I-49
Transgression model with derived sedimentation rates (A) for Rehoboth Bay (After Belknap and Kraft 1977). Generalized cross section (B) for the Rehoboth Bay vicinity based on Kraft (1977) and Sheridan et al (1974) and adapted from Belknap and Kraft (1977). Extensive preservation of the pretransgressive surface indicated on the basis of stratigraphy and radiacarbon dates. Location of cross section given in Fig. I-42

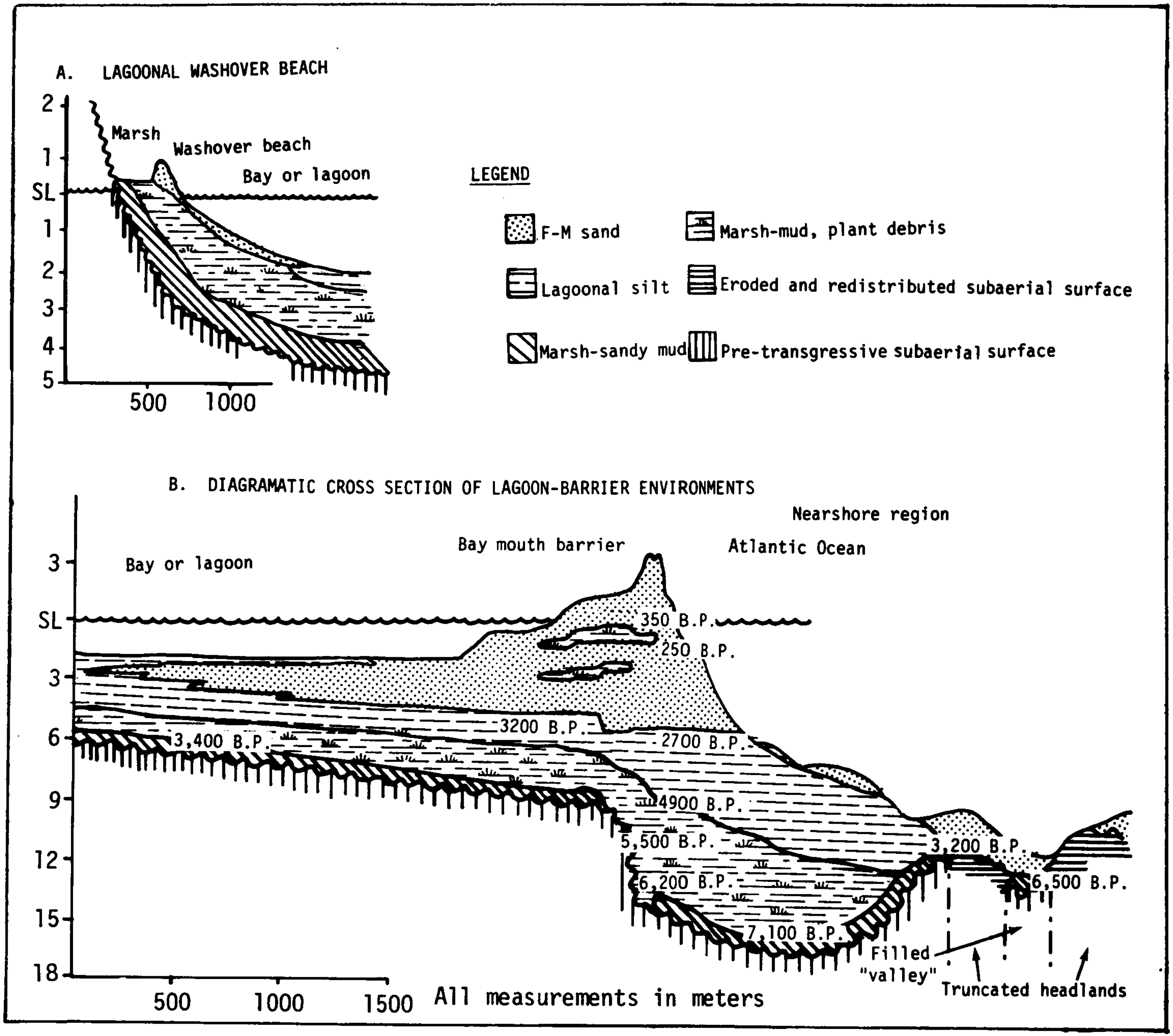


Fig. I-50
Diagrammatic cross section of barrier-lagoonal sequences preserved frequently along coastal Delaware (adapted from Kraft 1971).

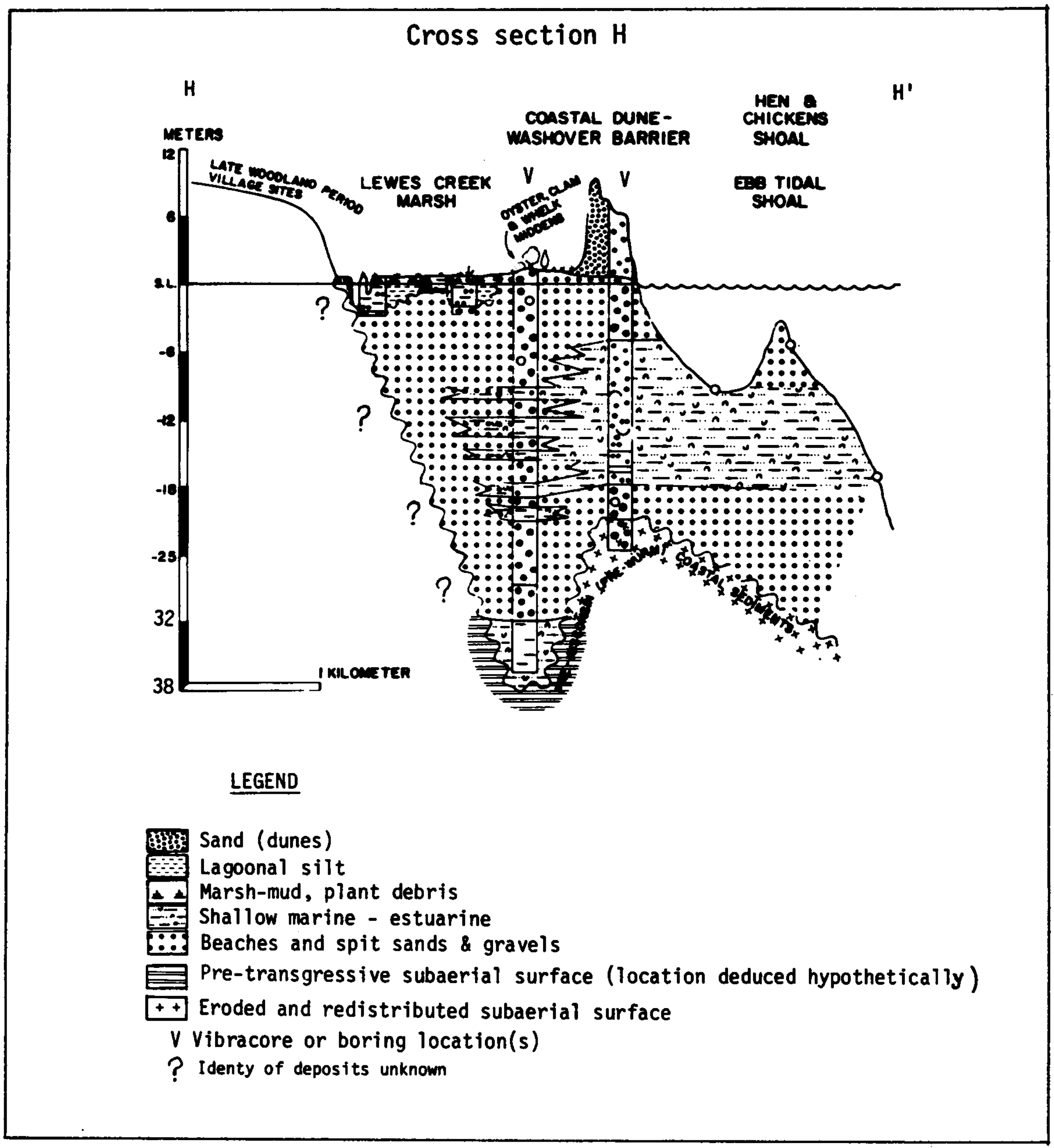


Fig. I-51

Cross section near Lewes Creek, Delaware, showing the stratigraphic units left after the ancestral Cape Henlopen had migrated past this area during the Late Holocene. Lewes Creek Marsh has formed over an area which used to be an open lagoon. Abundant Late Woodland Period shell middens are found along the edges of the marsh and presumably date from the time of the shallow lagoon environment (Kraft 1977). The position of this cross section is shown on Fig. I-42. After Kraft (1977).

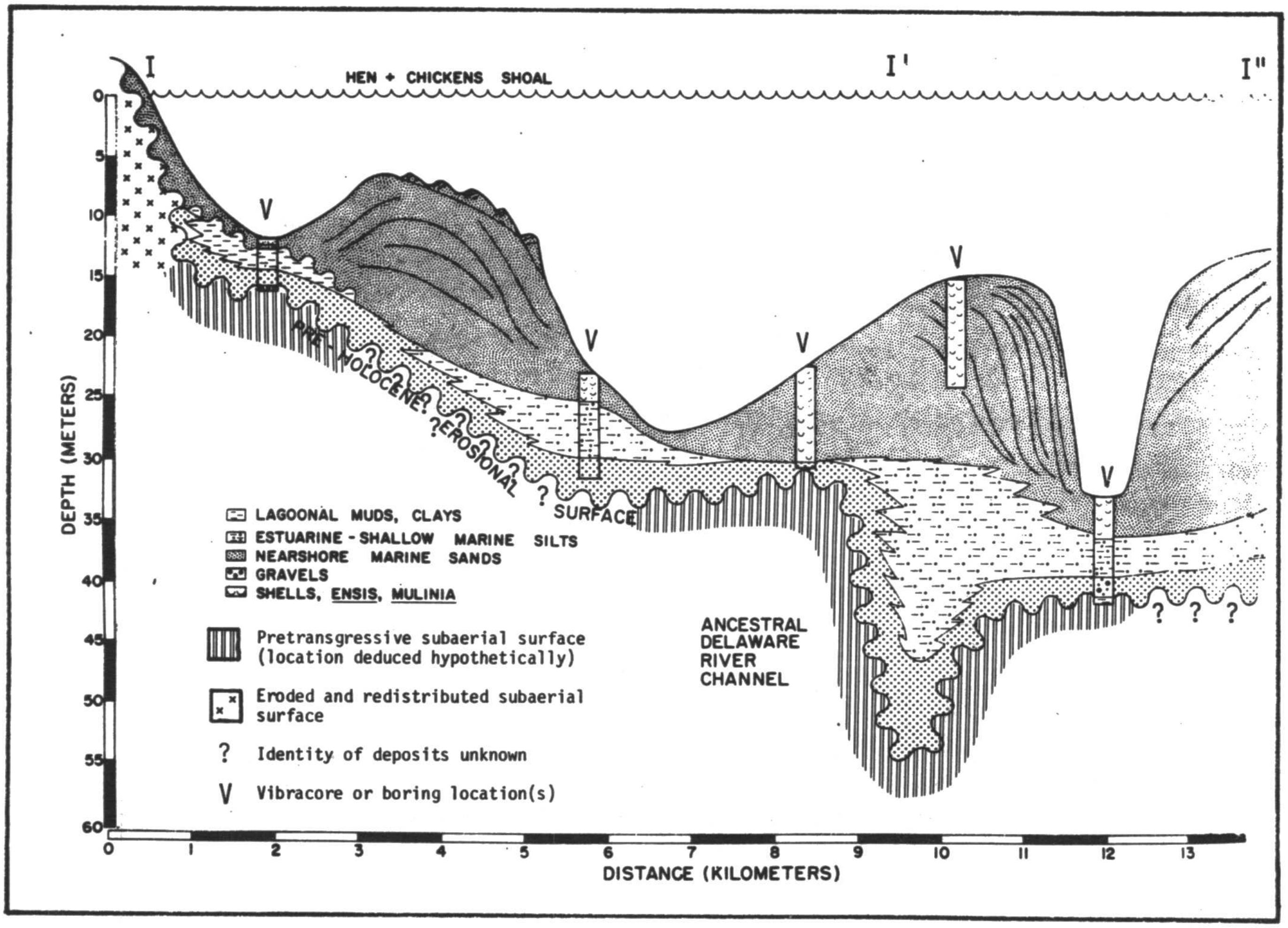


Fig. I-52a
Cross section northeast of Rehoboth Beach, Delaware showing evidence of the ancestral Delaware River Channel. Above the ancestral channel are two flood and ebb-dominated tidal currents. After Sheridan and others (1974).

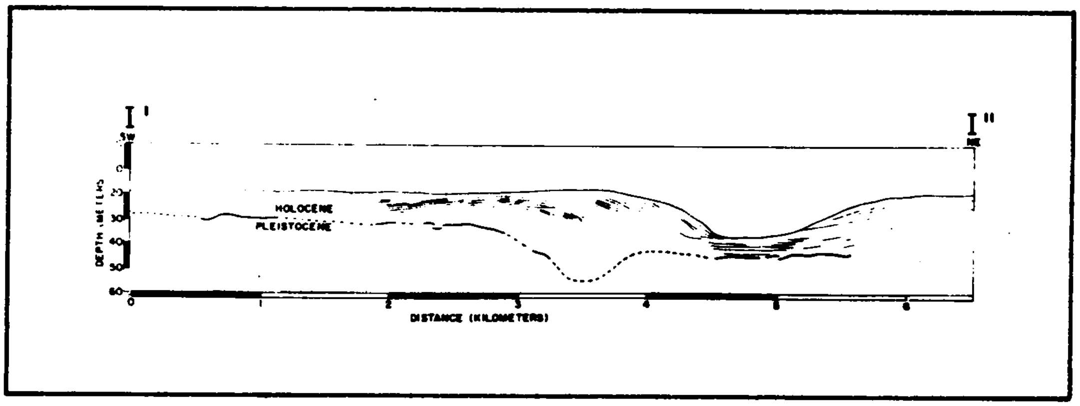


Fig. I-52b

Interpretive tracing of seismic profile along the inner shelf of Delaware, northeast of Rehoboth Beach. Structure of sand shoals near existing channel suggests that it is filling in this depression. After Sheridan and others (1974).

a cross section through this area (section K) showing the lower section of several truncated Pleistocene valleys.

Cape Henlopen has been migrating toward the northwest rather rapidly in recent times. Fig. I-56 shows its positions over the last few hundred years. Fig. I-57 gives several paleogeographic reconstructions of the "Cape Henlopen" region since 7500 B.P. (Kraft 1977; Kraft and others 1978).

The Cape Henlopen Spit complex consists of a variety of morphological features. Kraft and others (1978) identify dunes, swamps, marshes, recurved spit tips, beach accretion plain, and beach and berm regions. Unfortunately, erosional shoreface retreat removes these subaerial surfaces as the spit migrates northwestward. Only the lower portions of deeply incised Pleistocene-Early Holocene river valleys escape this process, as outlined in our model. Consequently, it is of little use to devote space to spit complexes, since only their basal (submarine) sequences are capable of being preserved in the sedimentary record. Needless to say, the depositional environment of the basal units was not subaerial but instead nearshore marine.

Fig. I-58 gives a north-south-trending cross section from Cape Henlopen to Bethany Beach. The moderately incised pre-transgressive topography and the prograding shoreline at Cape Henlopen are 2 characteristics common to this area. Similar environments during the Holocene may have existed on the north side of the ancestral Delaware River valley as well as adjacent to other large river systems.

Kraft (1977) and Kraft and others (1974) have extended their work into Delaware Bay and offer a model for making paleogeographical reconstructions. Fig. I-40 gives the location of the cross sections along Delaware Bay. Kraft (1977) considers the present-day geomorphic environments found along middle and upper Delaware Bay similar to types which used to exist on the Inner and Middle Shelves. Environments found by moving northward along Delaware Bay are similar to older environments which used to exist further south-southeast (seaward) when sea level was lower. Moving north-northwestward along the Bay is similar to moving backwards in time at the mouth of the Bay (Kraft and others 1974). Fig. I-59a and b illustrates this concept and shows several reconstructed profiles and their present-day counterparts (Kraft and others 1974).

Tributaries along the present Delaware Bay can also be used to infer the Middle to Late Holocene evolution of the estuary system. The environments of tributaries during the Early Holocene would have been somewhat different because their valleys were moderately incised during the Late Pleistocene lowstand. Along the Middle and Outer Shelves, rivers and streams would have had much greater gradients before transgression began than those found along the partially submerged tributaries of Delaware Bay today. There would have been moderately incised stream valleys, flanked by remnants of terraces developed during the preceding marine regression. Today, most of Delaware Bay's tributaries exhibit

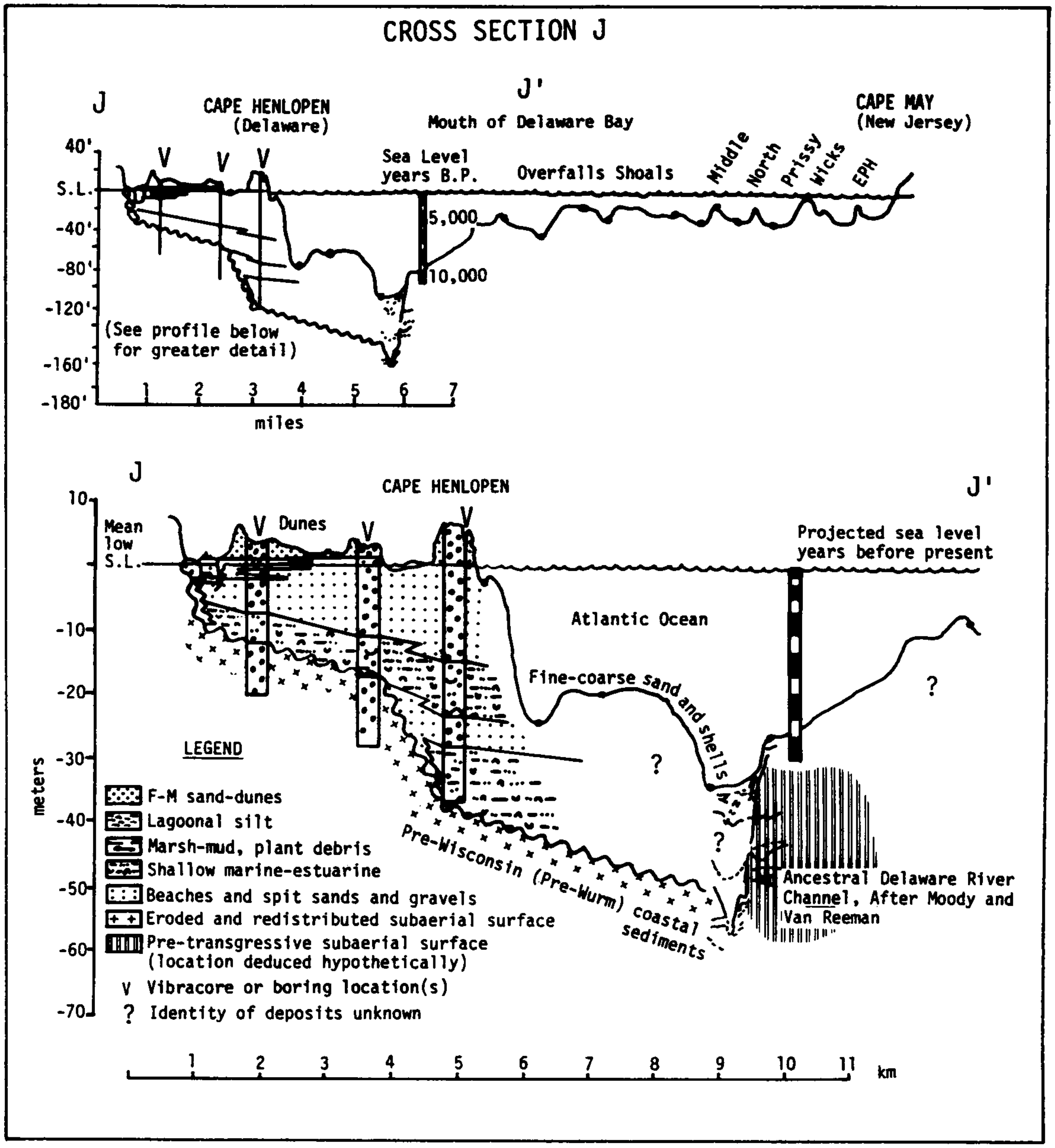


Fig. I-53

Cross section from Cape Henlopen, Delaware across the mouth of Delaware Bay showing the presence of at least one buried river valley. Adapted from Kraft (1971).

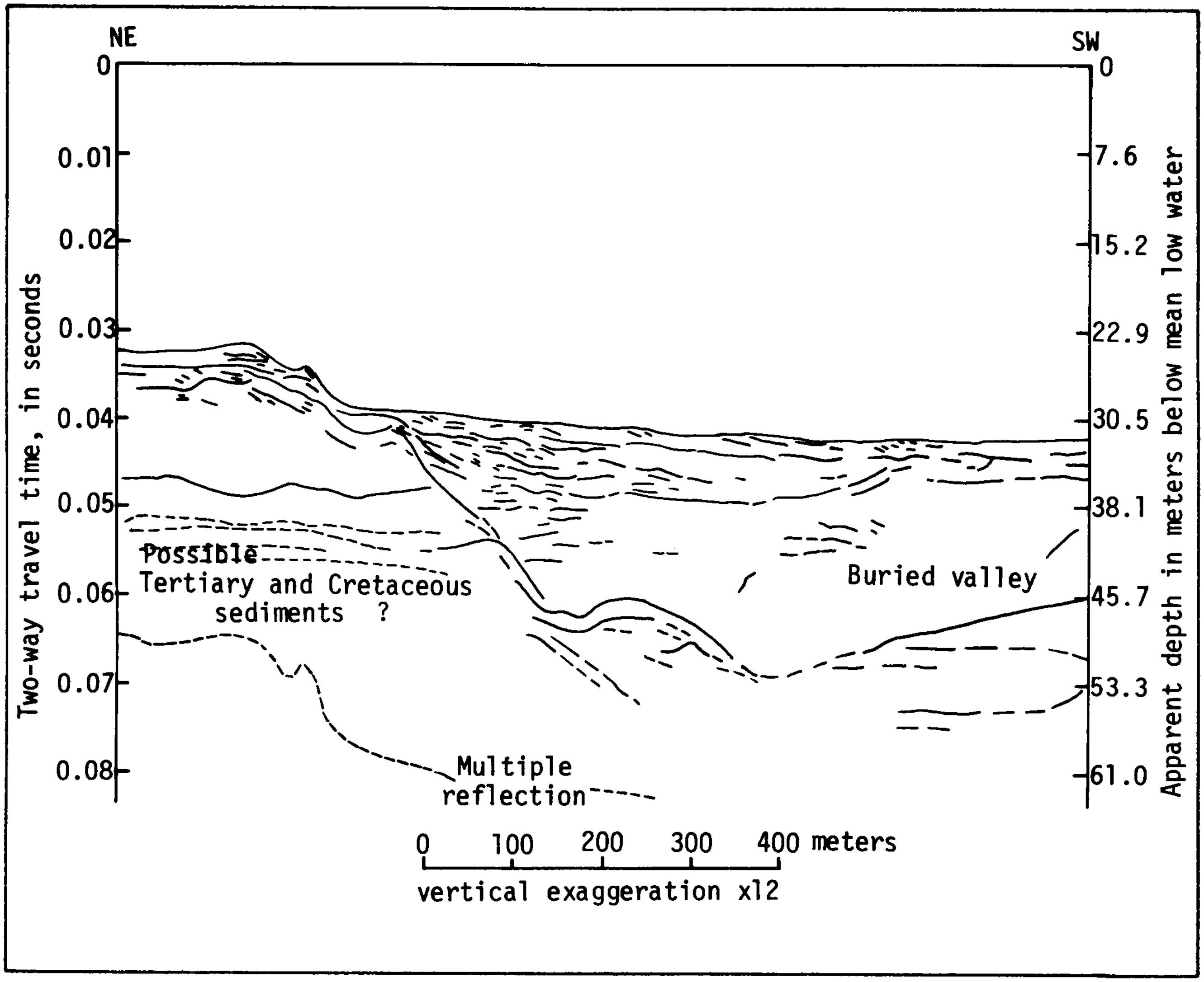


Fig. I-54

Line drawing of seismec profile from a section of the mouth of Delaware Bay showing a buried valley probably belonging to the late Wisconsin Delaware River. After Moody and Van Reenan (1967).

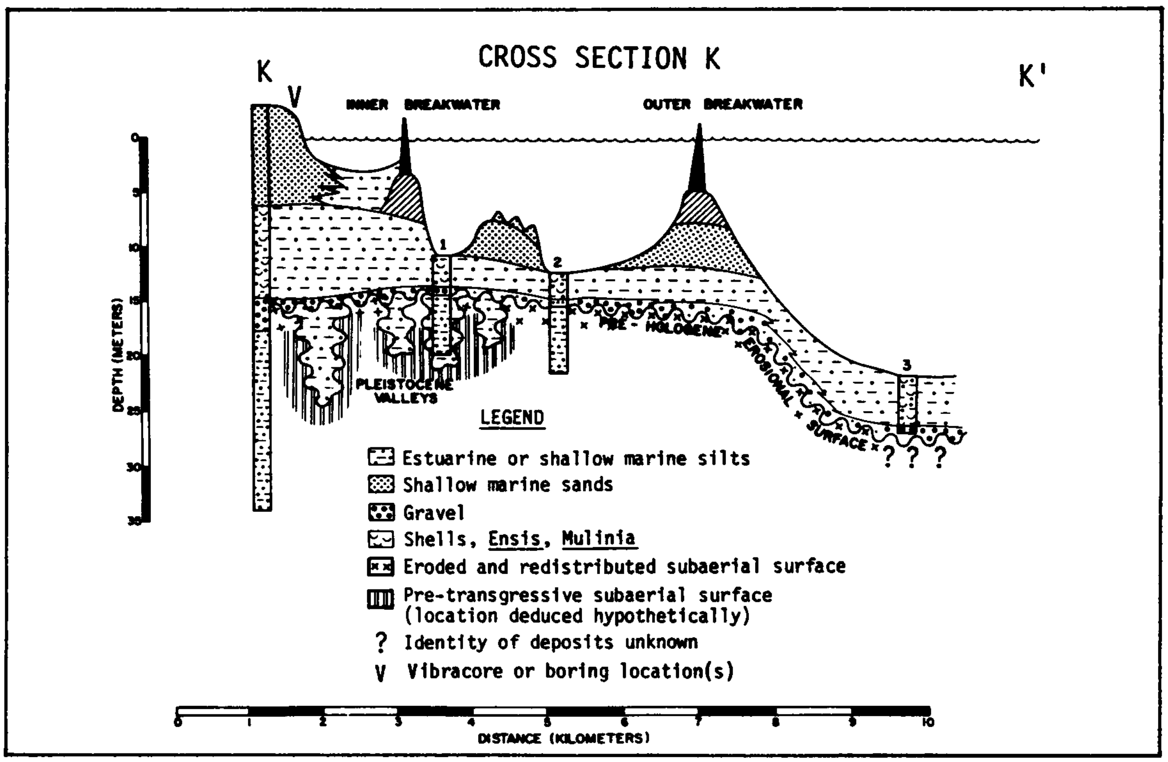


Fig. I-55
Cross section north of Cape Henlopen, Delaware showing several buried pretransgressive stream valleys. After Sheridan and others (1974).

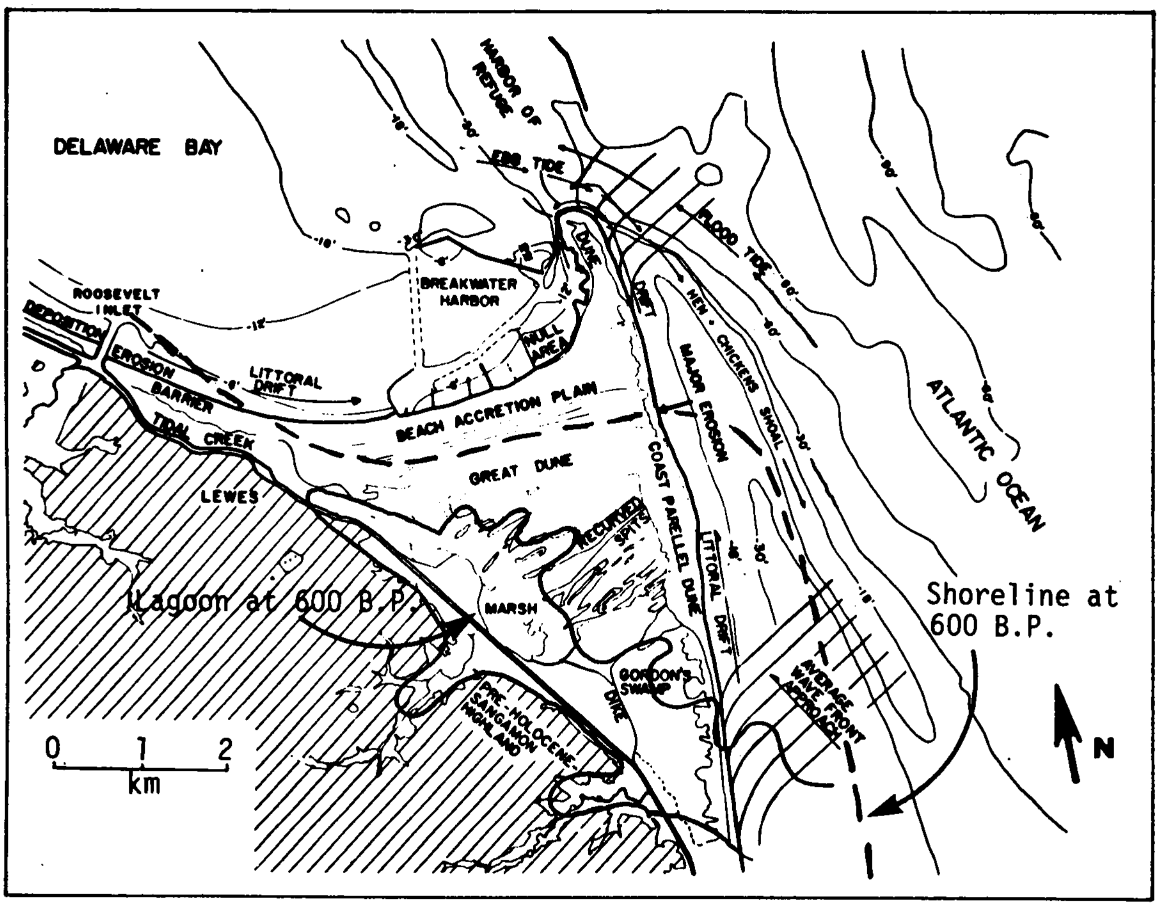


Fig. I-56

Geomorphic elements of the Cape Henlopen spit complex and some of the processes eroding and redistributing sediment. An interpretation of the shoreline for this area at about 600 B.P. is also illustrated based on Kraft (1971). As erosion and transgression continue along this section of the coast, the subaerial surface of this spit will be destroyed. After Kraft and others (1978).

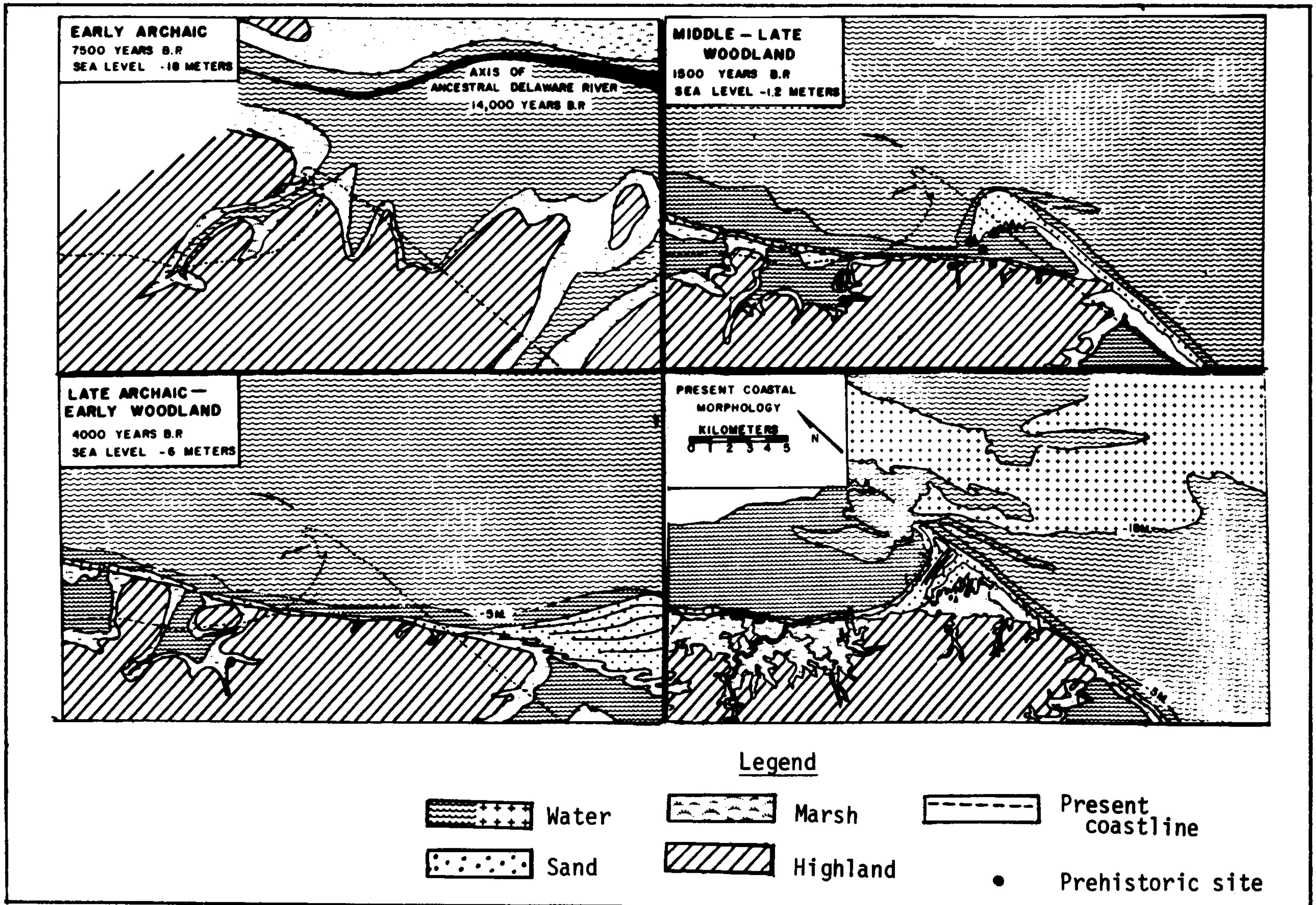


Fig. I-57 Paleogeographic reconstructions of the Cape Henlopen region. The path of the ancestral Delaware River at approximately 14,000 B.P. is shown. At 7,500 B.P. several headlands and bays existed in the Cape Henlopen region. The dashed line indicates the approximate position of the present coastline. Dots indicate the position of prehistoric sites associated with each period. After Kraft (1977).

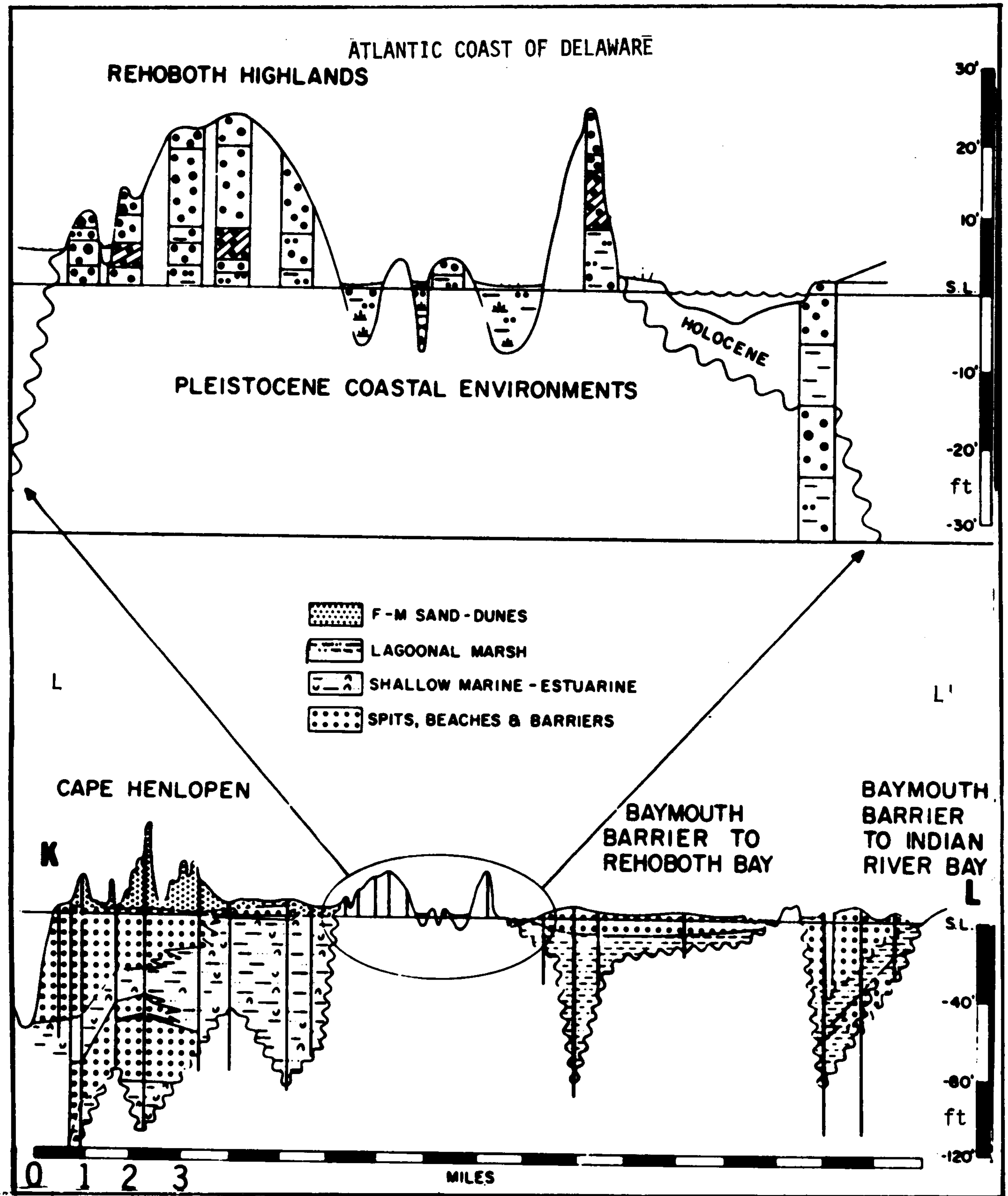


Fig. I-58 Interpretive cross section of the Atlantic Coast of Delaware showing Holocene coastal environments overlying incised pre-transgressive topography. After Kraft (1971).

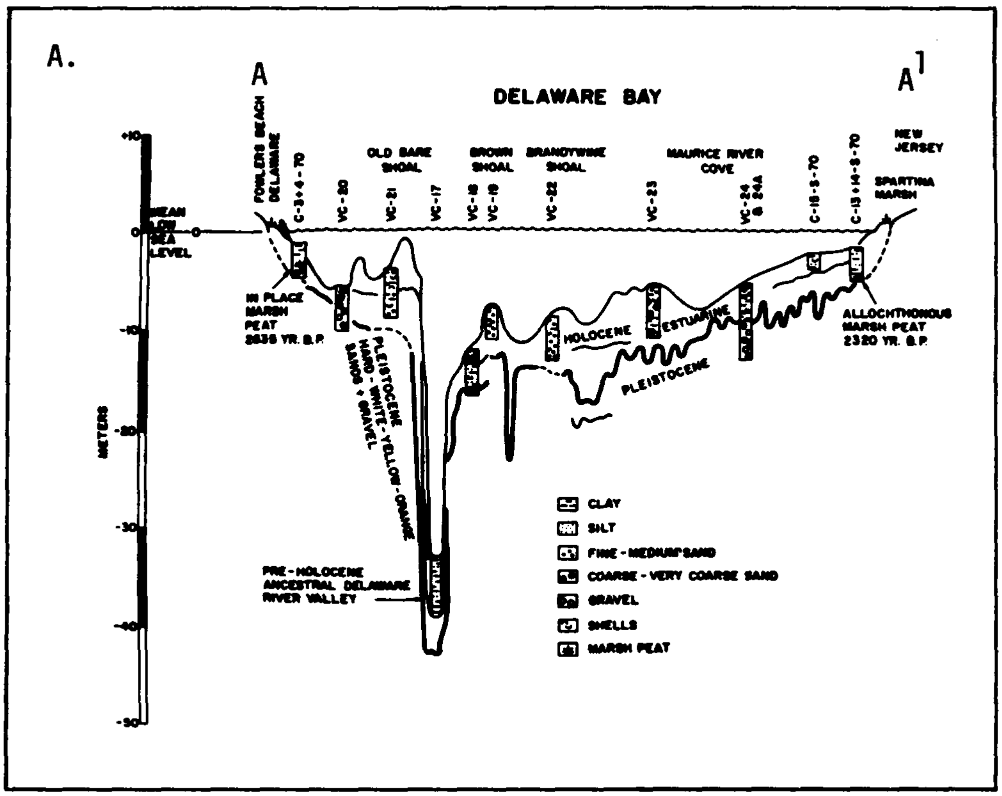


Fig. I-59

Cross sections and reconstructions of Delaware Bay based on the work of Kraft (1977) and Kraft et. al. (1974). (A) Cross section of Delaware Bay showing the amount of Holocene sediment in fill, based on seismic and core data (Kraft 1977). (B) Evolution of Delaware Bay during the mid- and late-Holocene based on seismic profiles, cores, and carbon-14 dates (Kraft et. al. 1974). (C) Reconstructed cross sections showing the ancestral Delaware Bay/River during the mid- and late-Holocene. Sea-level positions based on Belknap and Kraft (1977). The reconstructed cross sections are correlated with environments found along Delaware Bay today. They illustrate the concept that a fixed location along a river valley evolves through stages comparable to a northward movement along Delaware Bay today as transgression takes place. (A) and (C) after Kraft (1977). (B) after Swift (1976a).

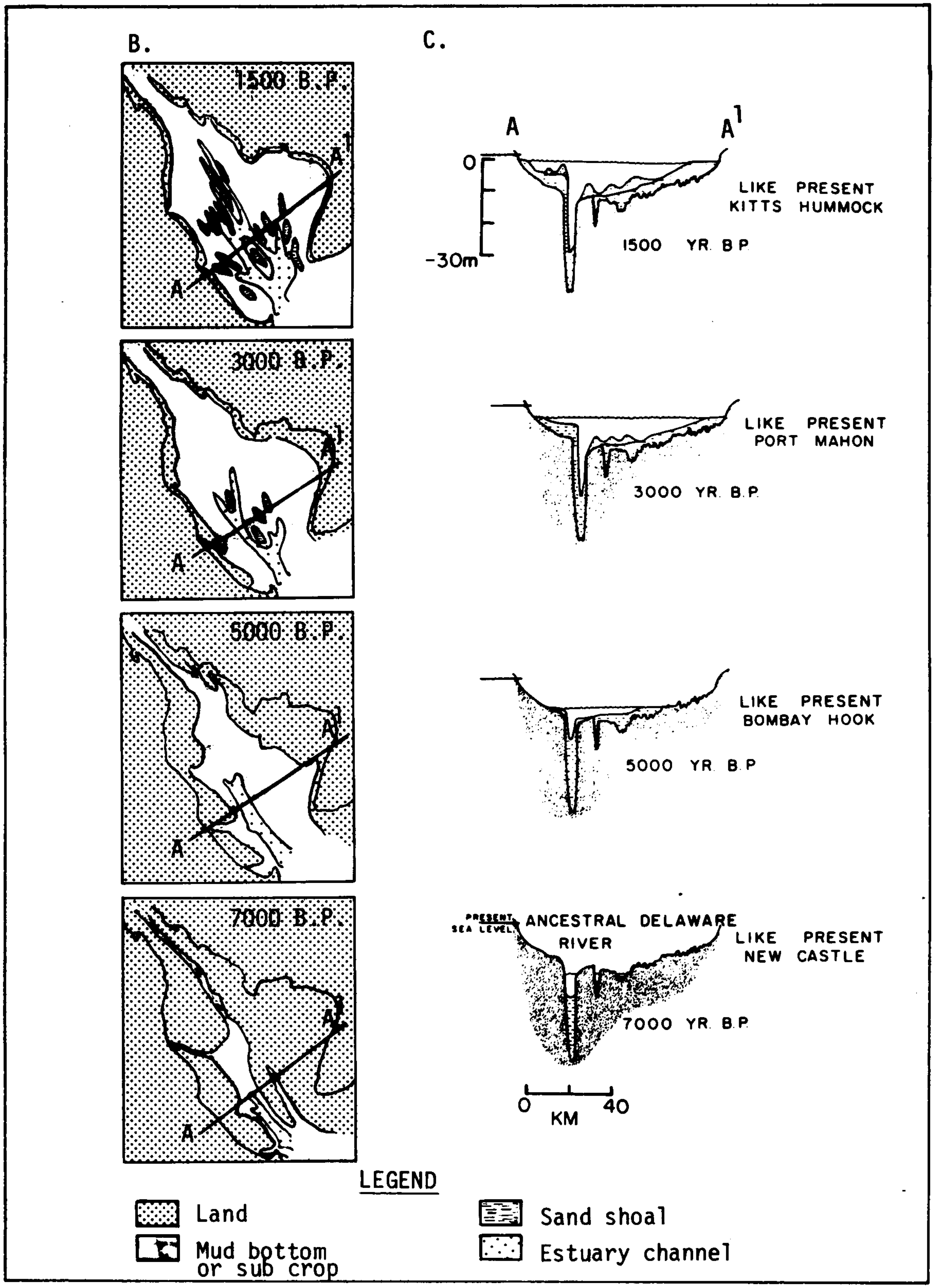


Fig. I-59 continued

quite different features. They consist of broad marsh-filled valleys in which seemingly undersized streams meander toward the Bay (Kraft 1977).

The development of Delaware Bay and the infilling of its tributaries have been investigated in detail (Kraft 1971, 1977; Kraft and others 1974). Figs. I-60 through I-63 illustrate the development of three regions along the Bay. In the Island Field area, marshes appeared as early as 10,000 years ago as a result of the tidal influence of the ancestral Delaware River (Fig. I-60). A similar situation existed along the Appoquinimink River section of the ancestral Delaware River at 10,000 B.P. (Figs. I-63 and I-65). The ancestral tidal Delaware, however, did not reach the Holly Oak region until more recently, and tidal marshes were nonexistant in the area before 6000 B.P.

As marine transgression proceeded up the Delaware River system, infilling took place along each estuary. The general sequence was that of marshes rising to cover the adjacent slopes and terraces, followed by tidal mudflats and the intrusion of stretching fingers of Delaware Bay down each of the adjacent tributaries. A cross section (Fig. I-61) through the Murderkill River Valley illustrates the extent of infilling found along the central portion of Delaware. Over 27 m of marsh and estuarine sediments have accumulated at the mouth of the present-day Murderkill River, spanning over 10,000 years (Fig. I-61).

The unconformity encountered at the base of the Holocene transgressive sequence represents the gradient of the ancestral Murderkill River between about 12,000 and 6000 B.P. At the earlier time the gradient would have been somewhat steeper than that shown on Fig. I-61 (greater than 1.6 m per km). As transgression reached the vicinity of drill hole DH 2-71 (Fig. I-61) downcutting soon halted and infilling proceeded along the central axis of the valley. Beveling, of course, also came into operation during this sequence of events. Fig. I-61 shows the beveling by the Delaware Bay systems in recent times in the vicinity of A^1 (eastern end). In other areas, sufficient sediment cover (about 25-30 m) is present to protect the buried pre-transgressive subaerial surface from erosion by the encroaching shoreline of Delaware Bay. Evidence of exploitation of the tidal marsh surface during the last few thousand years, however, is destroyed in some areas as the upper few meters of the Murderkill River tidal marsh are eroded by either the Delaware estuary or the Murderkill River (Belknap and Kraft 1977).

The northern section of the tidal Delaware River abuts the piedmont. In this area, the incision of Delaware River tributaries was greatly restricted by earlier channels cut into the resistant crystalline bedrock. Tidal marsh and estuaries are severly limited in this area (that is, northwest of the Delaware River). On the southeast side of the Delaware River, emerged coastal plain topography provided larger marshlands and estuaries and allowed them to extend further inland. Kraft (1977) takes into account these environmental differences in his paleogeographic reconstruction of the Holly Oak area in Fig. I-62.

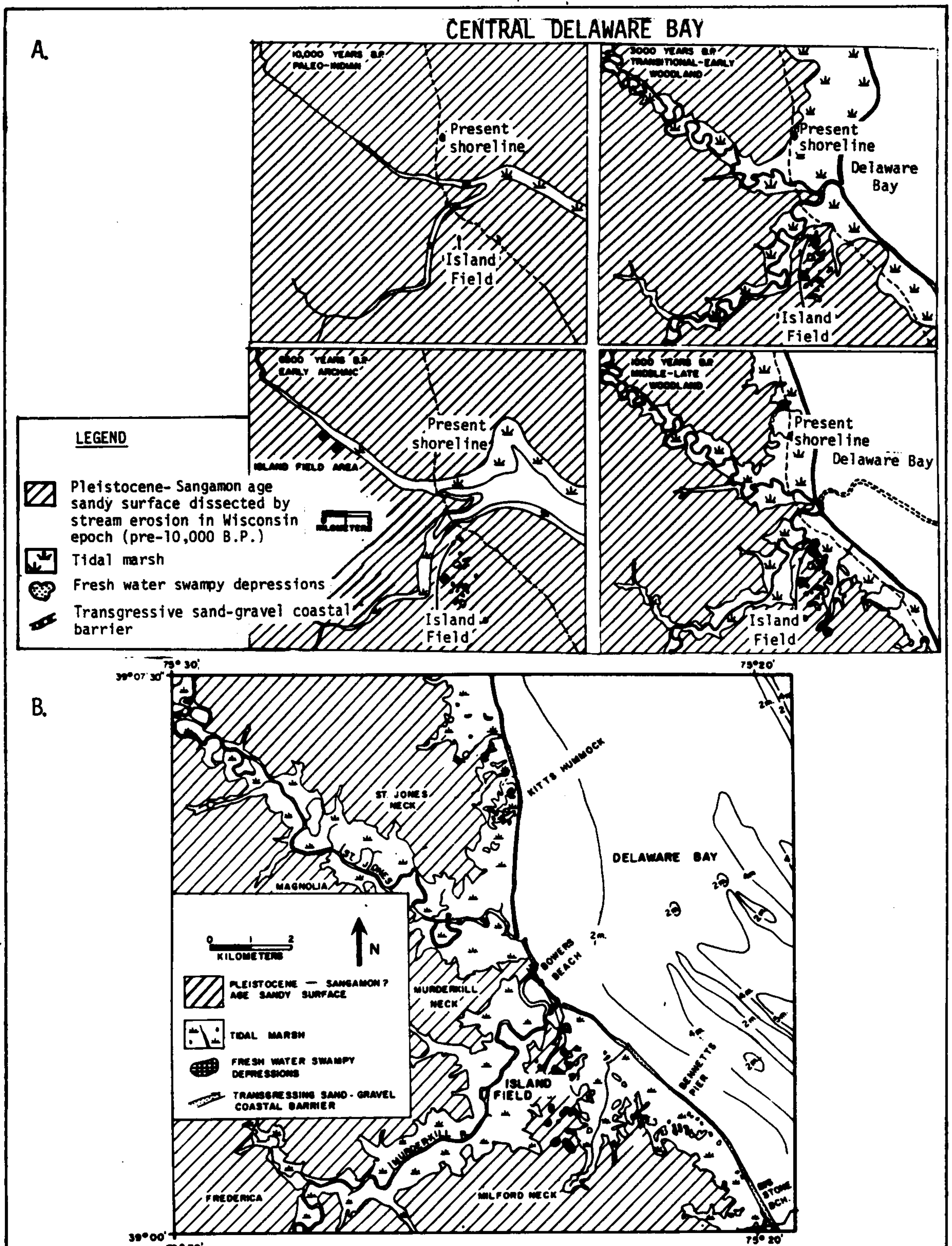


Fig. I-60 Paleogeographic reconstructions of the Murderkill-St. Jones River systems in Late Woodland Period (1,000 B.P.), Early Woodland Period (3,000 B.P.), Archaic Period (6,500 B.P.), and Paleo-Indian times (10,000 B.P.) are given in Fig. A. Dots indicate locations of prehistoric sites for the periods shown. The present geomorphology of the area is shown in Fig. B. After Kraft (1977).

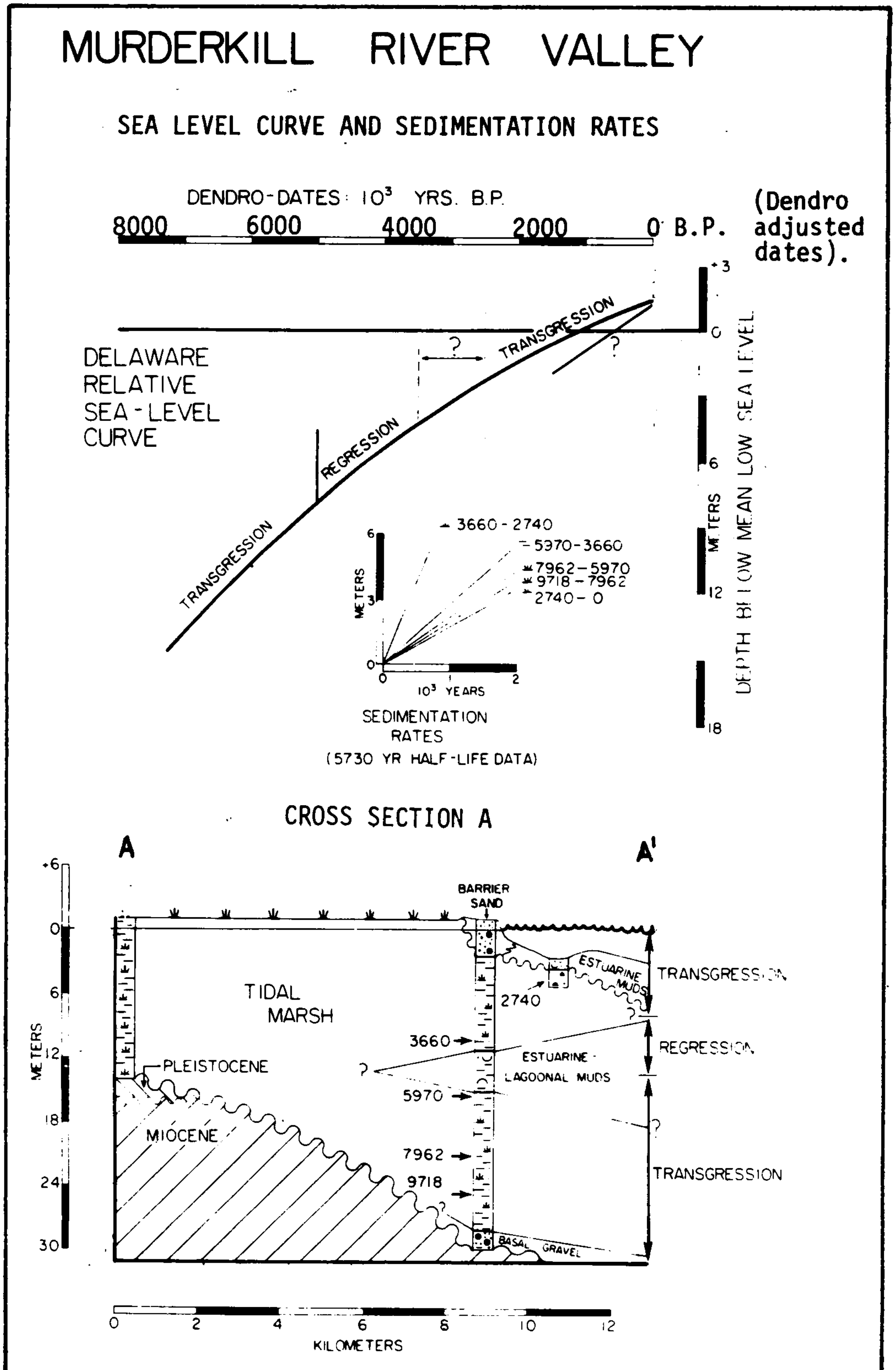


Fig. I-61 Cross section, local sea-level curve, and sedimentation rates for the Murderkill River Valley along Central Delaware Bay. After Belknap and Kraft (1977).

So far, the discussion has focused on evidence concerning environments along the inland portion of Delaware Bay. Some information concerning the path of the ancestral Delaware River on the Continental Shelf may be found in the literature (for example, Cousins and others 1977; Kraft 1971; Swift 1973, 1976b; Swift and others 1972; and Twichell and others 1977).

Swift (1973) has identified the Late Pleistocene-Holocene path of the Delaware River Valley by means of shelf bathymetry. In the same place, he notes that shelf highs and lows seaward of the present mouth of Delaware Bay are basically the result of estuary, nearshore, and mid-shelf submarine processes which have modified the truncated pre-transgressive surface. The Holocene Delaware River Valley provided a path for the ancestral Delaware Bay as the sea level rose. Estuary sedimentation filled in large sections of the valley as water invaded them. Tidal currents have played an important role to reshape the estuary bottom. Sedimentation associated with littoral drift-deposition centers flanking the mouth of the bay also have changed the bathymetry in the area. These processes eventually produced the bottom configuration found seaward of Delaware Bay today. Storm and tidal currents still affect some bottom features and over the next few centuries the near-shore features will be changed significantly by new shoal configurations and sand ridges.

Fig. I-64 shows several components of the Delaware Shelf Valley identified by Swift (1973). Between the 40 m scarp and the 60 m scarp, a large cuspate delta has been identified from shelf bathymetry (Swift 1973). As the ocean transgressed this delta, it probably evolved into a broad lagoon behind a barrier chain capped by a cuspate foreland (Swift 1973). Continued transgression and erosional shoreface retreat eventually pushed the barrier chain and cuspate foreland landward, leaving behind a series of shoals as evidence of littoral drift convergence centers (Swift 1973; Swift and others 1972). After sea level had risen above the 40 m scarp, an ancestral Delaware Bay formed and an estuary environment extended a considerable distance inland.

By applying local sea-level data (Belknap and Kraft 1977; Dillon and Oldale 1978) to data on the evolution of the Late Pleistocene-Holocene Delaware River (Swift 1973; Swift and others 1972; Swift and Sears 1974), one may deduce that the following environments probably existed along this area. At 18,000 B.P. the ancestral Delaware River flowed across the Shelf and entered the open ocean somewhere near 38°N 74°W. Its course was buried during the Early Holocene and its exact location remains unknown. From 18,000 B.P. to about 13,000 B.P., rapid transgression occurred. The Delaware Valley on the Outer Shelf during this period was V-shaped and narrow with a much greater seaward grade than the section that flowed over the Middle and Inner Shelves. Transgression filled in the subaerial valley making it indistinguishable from the adjacent shelf floor.

After about 13,000 B.P., when sea level reached the 60 m scarp, there was a broad delta along the Middle Shelf where the Delaware entered the ocean

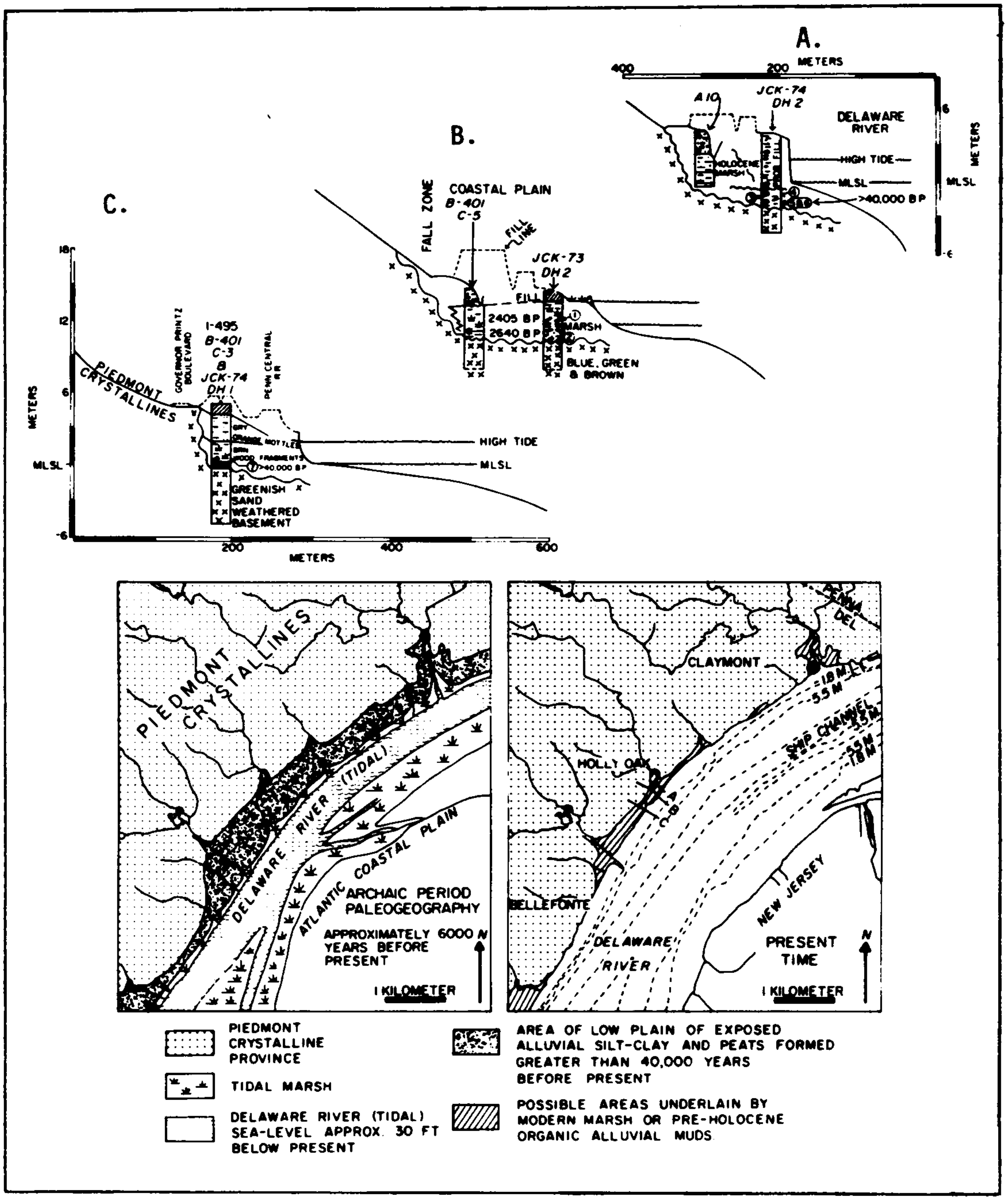


Fig. I-62

Paleogeographic reconstruction for 6,000 B.P. of a section of the Delaware River, Delaware, near the Pennsylvania border. Locations of cross sections illustrated above are shown near Holly Oak. The cross sections indicate that this late-Holocene sediment covers much older (Sangamon?) deposits. After Kraft (1977).

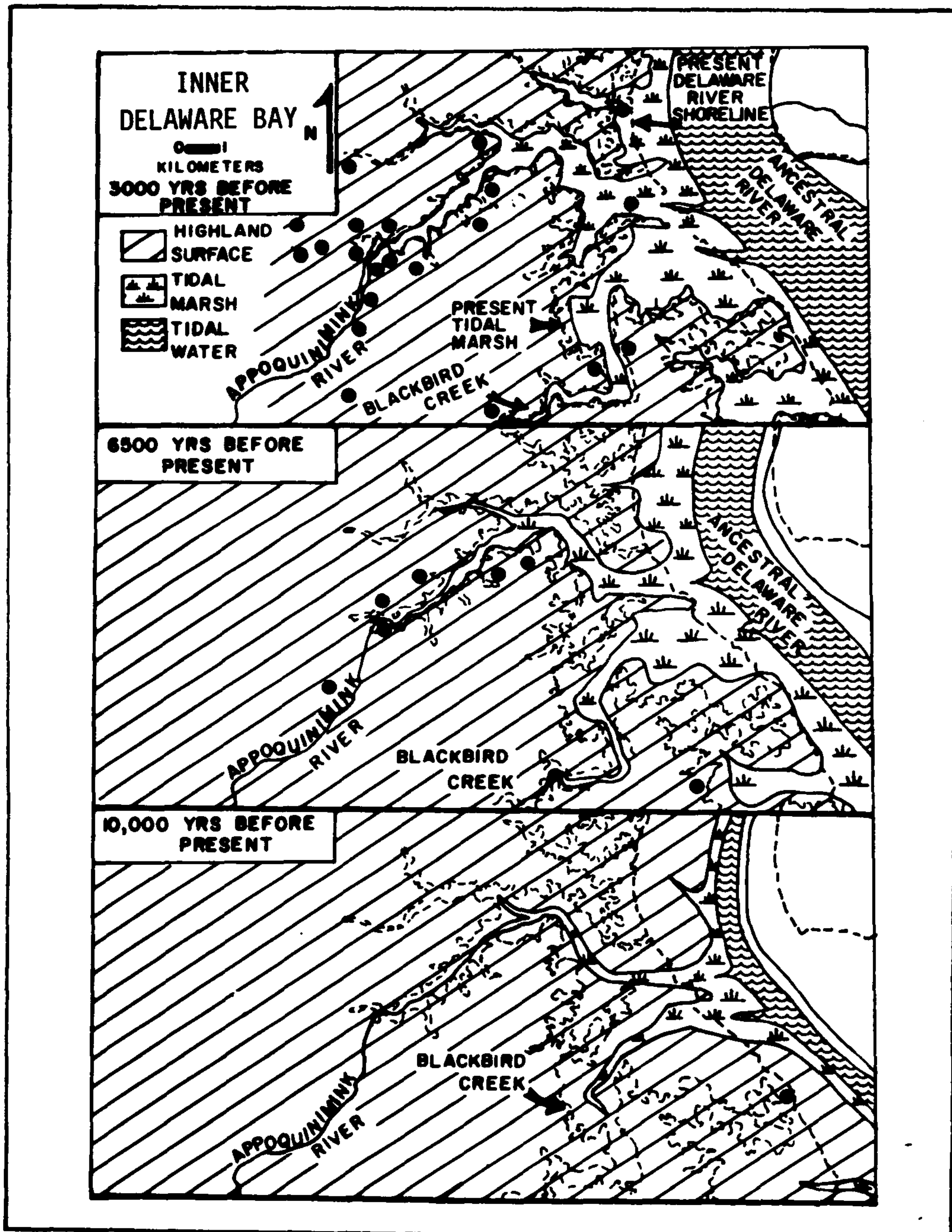


Fig. I-65
Paleographic reconstructions of a section of inner Delaware Bay, Delaware, at 3 periods during the Holocene. Dots indicate the location of prehistoric sites associated with each time period.

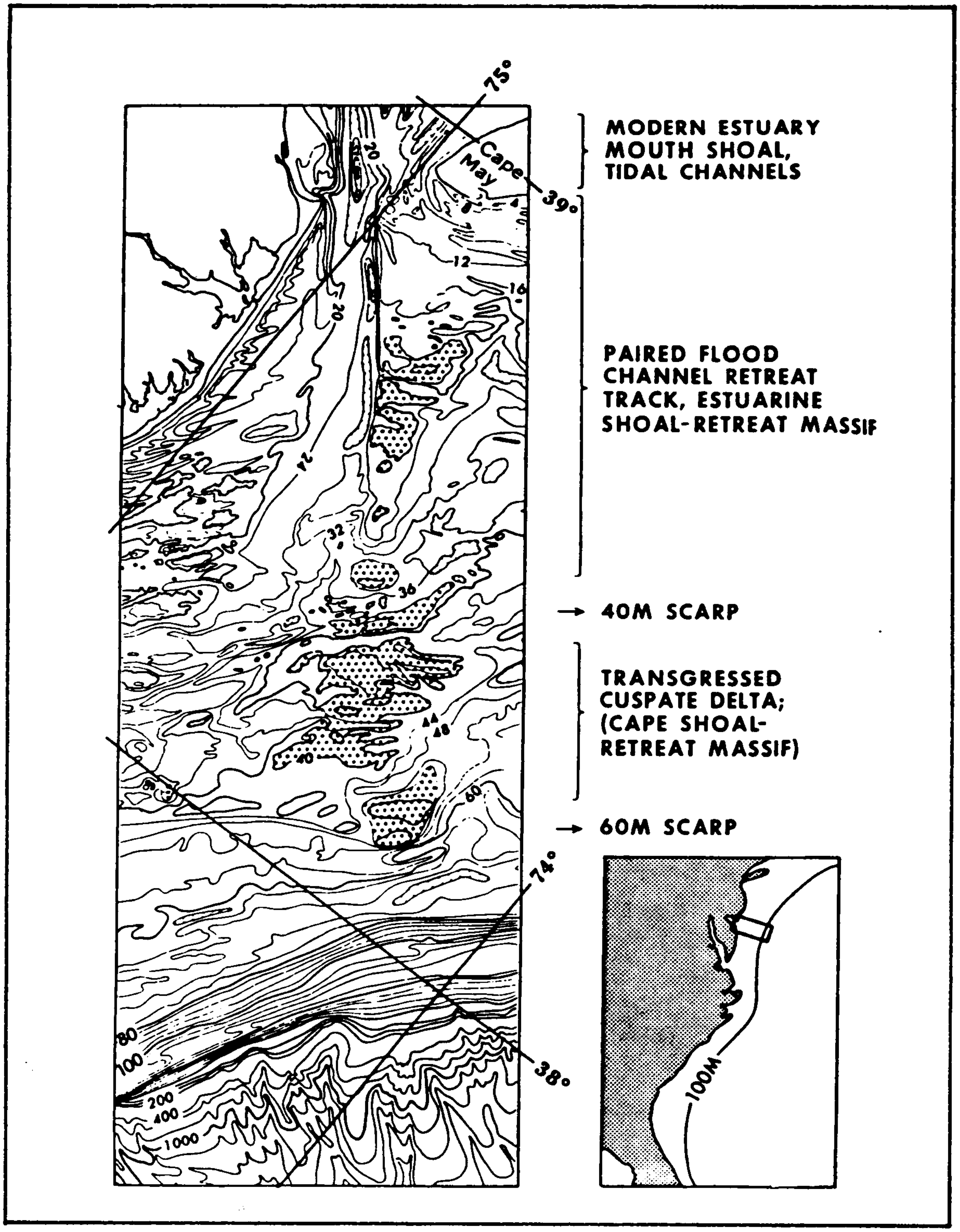


Fig. I-64

Delaware shelf valley, mid-shelf delta, and shoal retreat massifs. Stippled areas are shelf highs of submarine origin. After Swift 1973.

(Fig. I-64). This delta may have been active during a previous portion of the Pleistocene and reactivated by glacial metlwater sediment and littoral drift. From about 13,000 B.P. to about 10,000 B.P., the delta probably evolved into a cuspate foreland and barrier chain. During the earlier part of this stage, broad marshes may have developed as sea level eventually transformed the area into a large shallow lagoon or lagoon complex. By 10,000 or 9000 B.P. the delta-lagoon stage in the development of the Delaware River mouth had come to an end and had been replaced by the ancestral Delaware Bay as sea level rose about 40 m.

From the information given by Kraft (1971, 1977) and Belknap and Kraft (1977), estuary conditions extended as far inland as the Appoquinimink River by 9000 B.P. (Fig. I-65; Kraft 1977). Since then, Delaware Bay has moved further inland with further sea-level rise, and tremendous infilling has taken place, thus burying former river and stream valleys.

The path of an older (before 35,000 B.P.) Delaware River has been traced across the CS by Twichell and others (1977). Although this buried river valley, shown on Fig. I-66, predates the period of interest to archaeologists, a description of this particular buried valley is instructive. It was relatively broad and flat along the Inner and Middle Shelves, was 4-8 km wide, had relief of 10 to 15 m, and a gradient of less than 0.03° (Twichell and others 1977). On the Outer Shelf, the buried valley was much narrower (3-4 km wide), deeper (30 m), V-shaped in profile, and consequently had a steeper gradient. These circumstances are those normally expected from a Pleistocene lowstand and the associated entrenchment of a major river along the Outer Shelf. Once again, it is important to note that this buried river valley predates 35,000 B.P. The valley has been identified through seismic profiles, and it is not visible from existing shelf topography. Fig. I-67 shows 11 cross sections of this buried valley. The migration of shoals accompanying at least two sequences of marine regression and transgression, shoreline truncation, and submarine hydraulic processes have completely buried this Pleistocene valley. The preservation of this feature in the shelf geological record is evidence that marine transgression/regression does not destroy all former subaerial features.

Other investigations concerning the Middle and Outer Continental Shelves of the Delmarva compartment have provided little additional information on their pre-transgressive topography. The basal sections of ancient shorelines (scarps) have been identified along this compartment in the work done by Cousins and others (1977) and Dillon and Oldale (1978). The scarps are shown on Chart I-1b.

Cousins and others (1977) also provide some general information concerning major seismic reflectors observed along this portion of the CS. From seismic profiles, they have identified what appears to be the Holocene-Pleistocene unconformity (designated reflector II). This reflector is nearly horizontal and is observed in all of their profiles except at locations where it outcrops or is truncated by the sea floor. Beneath this reflector, probable stream channels and fluvial deposits

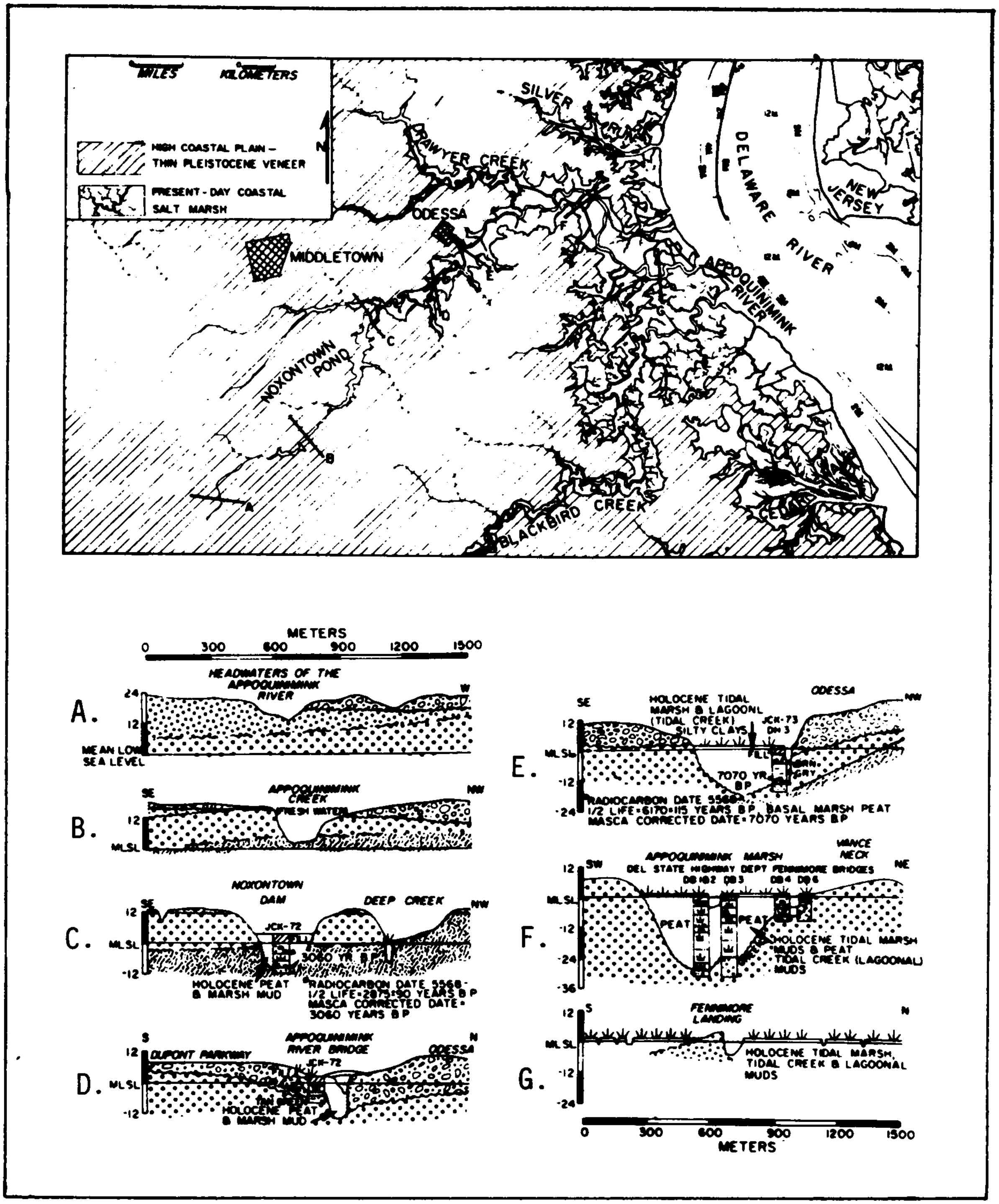
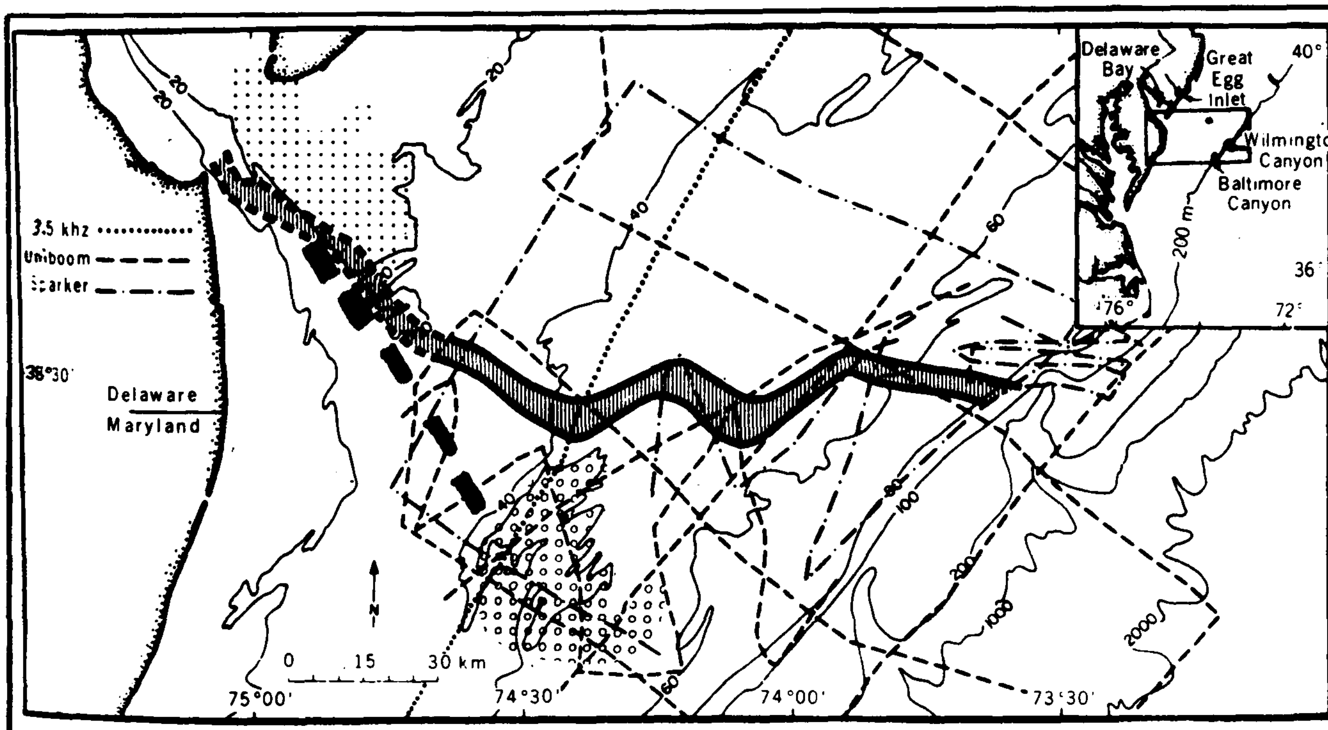


Fig. I-65

Present geomorphology of a section of inner Delaware Bay, Delaware. The locations of cross sections (shown below) along the Appoquinimink River are given on the above figure. Considerable sediment in-filling of the valley has taken place during the Holocene as indicated by carbon-14 dates. After Kraft (1977).



Shelf valley representing the retreat path of the mouth of the ancestral Delaware Bay. In the vicinity of this shelf valley lies the buried ancestral Delaware River Valley (Swift 1973; Swift and others 1972; Swift and Sears 1974).

Pre-late Wisconsin buried Delaware River Valley (Twichell and others 1977).

Delta identified on basis of bathymetry and morphology (Swift and others 1972; Swift 1973; Swift and Sears 1974).

Shoal retreat massif or shelf high representing the convergence region for littoral drift forming a local depositional center for sediment (Swift and others 1972; Swift 1973; Swift and Sears 1974).

Fig. I-66

Two Pleistocene paths of the Delaware River Valley. The stripe path represents a buried river valley of pre-late Wisconsin age (Twichell and others 1977). The shorter shelf valley leading to the mid shelf delta is approximately the late Pleistocene-Holocene path of the ancestral Delaware River (Swift and others 1972; Swift 1973; Swift and Sears 1974).

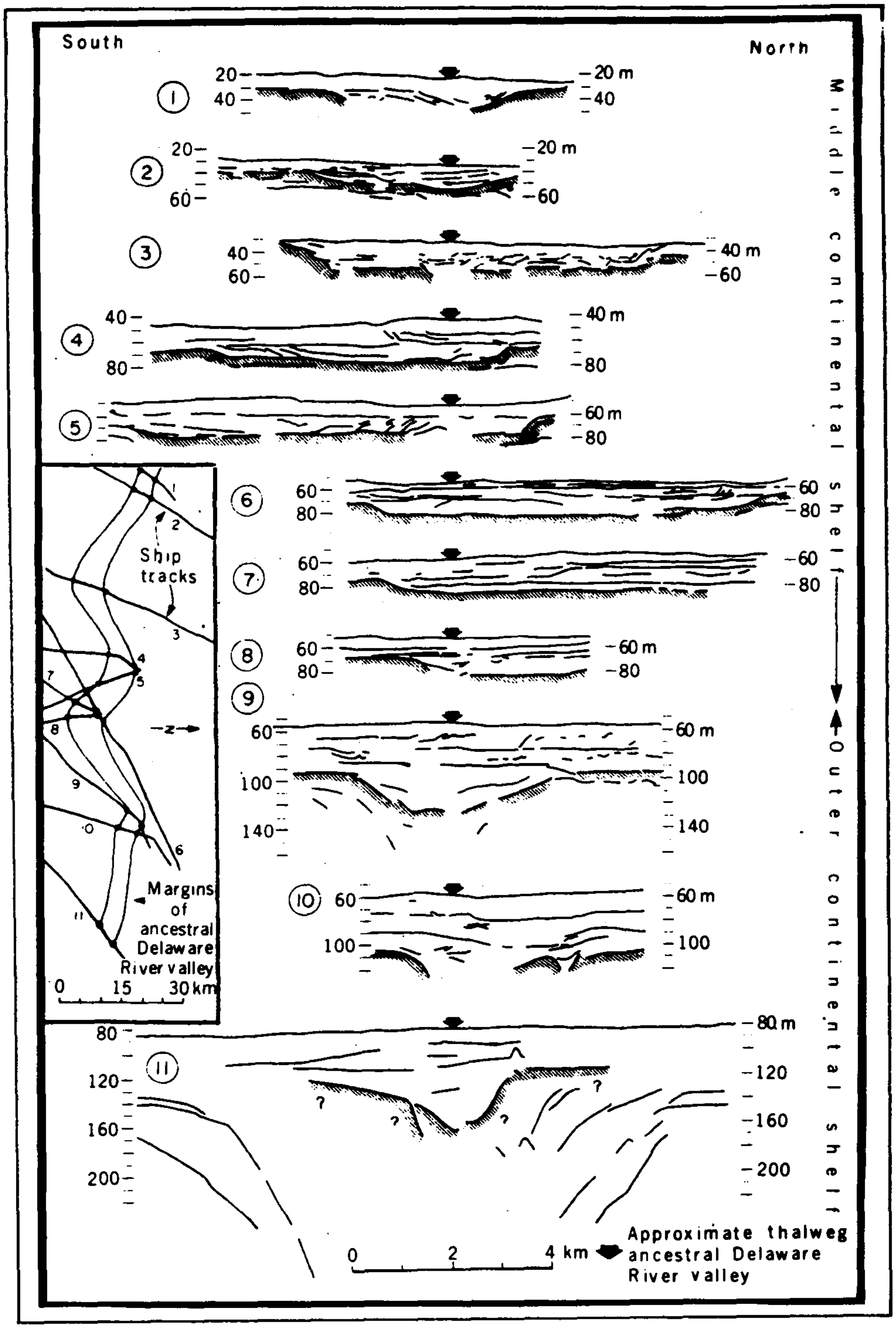


Fig. I-67

Line drawings of seismic profiles across a pre-late Wisconsin buried valley of the Delaware River. The buried valley is broad and shallow along the inner and middle shelf but becomes narrow and entrenched on the outer shelf. After Twichell and others (1977).

are visible in the upper section of the underlying sequence. Cousins and others (1977) rely on the dates offered by Shideler and others (1972) for assigning an age to the upper portion of this unit. As mentioned previously, it seems more reasonable to assign a pre-transgressive age to the upper unit found beneath reflector II. This would extend its possible age well into the Early Holocene, depending on its shelf location. The unit was originally formed during the last regression and consequently its basal portions are Middle Pleistocene in age (Knebel and Spiker 1977; Shideler and others 1972). The entire unit averages 8 m thick on the Inner Shelf and 15-18 m on the Outer Shelf (Cousins and others 1977). In some localities it may exceed 24 m.

Above reflector II, Cousins and others (1977) identify probably transgressive sediments (lagoonal muds, estuarine sediments, etc.), underlain locally by lag gravel. In some areas, the gravel may represent locations where major headlands have been truncated during transgression. In other localities, erosion has removed transgressive deposits leaving older sequences exposed. The upper portion of most transgressive deposits has been modified further, producing another reflector (I). This reflector represents the lower limit of the Holocene "surficial sand sheet." The reflector is only observable in a few locations since it is obscured in most areas by the acoustic return from the ocean floor. The "sand sheet" is a relatively thin, discontinuous layer of active (mobile) sand (Knebel and Spiker 1977; Stubblefield and others 1975; Swift and others 1972).

Several investigations into Late Pleistocene-Holocene sea-level positions have been carried out along the Delmarva Inner Shelf and coastal region (Belknap and Kraft 1977; Kraft 1971, 1977; Meyerson 1972; Newman and Rusnak 1965). The investigations done by Belknap and Kraft (1977) provide one of the most complete sea-level curves (extending back to at least 8000 B.P.) of any found along a single shelf compartment. Fig. I-68 presents the data Kraft (1977) compiled from the Delaware Bay region and coastal zone of Delaware. Table I-4 is a compilation of dates taken from the literature which have been considered to represent additional information on past sea-level positions. As both Fig. I-68 and Table I-4 illustrate, determinations of sea-level positions older than 9000 B.P. are questionable at this time.

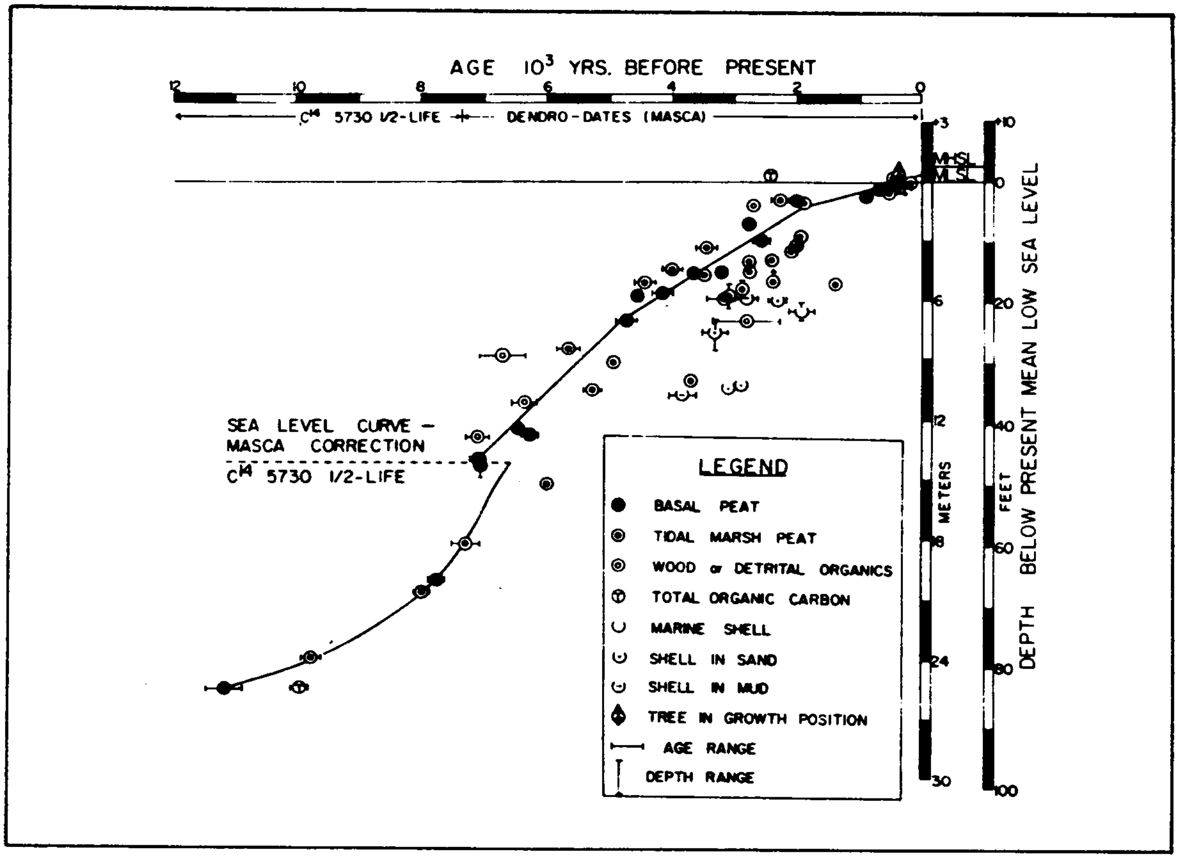


Fig. I-68
A relative sea level curve for coastal Delaware corrected according to MASCA dendrochronology dates. Early Holocene dates are based on 5730 one-half life. Data based on work of Kraft 1971, Belknap and Kraft 1977, and Kraft 1977. After Kraft 1977.

Table I-4: Approximate sea-level positions at 3,000-year intervals for the Delmarva Shelf Compartment.

	Range	Best Estimate	Source(s)
3000 B.P.	1-4 m	3 m	1, 2, 4
6000 B.P.	7-12 m	12 m	1, 2
9000 B.P.	21-27 m	22 m?	1, 3
12,000 B.P.	25-70 m	40 m?	3
15,000 B.P.	60-100 m	70 m?	3
18,000 B.P.	?	100 m?	3

Sources

1. Belknap and Kraft (1977)
2. Newman and Rusnak (1965)
3. Dillon and Oldale (1978)
4. Meyerson (1972)

8.0 NEW JERSEY SHELF

The New Jersey shelf compartment consists of a set of morphologic elements similar to those found along Delmarava. In general, the coast consists of a chain of barrier islands backed by extensive lagoons and marshes, except for its northern end, where headland erosion is in progress. In the middle of the New Jersey compartment, the ancestral Great Egg Valley has been identified (McClennen 1973; Swift and others 1972) as a major river system that formerly drained this region.

Along the New Jersey compartment, deposits from former lagoons on the Inner and Middle Shelves have been encountered (Stahl and others 1974; Stubblefield and others 1975; Stubblefield and Swift 1976). Scarps representing the truncated remains of former stillstands, have also been mapped along this section of the Atlantic coast(Chart I-1a and b), by Cousins and others (1977), Emery and Uchupi (1972), Ewing and others (1963), and Knott and Hoskins (1968). A number of researchers (such as Dillon and Oldale 1978; Edwards and Merrill 1977; Emery and Uchupi 1968) have noted several northward-dipping features (shelf break, scarps and terraces) along this section. Collapse of a glacial forebulge, differential water loading, and local tectonic activity have been suggested as causes. Dillon and Oldale (1978) noted specifically that scarp-elevation changes correspond between scarp systems and may be used to delineate inflection zones. The zones of inflection define a portion of the shelf which may have moved independently during the Late Pleistocene. The "Atlantis Shore" (also referred to as the Mid-Shelf shoreline: Swift and others 1972) may be used to establish a relative date for movement of this shelf block indicating that differential subsidence occurred sometime before 9000 or 10,000 B.P.

The boundaries for this shelf compartment are defined by the Delaware shelf valley (southern border) and the Hudson shelf valley (northern border). Bathymetry and some additional information are given in Fig. I-69. Major topographic elements along this shelf compartment are also shown on this figure.

The present-day coastal physiography for the New Jersey compartment consists of extensive tidal marshes lining the large lagoons found behind the barrier island.

A significant portion of the research done along the New Jersey shelf has focused on the "surficial sand sheet" and its ridge and swale topography (see for example Knebel and Spiker 1977; Knebel and Twichell 1977; McKinney and others 1974; Stubblefield and others 1975; Stubblefield and Swift 1976; USGS 1978). Some of the information derived from these investigations is of direct use to archaeological studies. Investigations along the New Jersey shelf have also shown that the sand sheet is not "relict" but is actively responding to storm-generated currents.

The Holocene path of the ancestral Delaware River was discussed in the preceding section on Delmarva.

Along the northern side of Delaware Bay, Meyerson (1972) has investigated a tidal marsh covering over a dozen square miles (Fig. I-70). Meyerson's study also includes pollen and paleosalinity investigations which help to support his paleoecological interpretations. The study indicates that fresh-water and tidal-marsh environments have fluctuated over the area in response to minor sea-level change and sedimentation. Meyerson considers the data to indicate that minor but significant changes in sea level have occurred over the last 3,000 years. It is his belief that these changes are hidden by the averaging process normally used to construct smooth sea-level curves. Meyerson (1972), however, does not uncover any evidence concerning Late Holocene sea level that is radically different from that given by the Belknap and Kraft (1977) curve. Some of the radiocarbon dates taken by Meyerson were not from basal peats (Fig. I-71) and therefore have little direct relationship to past sea levels. His data provide an example of nearshore "marine regression" resulting from slowing sea-level rise and increasing sedimentation during the last 3,000 years. The allochthonous peat encountered at the bottom of core #6 illustrates the need for using multiple peat cores as well as for making detailed analysis of the type of peat found in each core. In this particular case, these methods indicated that some peat (radiocarbon-dated at 2150 B.P.) was allochthonous and probably had been left behind by the meandering of a local tidal creed (Meyerson 1972).

Along the northern flank of the ancestral Delaware shelf valley, several areas have outcrops of Pleistocene or Holocene fine sediments (USGS 1978). These fine-grained deposits (clayey silt) generally occur in depressions and are strongly reflective, suggesting that they differ greatly from the adjacent surficial sands. Scouring between sand ridges has uncovered the finer-textured sediments. In water less than 40 m deep along the ancestral Delaware River, areas of fine-textured sediment made up 30% of the bottom topography. Away from the ancestral Delaware these sediments were less common, occupying about 16% of the Inner Shelf surface. Along the Middle and Outer Shelves, the incidence of outcrops of fine material dropped considerably, varying from 0 to 12% of the sea floor (USGS 1978). Although the origin of the fine-textured sediments is unknown, they probably represent estuarine and lagoonal deposits subsequently exposed by storm-dominated shelf currents.

Along the Inner Shelf portion of the ancestral Delaware Valley, Swift (1975a) has identified a "shoal retreat massif" from existing bathymetry. The northern side of this massif would have experienced considerable erosion well into the underlying pre-transgressive deposits, as outlined by our model. The southern edge of this massif has probably buried some estuarine deposits, protecting them from additional erosion by a substantial layer of sand.

Beachrock has been recovered from several areas along the Middle and Outer Shelves (Allen and others 1969; Folger 1977). In the area

LEGEND

- Buried valley on the shelf identified by seismic profiling and containing flood plain, river channel, river terrace and estuarine deposits
- Surface valley on shelf indicating that buried estuarine and possibly river valley deposits probably exist within the general vicinity
- Inferred buried valley approximately located and containing flood plain, river channel, river terrace and estuarine deposits
- Buried valley probably older than 35,000 B.P.
- Major shelf delta identified by bathymetry, morphology, and sometimes seismic data when available
- Shoal retreat massif (shelf highs of constructional origin due to littoral drift deposition centers)
- Buried river valley deposits with location questionable

Fig. I-69
Major Late Pleistocene-Holocene features on the Long Island continental shelf.

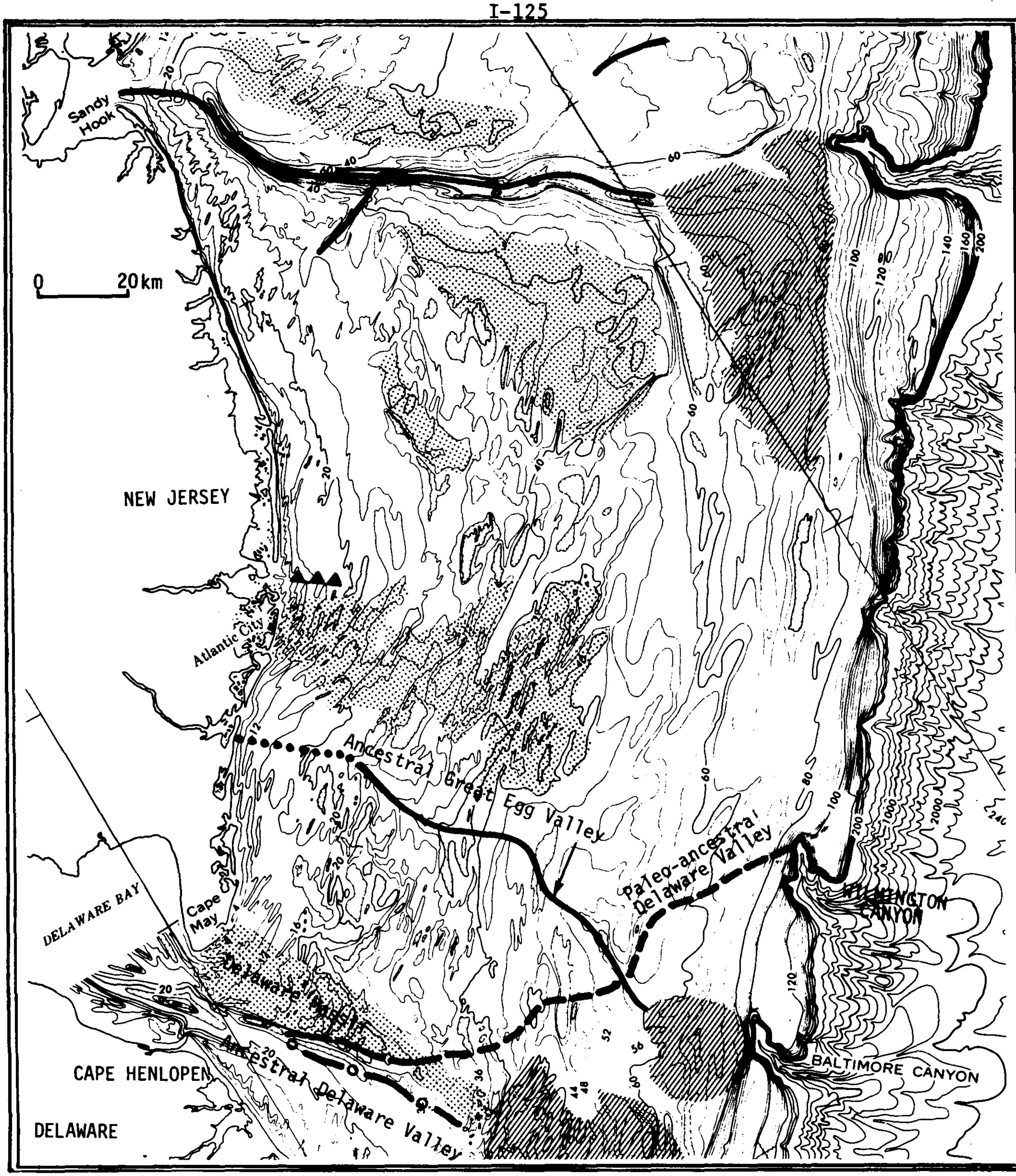


Fig. I-69 Major Late Pleistocene-Holocene features on the Long Island continental shelf. Features compiled from the following sources: Knott and Hoskins (1968); McClennen and McMaster (1971); Swift (1973); Swift and others (1972); Swift and Sears (1974); Twichell and others (1977). Contours in meters.

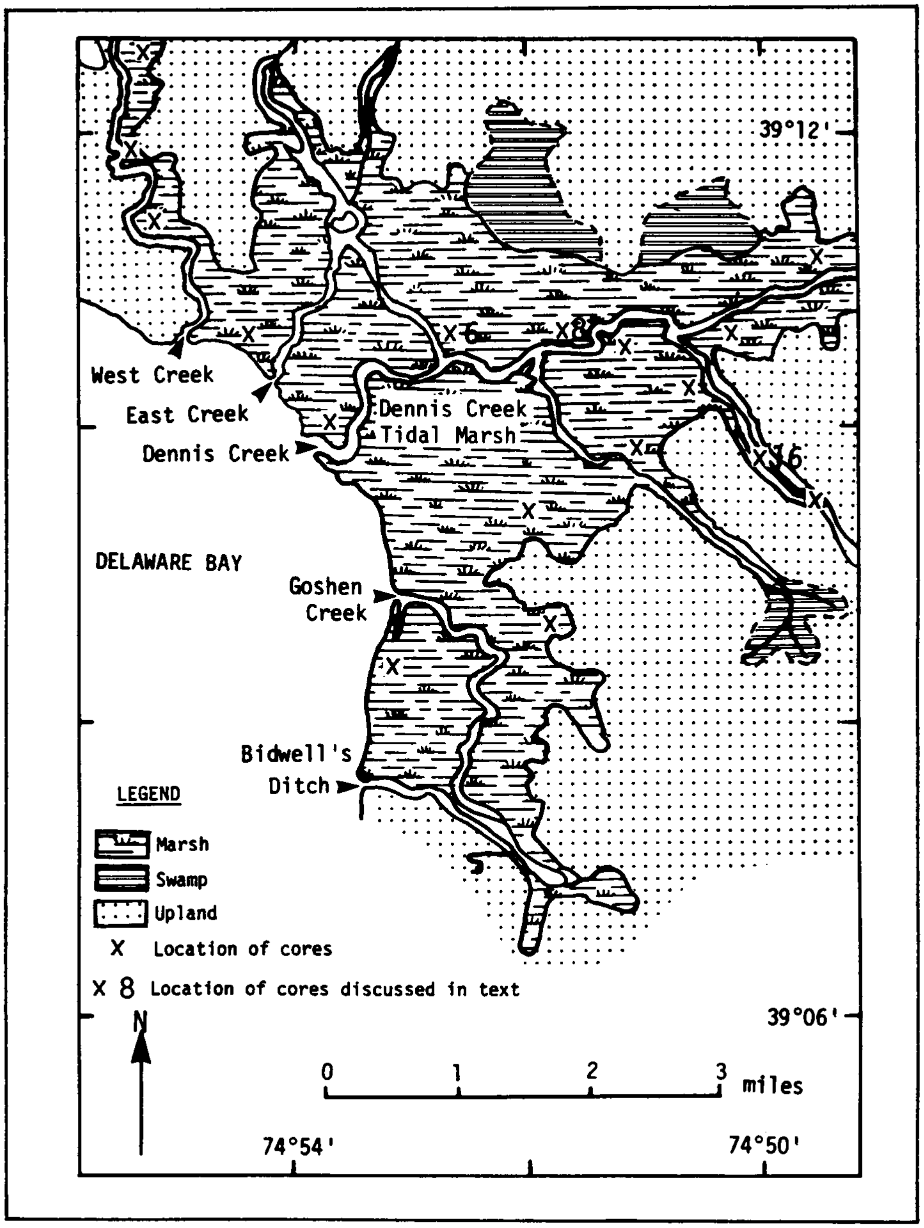


Fig. I-70

Dennis Creek tidal marsh and location of cores studied by Meyerson (1972). Cores 6, 8, and 16 are illustrated in Fig. I-71. Adapted from Meyerson (1972).

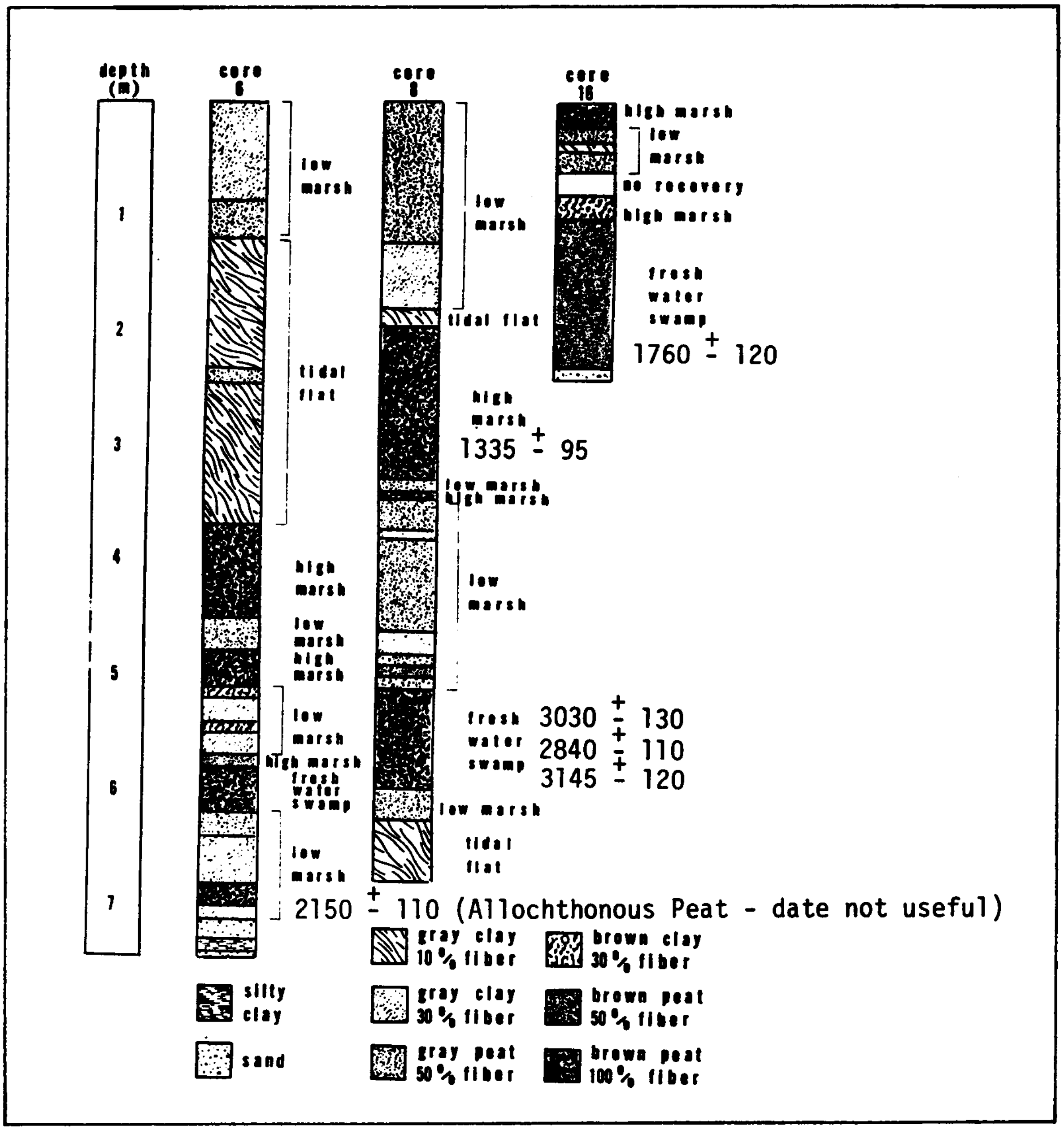


Fig. I-71

Lithology and inferred depositional environment for 3 cores from Dennis Creek Tidal Marsh, New Jersey. After Meyerson (1972). Dates given in years B.P.

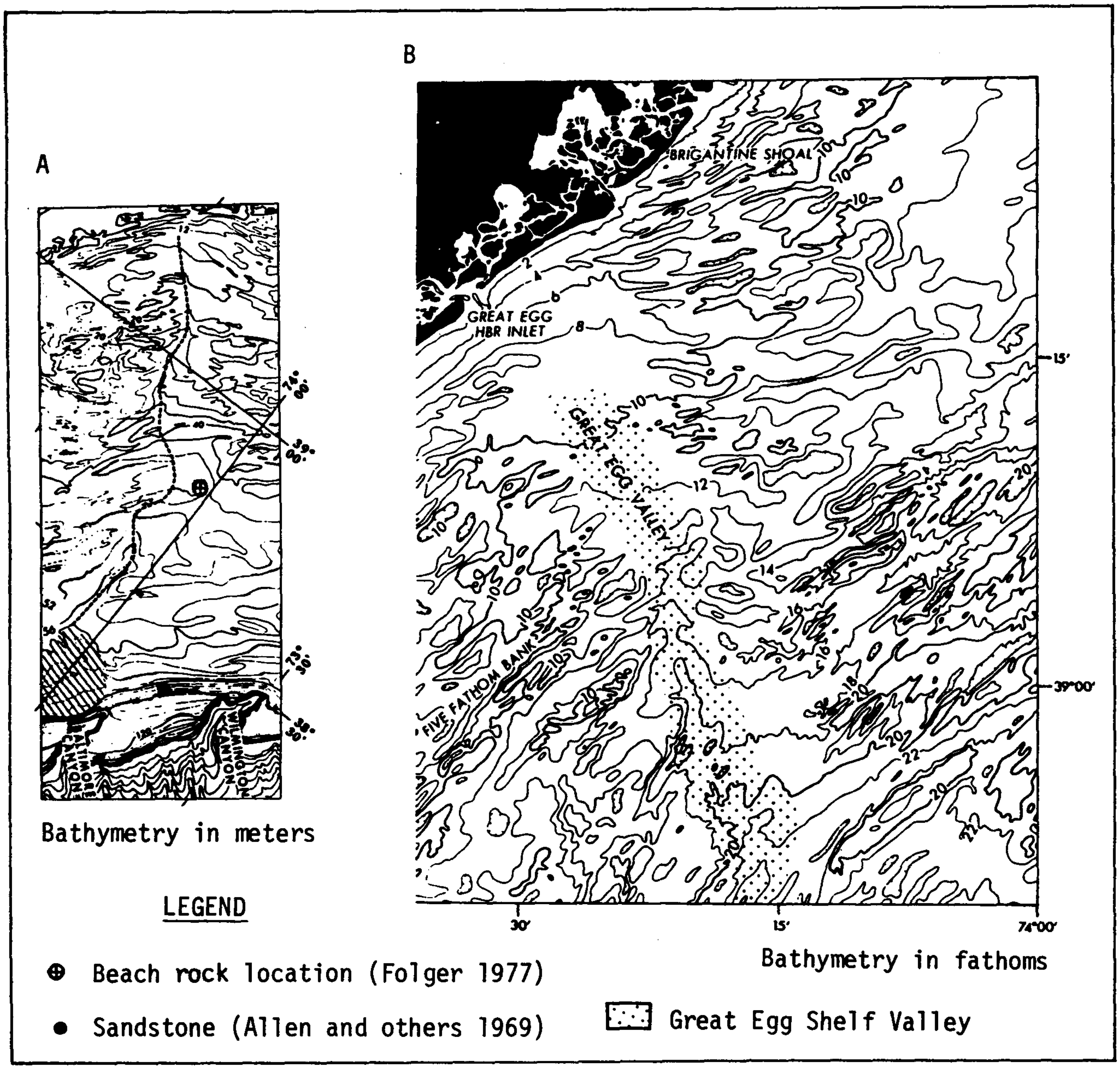


Fig. I-72
(A) Thalweg and probable delta associated with the Great Egg Shelf Valley. After Swift and others (1972). (B) Inner shelf section of the Great Egg Shelf Valley showing corridor in which buried estuary and possibly flood plain deposits exist. After Swift and others (1972).

discussed by Folger (1977) beach rock with shells dating to 21,600+ B.P. was recovered (Fig. I-72). The underlying silty horizon was dated at 29,500 B.P. Similar deposits on the shelf have been found to date from 24,000 B.P. to more than 40,000 B.P. (Knebel and Spiker 1977; Stubblefield and others 1975).

Reliance on "beach rock" for determining depth-age relationships is questionable unless detailed analyses are performed. Allen and others (1969) recovered "Beach rock" at a depth of 79 m (Fig. I-72) which produced conflicting dates. The shells dated from 4390 B.P. and the cement from "15,600" B.P., giving a "whole rock" date of 10,500 B.P. Allen and others (1969) suggested that the "beach rock" had been lithified at subtidal depths and that the process involved methane from buried marsh deposits. Their interpretations are backed up with carbon-isotope percentages, as well as textural and faunal analyses. Their research may offer very indirect evidence for possible buried tidal marsh deposits at least as old as 15,600 B.P. along the outer portion of the CS.

Along the central portion of the New Jersey compartment, investigations have revealed much about the area's Late Quaternary evolution. In this area, the ancestral Great Egg Valley has been traced across the Shelf (Fig. I-72; McClennen 1973: McClennen and McMaster 1971; Stout and McClennen 1977; Swift and others 1972:. Within this valley there should be buried river-channel and flood-plain deposits beneath a cover of estuarine and marine sediments. Twichell and others (1977) encountered evidence of a south-trending channel crossing the path of the pre-35,000 B.P. Delaware River near the Outer Shelf. Swift and others (1972) identify a probable Outer Shelf delta for the ancestral Great Egg River (Fig. I-72). The Holocene Great Egg River was much smaller than either the Delaware or Hudson River systems. Most of its waters came from the exposed shelf and a small portion of the New Jersey coastal plain southeast of the Delaware River. Before the Late Pleistocene, the Great Egg River system included the ancestral Schuylkill River, but the latter was lost when it was captured by the Delaware River system. It is quite obvious that its catchment basin is much smaller than either the Delaware or Hudson drainage system.

The location of the ancestral Great Egg Valley has been determined by high-resolution seismic reflection techniques (McClennen 1973; Stout and McClennen 1977). Fig. I-72 shows the path of the ancestral Great Egg Valley. Within this valley there is a chance of encountering intact flood-plain deposits and similar subaerial surfaces beneath estuarine deposits.

Several areas have been investigated in detail along the central portion of the New Jersey shelf compartment. These studies have provided some information on pre-transgressive deposits, but most research concentrated on the "surficial sand sheet" and its ridge-and-swale topography. Fig. I-73 shows the area reviewed by Stubblefield and others (1975) and Stubblefield and Swift (1976) who used seismic reflection and cores. In this area they encountered at least three stratigraphic units. The lowermost unit was a fine-grained, shell-free sand which had been deposited in a nearshore environment probably during a major transgression. Stratigraphic constraints placed the age of this unit at greater than 36,000 B.P.

The next unit above was a layer of medium gray, silty clay forming a widespread reflector over this area. The unit's age ranged from 25,000 to 36,000 B.P. The younger section of this unit probably represents an offshore deposit formed adjacent to a prograding shoreline during a period of sea-level fall. That is, it represents material deposited in a nearshore environment by rivers at the time of a marine regression. During this period, a lobate coastline with broad peninsulas separating large interfluvial embayments probably existed. The lowest sediment of this unit probably represents marine deposits formed during the Plum Point interstadial (about 36,000 B.P.).

Overlying the nearshore sediments is a deposit of medium-grained upward-coarsening sand varying in thickness from 1 to 8 m. An age of 22,305 B.P. suggests that this material was deposited in a prograding shoreline which overrode the nearshore deposits mentioned above. Above this unit, Stubblefield and Swift (1976) identified lagoonal deposits ragning from 36 to 52 m below sea level and varying up to 6 m in thickness. The deposits consisted of medium gray silty clay and were discontinuous as a result of erosion and scour. This unit is considered to fall between 22,000 and 11,000 B.P. in age. Of the 115 km of seismic profiling investigated by Stubblefield and Swift (1976), only 9 km indicated that erosion had completely removed the lagoonal unit, exposing the underlying sand substrate. If their sampling is representative of the general area, then there appears to be strong evidence that considerable amounts of lagoonal deposits are preserved beneath the "surficial sand sheet." Crassostrea virginica and Mercenaria mercenaria shells were recovered in the lagoonal muds (see Table I-5) supporting the environmental interpretation suggested for this unit. As explained in our model, the base of this unit has the potential for containing an intact subaerial surface. Fig. I-74 illustrates some of the sequences encountered by Stubblefield and Swift (1976).

Knebel and Spiker (1977) investigated two areas (Fig. I-75) along the New Jersey Shelf compartment. Although they were primarily interested in the age and thickness of the "surficial sand sheet," their investigations also shed additional light on the extent of the underlying substrate. Vibracores and seismic reflection indicated that the surficial sands in these two areas rest upon a much older, texturally diverse unite. Table I-6 gives the data they collected from vibracores. In general, silty and clayey sands, sandy and silty clays, sand-silt-clay, and clays are encountered beneath the "surficial sand sheet." Shell fragments are scarce throughout this unit and, when present, usually are associated with sandy zones. Radiocarbon dating of material from below the "surficial sand sheet" was done on only three of the cores. Unfortunately, two of these were located in depressions (core #111-030 in subarea 3, and 095-002 in subarea 1). Scouring has probably removed

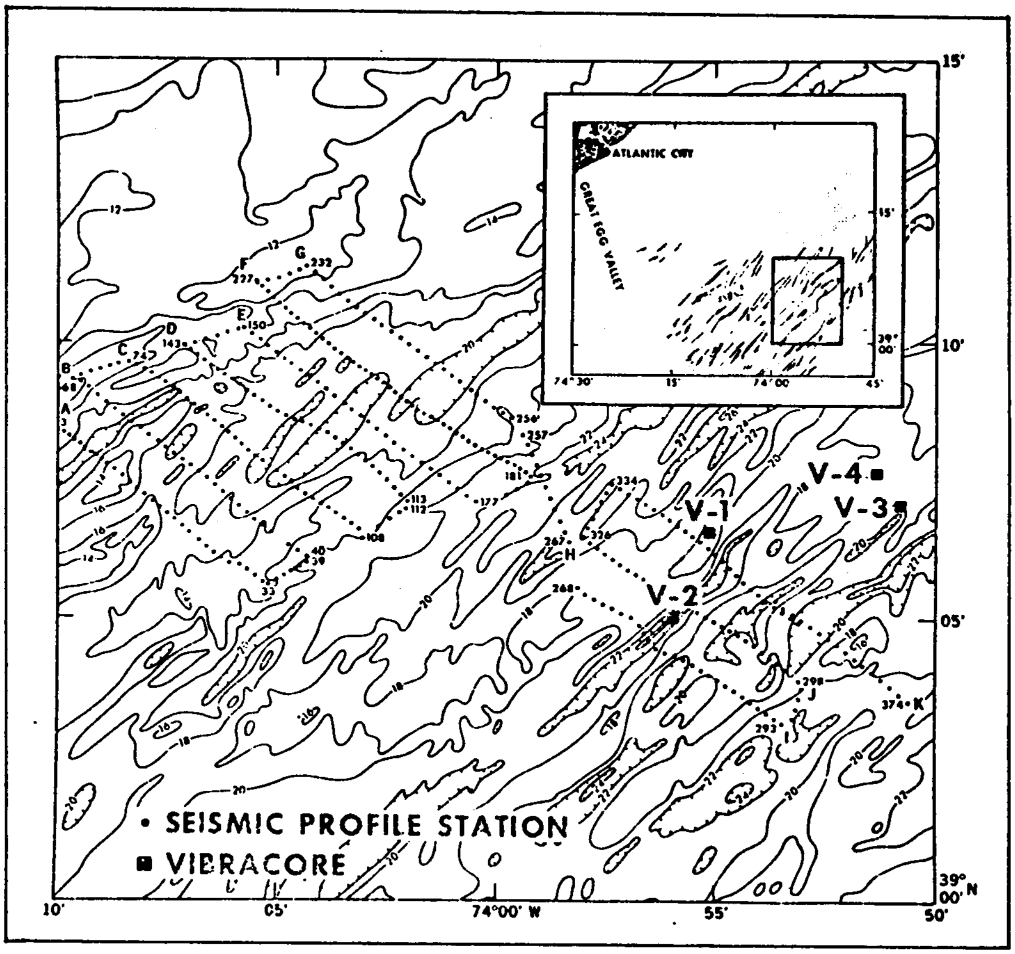


Fig. I-73

Index map of area studied by Stubblefield and Swift (1976) showing seismic profile stations and vibracore locations. Sediment sequences encountered in vibracores and general interpretations are illustrated in Fig. I-74. After Stubblefield and Swift (1976).

Table I-5: Radiocarbon dates of selected fauna from the middle continental shelf off New Jersey (after Stubblefield and others 1975). Location of cores shown on Fig. I-73.

Core	Position	Fauna dated	Depth below sea level	Depth from top of core	Apparent age (years B.P.)
V-2	39°05.1'N 73°55.6'W	*Mercenaria mercenaria*	45.63 m	(83) cm	29,700 ± 650
V-2	Same	*Crassostrea virginica*	46.95 m	(125) cm	32,150 ± 600
V-3	39°05.7'N 73°50.5'W	*Mercenaria mercenaria*	41.88 m	(8) cm	10,950 ± 360
V-3	Same	*Ensis directus*	43.30 m	(150) cm	22,035 ± 665
V-3	Same	*Mercenaria mercenaria*	44.30 m	(250) cm	25,300 ±1040 1200
V-3	Same	*Mercenaria mercenaria*	45.50 m	(370) cm	36,000
V-4	39°06.9'N 73°31.3'W	*Placopecten magellanicus*	36.90 m	(60) cm	500
V-4	Same	*Placopecten magellanicus*	37.55 m	(125) cm	3,760 ± 70
Trough Sample # 156	39°05.6'N 73°55.1'W	*Crassostrea virginica* *Mercenaria mercenaria*	44.00 m		10,050 ± 170

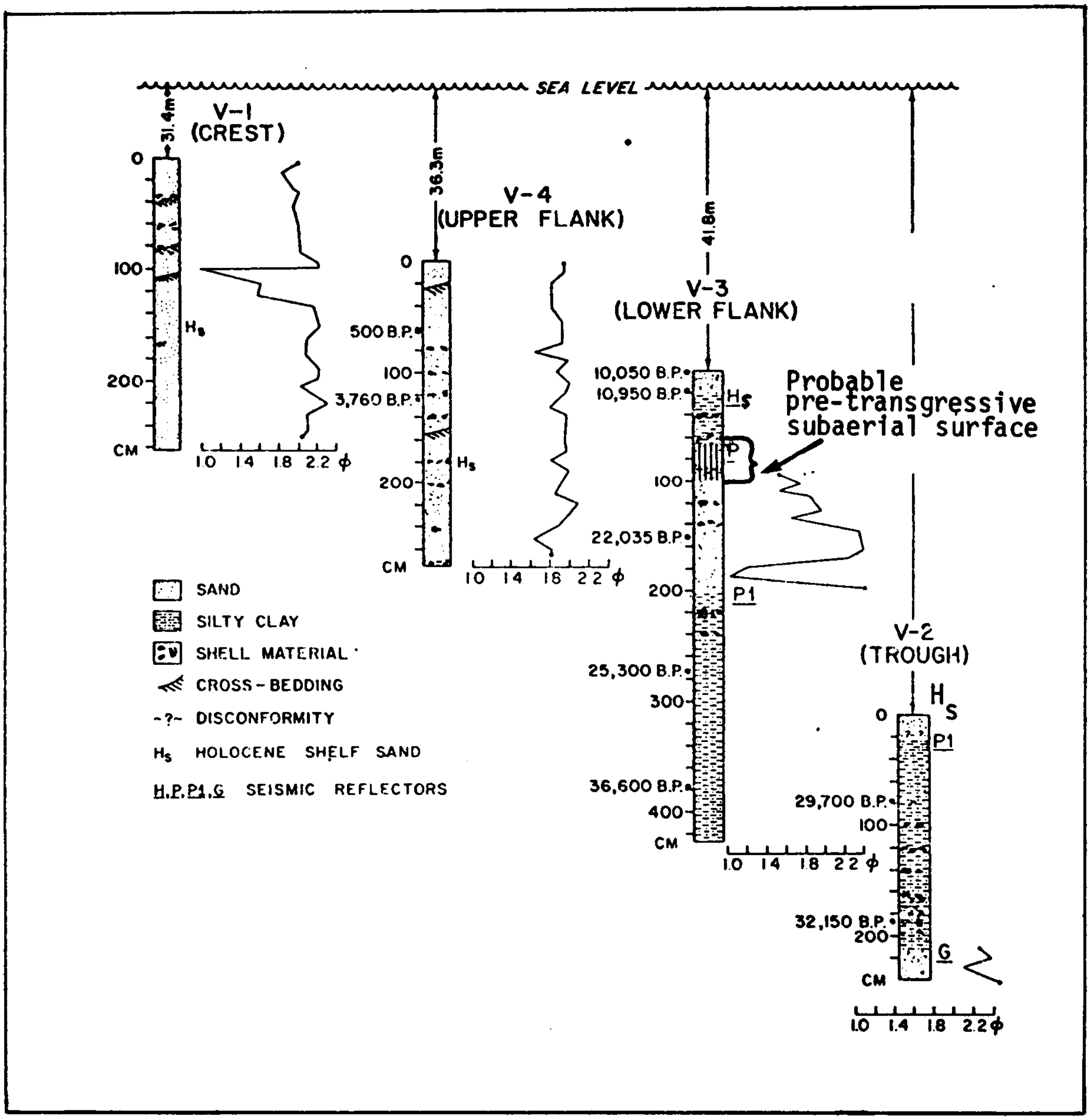
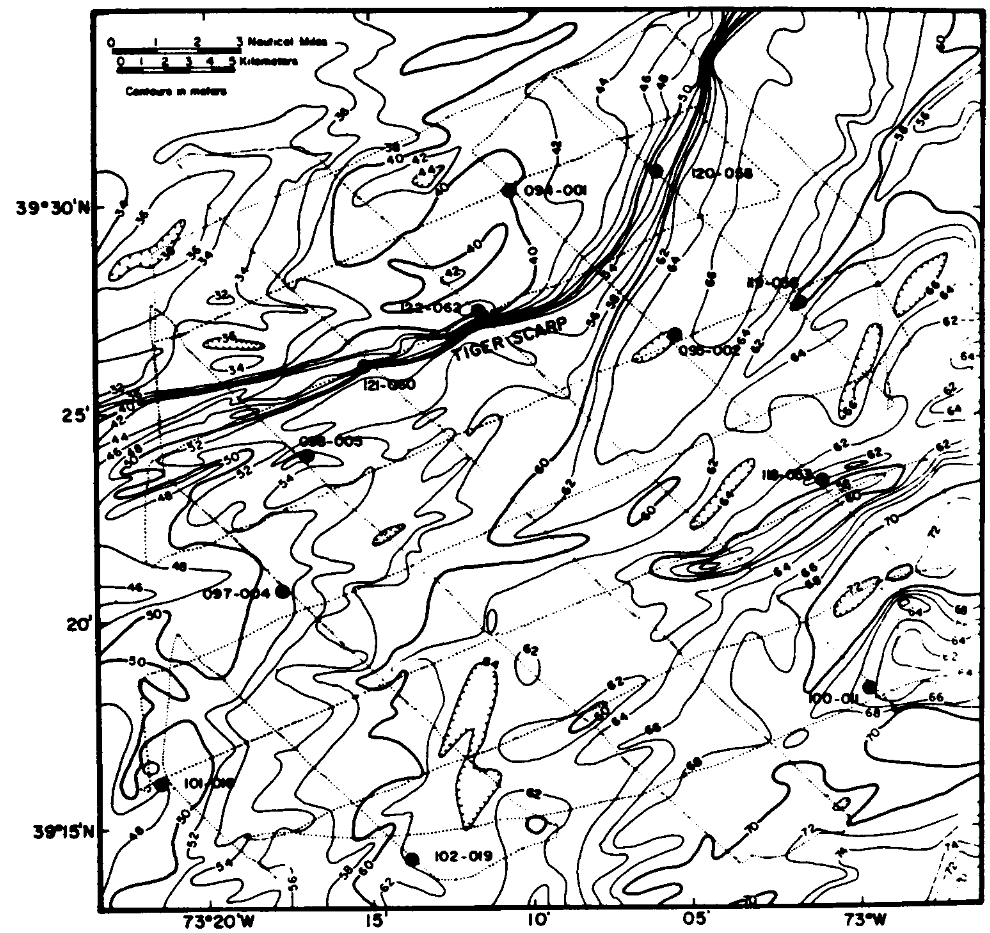


Fig. I-74

Sediment sequences from four vibracores from the New Jersey shelf showing lithic types, structure, and mean grain sizes. After Stubblefield and Swift (1976). Locations of vibracores shown in Fig. I-73.

SUBAREA I



LEGEND

- Core locations

······Bathymetric and seismic tracklines

Fig. I-75a

Subareas studied by Knebel and Spiker along the New Jersey continental shelf. The location of cores given in Table I-6 are indicated for each subarea. After Knebel and Spiker (1977).

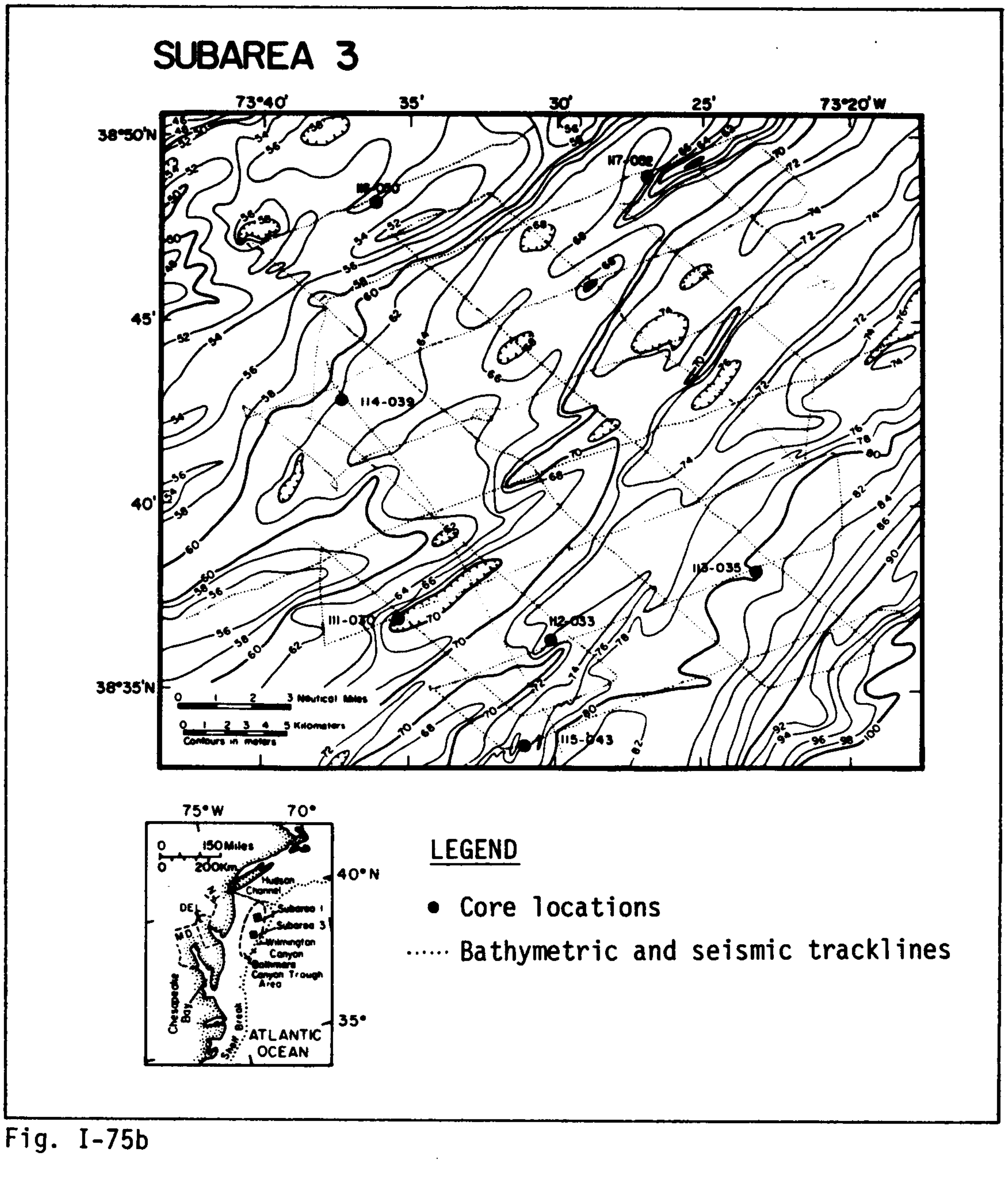

SUBAREA 3
73°40'
35'
30'
25'
73°20'W
38°50'N
45'
40'
38°35'N
114-039
113-035
111-030
112-033
115-043
117-002
0 1 2 3 Nautical Miles
0 1 2 3 4 5 Kilometers
Contours in meters
75°W
70°
150 Miles
200Km
40°N
35°
Hudson Channel
Subarea 1
Subarea 3
Wilmington Canyon
Baltimore Canyon Trough Area
DE
MD
Chesapeake Bay
Shelf Break
ATLANTIC OCEAN
LEGEND
• Core locations
······ Bathymetric and seismic tracklines

Fig. I-75b

Table I-6: Radiocarbon Analysis (after Stahl, Koczan, and Swift 1974).

Sample	Elevation (mlw) (m)	(ft)	Material	Unit	Date (yr B.P.)
816-35	-20	-66	Peat	Holocene H1	8.210 ± 120
821-28	-19.5	-64	Peat	Holocene H1	8,595 ± 215
825-21	-19	-62	Total organics	Late Pleistocene-Early Holocene	29,320 ± 1,300
829-12	-12.5	-41	Mollusk	Holocene H2	6,685 ± 170
833-25	-19	-62	Peat	Holocene H1	7,860 ± 190
837-15	-16.5	-54	Total organics	Late Pleistocene-Early Holocene	11,800 ± 520
839-20	-16.5	-54	Total organics	Holocene H1	6,835 ± 170
842-11	-15.5	-51	Total organics	Late Pleistocene-Early Holocene	25,100 ± 770
848-26	-19	-62	Peat	Holocene H1	7,790 ± 130
850-08	-13.5	-45	Total organics	Late Pleistocene-Early Holocene	16,900 ± 620
851-09	-13	-42	Total organics	Late Pleistocene-Early Holocene	22,870 ± 530
862-32	-21	-68	Peat	Holocene H1	7,880 ± 125

the pre-transgressive deposits in these areas (Stubblefield and others 1975; Stubblefield and Swift 1976). Consequently, the apparent absence of Early Holocene lagoonal deposits directly beneath the "surficial sand sheet," in core 111-030 (subarea 3), for example, may be the result of local scouring that occurred within this depression. Similar scouring may also have occurred in the vicinity of core 095-002. In subarea 1, the shell dated 26,650 B.P. came from a sandy unit about 1.8 m below the bottom of the "surficial sand sheet." The work done by Stubblefield and others (1975) and Stahl and others (1974) suggests that lagoonal deposits may be present just below the surficial sands where Knebel and Spiker (1977) note the presence of an interval of clayey sand. As our model has explicitly pointed out, lagoonal deposits preserved beneath the "surficial sand sheet" generally represent only the basal portions of sequences truncated by transgressive and post-transgressive hydraulic processes.

All of the remaining radiocarbon dates (16 of the 19 examples) in the study by Knebel and Spiker (1977) were from within the "surficial sand sheet." Except for one core, all of the dates derived from these material fall within the last 10,500 years. In most cases, the shells seem to represent the remains of post transgressive nearshore and mid-shelf populations. Knebel and Spiker's (1977) research, when reviewed in light of the other studies conducted along this portion of the Shelf, lacks sufficient evidence to discredit the possible existence of lagoonal deposits. More critical examination of the deposits beneath the "surficial sand sheet" is required in order to understand their origin, age, and spatial extent.

Along the Inner Shelf of central New Jersey, Stahl and others (1974) report on investigations of a proposed nuclear power plant location. Seismic profiling and drilling allowed them to investigate well over 20 m of shelf sediments. Fig. I-76 gives a profile across a portion of their study area. Lagoonal deposits were encountered throughout most of the area. Three major stratigraphic units were encountered, ranging from Miocene to Holocene in age. The Miocene unit was correlated with an outcrop on the nearby New Jersey coastal plain. Above this unit, dense sands and clays up to 8 m thick and containing shells dating between 22,870 and 29,320 B.P. were encountered. The basal portion of this unit represents nearshore deposits, while the uppermost sequences are probably composed of deltaic-fluvial sediments. The very top of this sequence may contain a soil horizon dating between 16,900 and 11,800 B.P. The "soil horizon" forms a weak but continuous acoustical horizon, suggesting that it may represent a well-developed Early Holocene subaerial surface. This is one of the few studies reviewed which suggests evidence for a buried soil horizon on the Shelf. Above these deposits was silty clay separated by a disconformity. The silty clay varied from 2 to 9 m in thickness and gave radiocarbon ages of from 8595 to 6685 B.P. Stahl and others (1974) and Swift (1976b) view this unit as the basal portions of a lagoonal sequence. The upper sequence, which was formed during the last marine transgression, has been mapped in detail. The trough-like portion of this unit (Fig I-76) has been interpreted as the filled-in remains of a coast-parallel tidal channel. The absence of basal peat

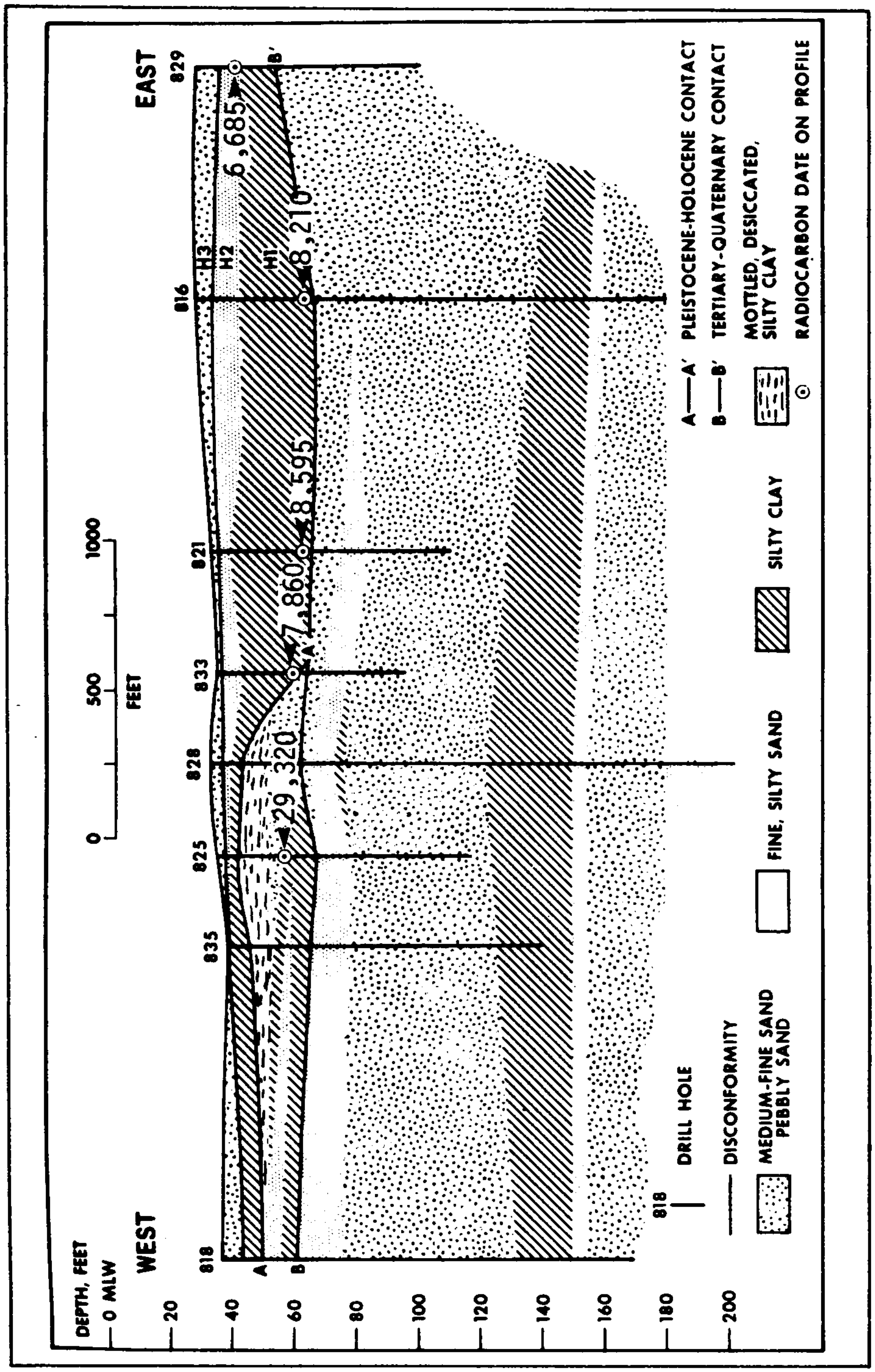


Fig. I-76

Cross section from the inner continental shelf along New Jersey. Surficial sand sheet (H3) covers back barrier (H2) and lagoon (H1) sequences. Some evidence of a pretransgressive subaerial surface is suggested by the zone of mottled desiccated silty clay. Thick zone of H1 between drill holes 828 and 829 represents a filled tidal inlet viewed across its narrow axis. From Stahl and others (1974). Table I-6 presents data collected from several of the drill holes shown in this figure.

or desiccated clay beneath the eastern section of the lagoon (Fig. I-76) is the result of tidal scouring through pre-transgressive deposits into the underlying tertiary substrate (Stahl and others 1974).

Overlying the lagoonal fine-grained sediments are light-gray silty sands which locally interfinger with the underlying muds. Radiocarbon dates suggest that they are roughly contemporaneous with the lagoonal sediments, and Stahl and ohters (1974) interpret these deposits as representing (truncated) back barrier and lagoon washover sands.

Finally, on top of this sequence is a very thin Holocene "transgressive sand sheet" formed during erosional retreat of the shoreface. It has been reworked by modern hydraulic processes. This sequence is one of the most complete records of a transgressed lagoon from the Middle Atlantic Bight.

At the northern end of this shelf compartment, some investigators have studied the Hudson Shelf Valley and the adjacent areas. The Hudson Shelf Valley is one of the few valleys which has not be completely buried by erosion or deposition since Holocene transgression. The thalweg of the present shelf valley is somewhat concave (Fig. 77). Although some deposition has occurred along the valley since its submergence, its concave-like profile is considered evidence of a former forebulge(s) (Edwards and Merrill 1977). The depth of the valley is viewed as a product of downcutting by Pleistocene glacial meltwater, possibly including some runoff from the Great Lakes Region (Veatch and Smith 1938).

Near the shelf break, several investigators have identified a delta (that is, the Hudson Delta) associated with the former Hudson River (Cousins and others 1977; Emery and Uchupi 1972; Ewing and others 1963; Knott and Hoskins 1968; Swift and others 1972; Veatch and Smith 1938). The delta is characterized by irregular reflectors probably representing cut-and-fill sequences (Cousins and others 1977). The delta, however, represents several Pleistocene depositional and erosional events.

Knott and Hoskins (1968) identified some filled channels cutting across the Middle and Outer New Jersey Shelves. These buried channels are much older and are considered to be pre-Wisconsin paths for the Hudson River. Kelling and others (1975) also discuss additional paths for the Pleistocene Hudson River which fall beyond the period of interest to archaeologists.

Along the southern flank of the Hudson Shelf Valley, Swift and others (1972) have outlined several shelf highland areas (Fig. I-69). These plateau-like uplands ("cuestas?") may be partly controlled by bedrock. Because these regions are above the surrounding shelf, they would have been exposed longer than the adjacent lowlands. Their elevation over the surrounding areas would have made them deposition centers for littoral transport during transgression. After transgression had passed this area, these shelf highs would have remained unprotected

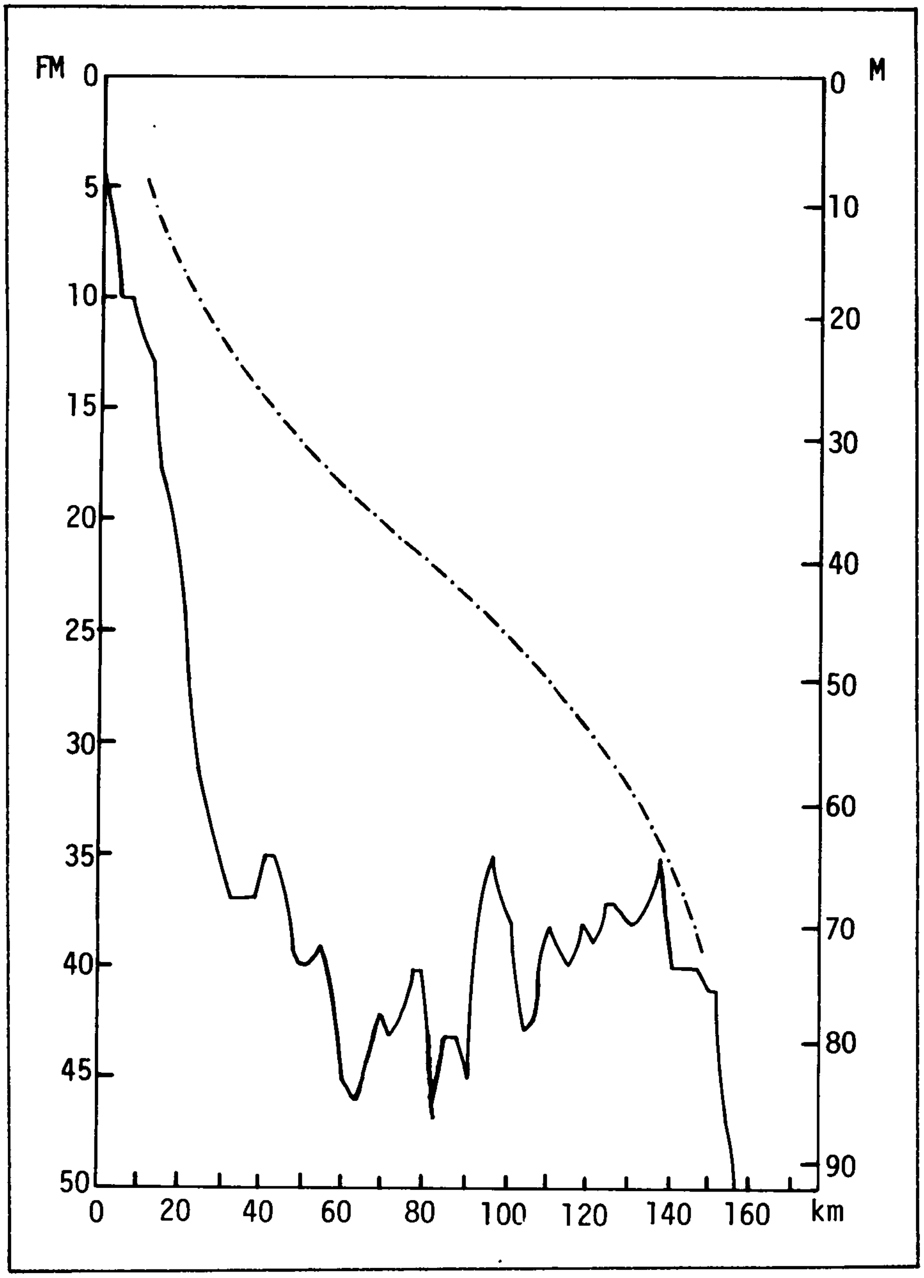


Fig. I-77
Profile of the Hudson shelf valley thalweg (solid line) and generalized profile of the slope of the adjacent shelf. Edwards and Merrill regard the general concavity of the shelf valley as evidence in support of a late Wisconsin glacial fore-bulge along the continental shelf. After Edwards and Merrill (1977).

from shoaling waves for a longer period than the adjacent shelf lows. Consequently, the unconsolidated deposits on these high areas probably have received considerable erosion and redeposition. This may partly explain the absence of lagoonal deposits in the cores collected by Knebel and Spiker (1977) from one edge of this "plateau."

Several smaller valleys may still be visible along the southern flank of the Hudson Shelf Valley. Buried river-valley deposits have not been identified in these valleys because no detailed seismic profiling or coring has been done.

The Tiger Scarp, located along the southern flank of the Hudson Shelf Valley, is probably a multicomponent feature. Schlee (1973) and Schlee and Pratt (1970) have suggested that the Scarp is part of a fluvial terrace belonging to the Pleistocene Hudson River. Knebel and Spiker (1977), using coring and seismic reflection, recovered evidence indicating that considerable erosion and redeposition has taken place during the last 8,000 years. Although the development of this feature is not adequately understood, Knebel and Spiker (1977) view its origin as partly due to deltaic or littoral sands' being added to an older Hudson River terrace. Foraminifera, possibly representing a lagoon or nearshore assemblage, were recovered 4.3 m below the surface at a depth of 52 m near the center of the scarp and suggest yet another facet of the development of this complex feature (Knebel and Spiker 1977). The descriptions of this feature also include the necessary components of a shoal retreat massif (Swift, personal communication).

Knebel and Spiker (1977) also investigated an area slightly north of Wilmington Canyon (subarea 3) using seismic profiling and coring (Fig. I-75). In this region, the "sand sheet" was found to range from 1 to 20 m thick. Thickness was largely related to the bathymetry, with the greatest thickness of sand occurring in the shallowest areas. Radiocarbon dates placed the "sand sheet" well within the Holocene (Table I-6).

Beneath the "surficial sand sheet" was a muddy deposit which was texturally diverse and which contained shells dating earlier than 24,000 B.P. Knebel and Spiker (1977) do not identify any lagoonal deposits in their cores from subarea 3 but the texture of some of the units encountered may possibly indicate such features. As mentioned previously, these two researchers were most interested in the age and thickness of the "surficial sand sheet" and not in the deposits beneath it. The two carbon dates of material from beneath the sand sheet do not indicate the presence of lagoonal deposits (compare Stubblefield and others 1975 and Stubblefield and Swift 1976).

Cousins and others (1977) have reviewed high-resolution seismic profiles from the Middle and Outer Shelves along the New Jersey shelf. They identified a reflector (reflector II) as the remains of the subaerial surface cut during the Late Wisconsin period. It is best observed on the Outer Shelf within 10 to 15 km of the shelf break. In some profiles it outcrops between 140 and 180 m and is also found to underlie the

Fortune shore. The work done by Knebel and Spiker (1977) has indicated that reflector II is underlain by Late Pleistocene deposits and overlain by Holocene sands dating 9830 B.P. and younger. Cousins and others (1977) report being able to trace the unconformity shoreward from subarea 3 (see Fig. I-75) for at least 24 km. Above reflector II are sediments considered to have originated in the last marine transgression. They are discontinuous in extent, vary from a few meters to 18 m thick, and are Holocene in age. These sediments are considered to be derived from older shelf deposits, transgressed headlands, presently eroding shorefaces, and from inland sediments carried to the shelf by fluvial processes (Frank and Friedman 1973; Keeling and others 1975). Lagoonal sediments, if present, would be found within this sequence. At the top of this sequence is usually another unconformity (reflector I) representing truncated lagoonal/estuarine deposits resulting either from a transgressing shoreline or from post-transgression submarine scouring. Reflectors I and II merge locally, indicating probable loss of the pre-transgressive subaerial surface. Cousins and others (1977) report encountering horizontally bedded non-marine blue clay beneath reflector I in many places.

Above reflector I is the relatively thin discontinuous "surficial sand sheet." In most places reflector I is not detectable because it is obscured within theacoustical return of the ocean floor.

Major scarps have been noted along the New Jersey shelf compartment (Cousins and others 1977; Dillon and Oldale 1978; Emery and Uchupi 1972; Frank and Friedman 1973; Knott and Hoskins 1968; McClennan and McMaster 1971; Sears and Swift 1974; Swift and others 1972; and Veatch and Smith 1939). The locations of all major scarps are shown in Chart I-1b. As Dillon and Oldale have observed, the scarps dip toward the north. The deepest, the Nicholls scarp, (160 to 120 m) is probably pre-Late Wisconsin in age. The remaining three (namely, the Franklin, Fortune, and Atlantis scarps) date from the last transgression. As Frank and Friedman (1973) have cautioned, not all escarpments actually represent former shorelines. Shelf depressions may incorrectly be identified as escarpments (see for example Allen and others 1969). Detailed mineralogical and textural analysis of relict sands along the New Jersey shelf by Frank and Friedman (1973) provide additional evidence for terraces corresponding with the 72 m and 35 m escarpments. Extensive reworking and loss of fine sediments along these terraces supports their interpretation.

Investigations concerning Late Quaternary sea-level change along the New Jersey shelf have provided little information on sea-level positions before 9000 B.P. (Knebel and Spiker 1977; Stahl and others 1974; Stubblefield and others 1975; Stubblefield and Swift 1976; Stuiver and Daddario 1973). Table I-7 lists sea-level positions at 3,000-year intervals since the Late Pleistocene. More investigation is desperately needed before we have a good understanding of sea levels before about 8000 B.P.

Table I-7: Approximate sea-level positions at 3,000-year intervals for the New Jersey Shelf Compartment.

	Range	Best Estimate	Source(s)
3000 B.P.	5.2 m	5.2 m	1
6000 B.P.	13.2 m	13.2 m	1, 4
9000 B.P.	20-32 m?	20-32 m? (tilted shelf)	2, 3, 4
12,000 B.P.	40-60 m?	40-60 m? (tilted shelf)	2, 3
15,000 B.P.	70-90 m?	70-90 m?	2
18,000 B.P.	100-130 m?	100-130 m?	2

Sources

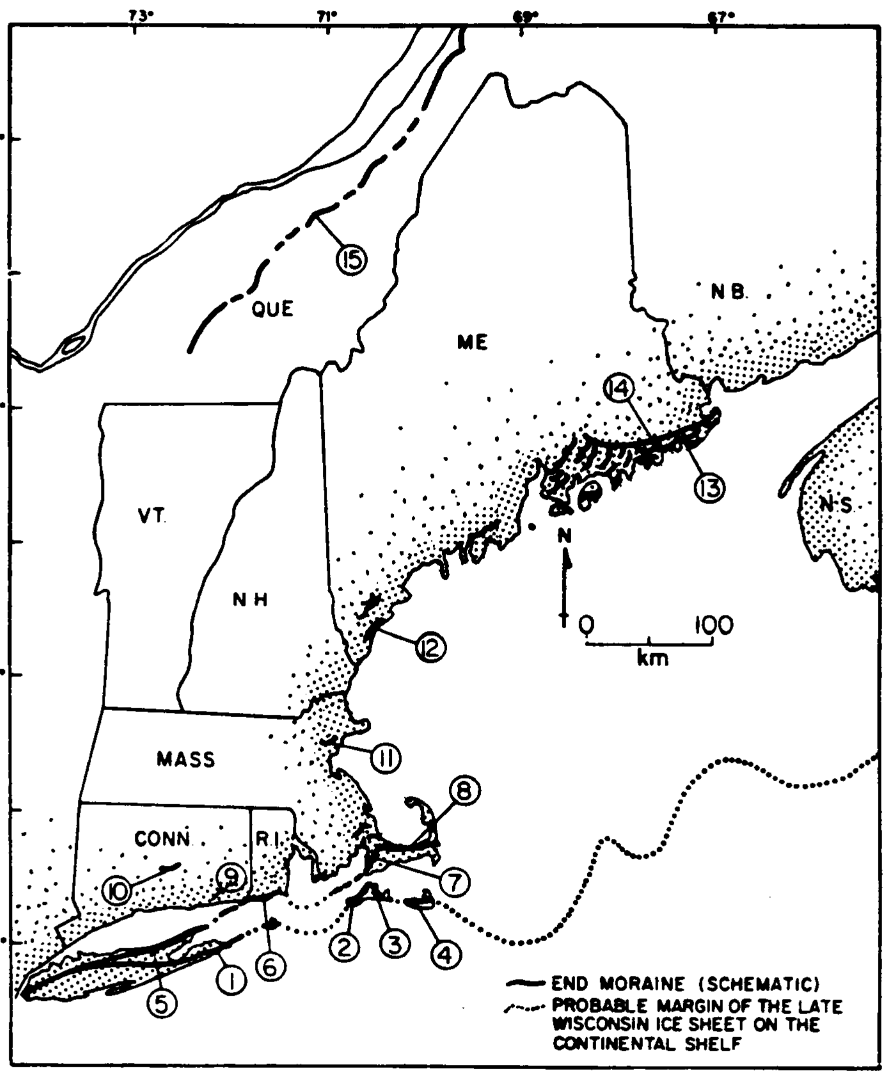
1. Stuiver and Daddario (1963)
2. Dillon and Oldale (1978)
3. Knebel and Spiker (1977)
4. Stahl and others (1974)

9.0 THE GLACIATED SHELF OFF THE NORTHEASTERN UNITED STATES

The continental margin northeast of New Jersey has been modified by Pleistocene glaciation to a greater extent than the shelf compartments discussed so far. The New Jersey and Delmarva continental shelves experienced increased river flows because of glacial meltwater but were not directly occupied by glacial ice during the Late Wisconsin (Borns 1973; Flint and Gilbert 1976; Pratt and Schlee 1969; Schafer and Hartshorn 1965). The inner Shelf along southern New England, Georges Bank, and the Gulf of Maine was occupied by glacial ice. Moraines, outwash plains, deltas, lakes, and other features formed during deglaciation. The boundaries and ages of many of these features inland in New England are difficult to define and often open to debate (Flint and Gilbert 1976; Schafer and Hartshorn 1965). The identification of submerged glacial features on the Inner Continental Shelf along this region is difficult because of the effects of transgression. The most difficult problem facing this study, however, revolves around local isostatic-eustatic relationships. Before reviewing Late Pleistocene-Early Holocene events on the northeastern CS, it is helpful to identify some of the important processes affecting our study area.

The maximum Late Wisconsin glacial limits shown in Fig. I-78 were reached sometime between 20,000 and 15,000 B.P. (Borns 1973; Flint and Gilbert 1976; Pratt and Schlee 1969; Schafer and Hartshorn 1965). As this figure partially indicates, the Ronkonkom moraine (Long Island), Block Island's surface till, and the Martha's Vineyard drift #5-Nantucket moraine are believed to represent the Late-Wisconsin glacial limits (Borns 1973; Flint and Gilbert 1976). The inferred position of terminal moraines further east (Fig. I-79) includes a lobe in the Great South Channel region as well as several lobes along the northern edge of Georges Bank (Pratt and Schlee 1969). The maximum limit across this entire region as shown on these figures does not necessarily represent the configuration of this glacial margin at any one time.

Deglaciation does not seem to have occurred simultaneously along this glacial margin (Borns 1973; Flint and Gilbert 1976; Oldale and others 1973; Tucholke and Hollister 1973). The Long Island section is believed to have been in retreat by 17,000 B.P. while the Buzzards Bay lobe does not seem to have begun to retreat until a millenium or two later (Borns 1973; Oldale and others 1973). Retreat of the Cape Cod and Great South Channel lobes may have been delayed slightly longer. By about 13,500 B.P., the ice margin had moved across the Gulf of Maine and was positioned slightly inland from the present coast of Maine (Fig. I-78). Marine transgression closely followed the ice margin during its retreat over the Gulf of Maine (Pratt and Schlee 1969; Tucholke and Hollister 1973). Consequently, no subaerial landforms are believed to have formed and the sills and ledges within the Gulf of Maine remained submerged.



1.	Ronkonkoma Moraine	
2.	Zack's Cliff	15,300 ± 800 yrs. B.P.
3.	Vineyard Moraine	
4.	Nantucket Moraine	
5.	Harbor Hill Moraine	
6.	Charlestown Moraine	
7.	Buzzard's Bay Moraine	
8.	Sandwich Moraine	
9.	Rogers Lake	14,240 ± 240 yrs. B.P.
10.	Middletown Readvance	
11.	Fresh Pond Moraine	<14,000 yrs. B.P.
12.	Kennebunk Advance	13,200 ± 120 yrs. B.P.
13.	Pond Ridge Moraine	13,520 ± 200 yrs. B.P.
14.	Pineo Ridge Readvance	-12,700 yrs. B.P.
15.	Highland Front Moraine	-12,600 yrs. B.P.

Fig. I-78
Schematic map illustrating the location of some Late Wisconsin end moraines in New England and on the continental shelf. After Borns (1973).

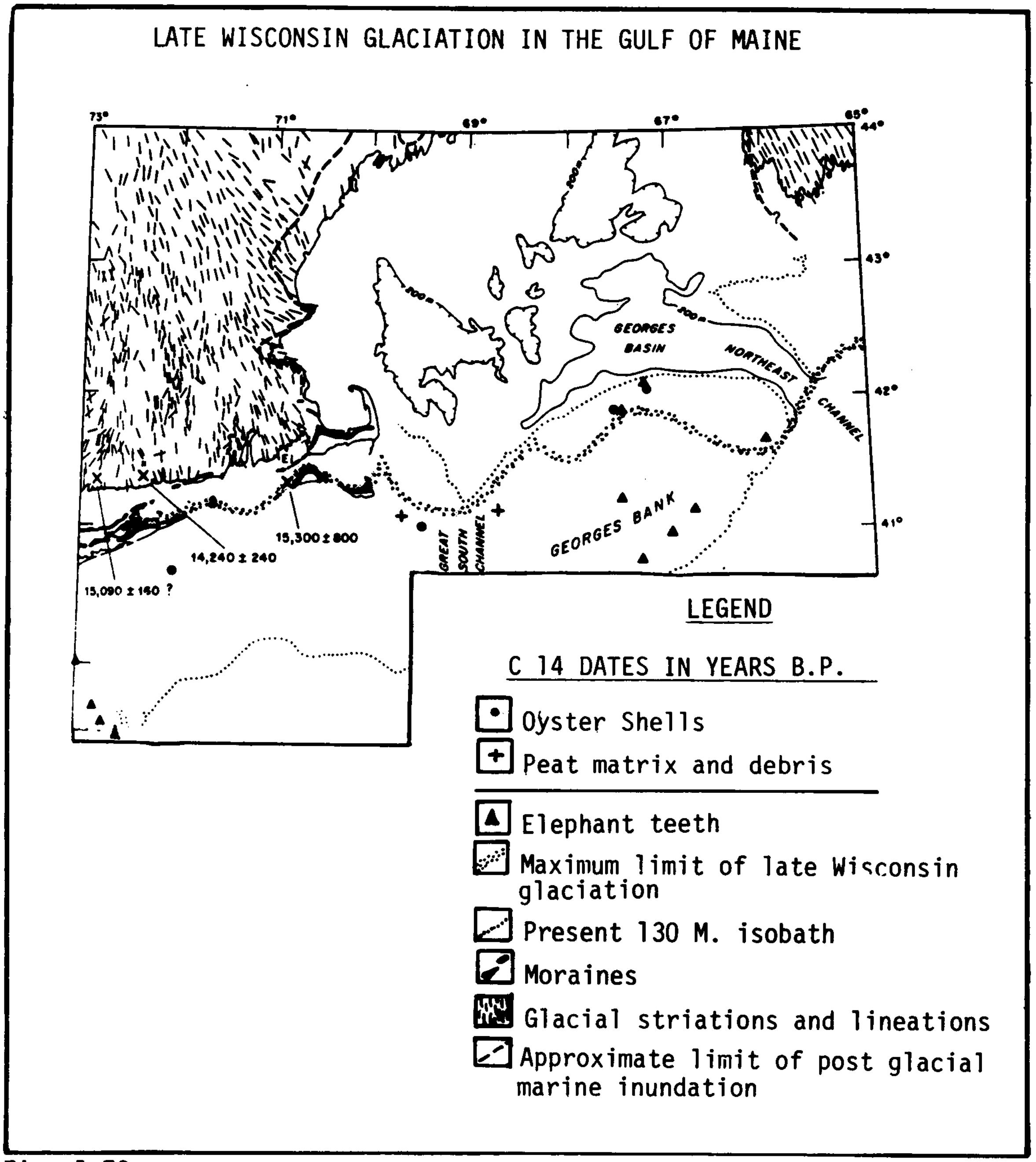


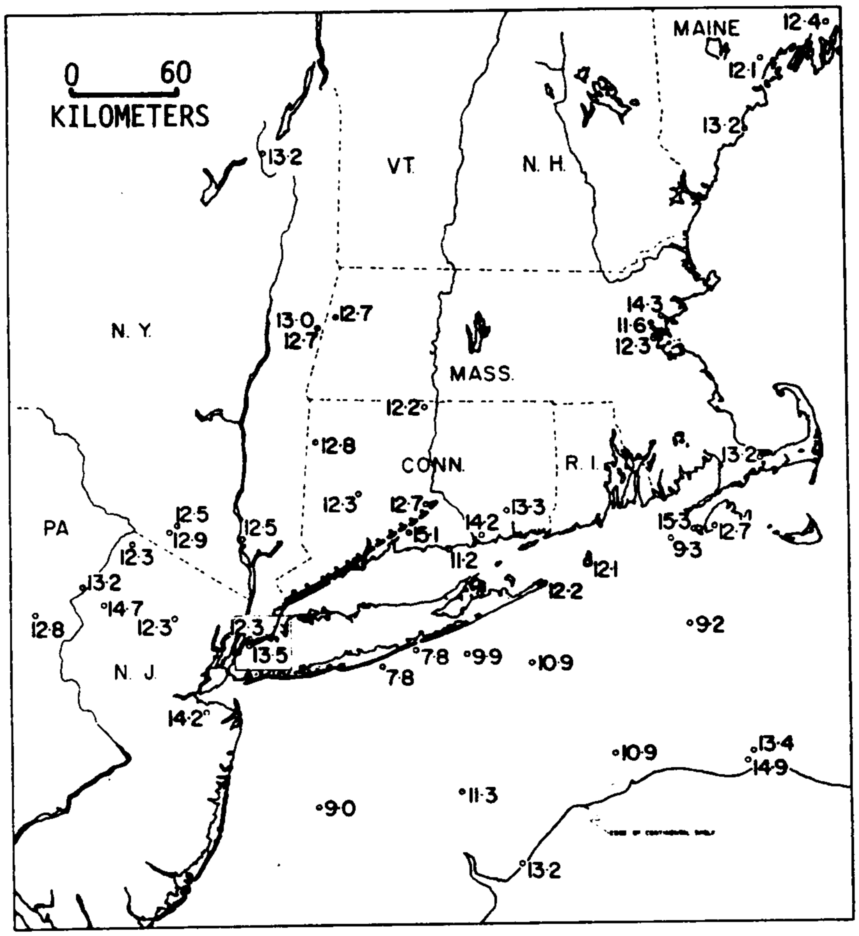
Fig. I-79 Maximum limit of Late Wisconsin glaciation and selected glacial features. The maximum limit of glaciation does not necessarily represent the configuration of this boundary at any one time. Not all the glacial striations and lineations relate to late Wisconsin glaciation but in general they offer a pattern for the direction of flow af glacial ice. Extent of postglacial marine inundation shown by dotted line along the northern New England coast. After Tucholke and Hollister (1973).

Suitable organic material for radiocarbon dating is hard to obtain, and consequently many of these glacial events are not accurately dated. The marine inundation of coastal Maine is one exception to this rule. Emergence of portions of the CS is hard to date, however, since peat and other surface plants seem to lag behind initial subaerial exposure by centuries or even several millenia (Newman 1977; Stuiver and Borns 1975). Fig. I-80 is a compilation by Newman (1977) of the oldest late-glacial radiocarbon dates in the Northeast.

Synchronic retreat of the ice sheet along the New England shelf does not appear to be the rule. Many factors influenced glacial retreat and it appears that different lobes retreated at slightly different times. Variation among local readvances has been well documented for inland portions of the ice sheet (Borns 1973; Schafer and Hartshorn 1965; Stuiver and Borns 1975) which illustrates the potential individuality between some sections.

It is beyond the scope of this project to review inland glacial events. Fig. I-78 shows the position of some important Late Wisconsin moraines. See Fig. I-81 for a discussion of glaciation in northeastern North America. At the time of maximum Late Wisconsin glaciation, land relationships from the glacial margin northward were isostatically depressed. Evidence of glacial downwarping has been identified for northeastern Massachusetts (Kaye and Barghoorn 1964), and along much of coastal Maine (Bloom 1960, 1963; Schnitker 1974; Stuiver and Borns 1975; Tucholke and Hollister 1973). Besides isostatic downwarping, there are some indications that the middle portion of the southern New England shelf was uplifted during the Late Wisconsin. Edwards and Merrill (1977) review some of the evidence which they believe indicates that a glacial forebulge existed. They point out the concave profile of the Hudson Shelf Valley (Fig. I-82) and suggest that it represents forebulge down-cutting. In addition, they note that scarps (old shorelines) drop to greater depths along the southern New England shelf. Dillon and Oldale (1978) also discuss shoreline tilt in detail and use submerged "shoreline" trends to define a shelf block that may have moved somewhat independently during the Late Pleistocene. Last of all, Edwards and Merrill (1977) note that canyon heads start at greater depths along this portion of the shelf compared to those found farther south.

Besides the evidence Edwards and Merrill (1977) present, the preservation of several small drainage systems, such as the Long Island Shelf Valley system (Swift and others 1972), also lends some support to the collapsing-forebulge hypothesis. Small drainage systems like that of the Long Island Shelf Valley do not usually remain intact after marine transgression. The preservation of the Long Island Shelf Valley, however, may have been enhanced by more rapid transgression which resulted when the forebulge "deflated" during deglaciation. Preservation of the Long Island Shelf Valley system may also have been enhanced by adjacent protective headlands which partly shielded it from destructive transgression (Swift, personal communication; Swift, Kofoed and others 1972). Last of all, sediment for infilling the system may have been in short supply. The



KEY

∘ Location of C^{14} dates

∘10·9 Age in C^{14} years (x 10^{3}) before present

Limits of Einhurst Readvance (dashed where inferred)

Fig. I-80

Distribution of oldest dates associated with the Late Wisconsin in southern New England and adjacent areas. Dates from the Long Island continental shelf illustrate the paucity of Late Pleistocene organic material collected and radiocarbon dated from this area compared with mainland regions. After Newman (1977). See Table I-8 for additional information on radiocarbon dates.

Table I-8; Tabulation of oldest Late-Glacial radiocarbon dates reported from various localities in the northeastern United States.

Lab No.	Date (yr B.P.)	Location	References
W-1181	15,300 ± 800	Martha's Vineyard	20
Y-446a	15,090 ± 160	Southern Connecticut	15
I-2544	14,850 ± 250	Continental Shelf edge	39
I-4162	14,720 ± 260	Northwestern New Jersey	40
W-735	14,250 ± 250	Boston	12
Y-950/51	14,240 ± 240	Southern Connecticut	15
W-1457	14,150 ± 450	East-central New Jersey	41
RL-157	13,470 ± 380	Northwestern Long Island	this paper
I-2473	13,420 ± 210	Continental Shelf edge	39
Y-447d	13,290 ± 120	Southeastern Connecticut	15
OWU-430	13,235 ± 1620	Northeastern Pennsylvania	24
I-4648	13,200 ± 220	Cape Cod	42
I-2545	13,200 ± 210	Continental Shelf edge	39
Y-2208	13,200 ± 120	Southwestern Maine	43
I-4986	13,150 ± 200	Near Lake George, N.Y.	21
Y-2247b	12,960 ± 180	Massachusetts-N.Y. border	44
L-1157A	12,850 ± 250	Southeastern New York	21
SI-1341	12,760 ± 135	Northeastern Pennsylvania	24
RL-245	12,750 ± 230	Northwest Connecticut	45
W-710	12,700 ± 300	Martha's Vineyard	15
W-46	12,700 ± 280	Central Connecticut	46
OWU-481	12,680 ± 480	Western Massachusetts	47
Y-2247a	12,680 ± 200	Massachusetts-N.Y. border	44
I-4137	12,530 ± 270	Southeastern New York	48
L-1141	12,500 ± 600	Southeastern New York	49
W-2117	12,380 ± 350	Southwestern Maine	50
Y-1865	12,330 ± 250	West-central Connecticut	51
W-2562	12,300 ± 300	Northern New Jersey	52
GXO-330	12,290 ± 500	Northern New Jersey	53
W-1801	12,275 ± 350	Boston	54
I-5663	12,270 ± 180	Northwestern Long Island	this paper
W-828	12,200 ± 350	North-central Connecticut	55
L-678	12,100 ± 300	Southwestern Maine	56
W-255	12,080 ± 200	Block Island	15
W-1801	11,600 ± 300	Boston	54
Y-1178	11,240 ± 160	Coastal Connecticut	9
	All younger shelf dates		39

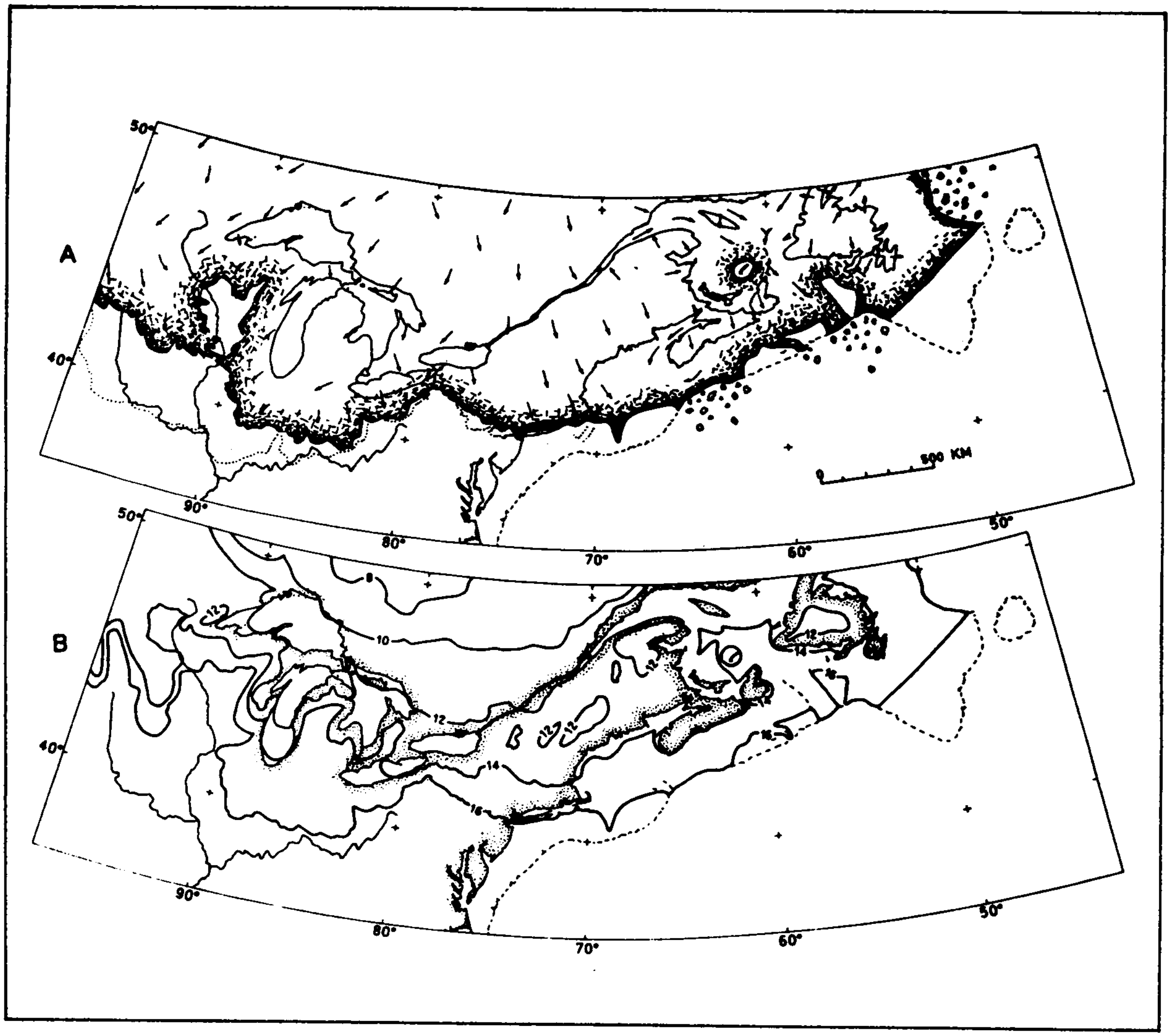


Fig. I-81

Late Wisconsin glaciation in northeastern North America. Extent of Wisconsin glaciation shown in (A) with some ice bergs indicated where the open ocean was reached. Arrows indicate inferred direction of ice movement. Dotted line gives Pre-Wisconsin glacier fronts. Stages of retreat of the Late Wisconsin glaciers is illustrated in (B) for two thousand year intervals. After Emery and Uchupi 1972.

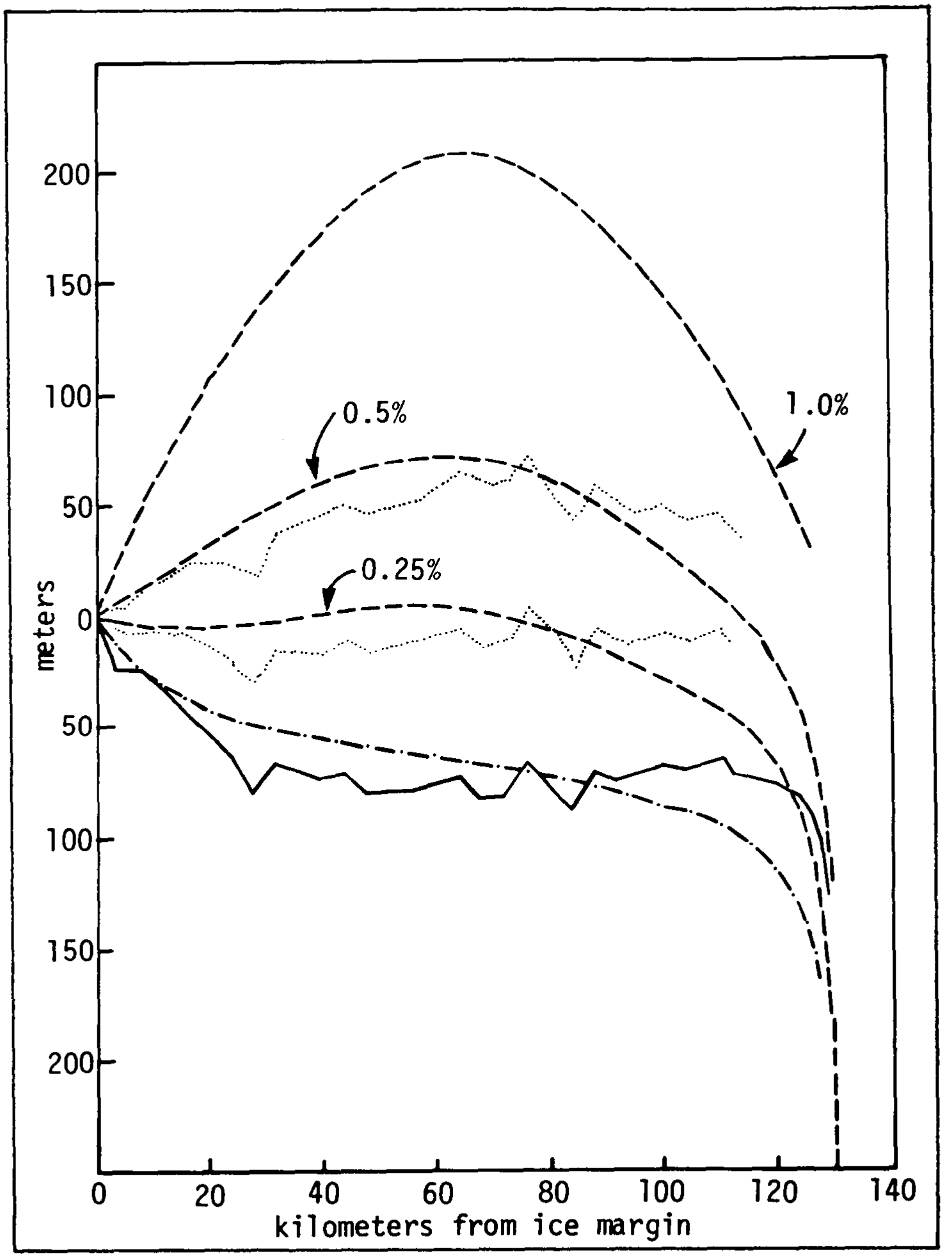


Fig. I-82

Profiles of the present Hudson River Channel (solid line) plotted against an idealized profile (dash-dot line) using three forebulge slopes (0.25, 0.5 and 1.0%). Edwards and Merrill feel that a slope of 0.25% seems realistic even if the depth of the thalweg is deeper than shown. The forebulge was calculated using the Heiskanen and Vening-Meinesz (1958) equation as modified be McGinnis (1968). After Edwards and Merrill (1977).

Hudson and Block Shelf Valleys would have acted as effective sediment sinks, collecting their share of longshore sands and minimizing the amount of sediment reaching the Long Island valley system by longshore processes during transgression.

Edwards and Merrill (1977) developed a set of isolines for visualizing the amount of forebulge affecting portions of the shelf. Fig. I-83 gives their reconstruction of the forebulge, based on estimates of several important factors and using specific forebulge equations (Edwards and Merrill 1977). Theoretically, the forebulge should reach its highest point about 66 km in front of the glacier (Fig. I-84, Edwards and Merrill 1977). At the time of maximum forebulge uplift, a trough of fresh or brackish water possibly existed along the present-day Inner Shelf region (Edwards and Merrill 1977). Glacial downwarping and forebulge uplift make it extremely difficult to reconstruct land-ocean relationships during the Late Pleistocene. By about 12,500 B.P., however, land relationships are believed to have been restored essentially to those found today (Edwards and Merrill 1977; Stuiver and Borns 1975).

The moraines, tills, and glacial outwash deposits on Long Island, Block Island, Martha's Vineyard, and Nantucket provide information on the southern limit of Late Wisconsin glaciation. In the areas between these islands and from Nantucket Shoals eastward across Georges Bank, the limit of glaciation has been inferred from sediment texture and shelf morphology (Pratt and Schlee 1969; Schlee 1973; Schlee and Pratt 1970). By plotting the seaward boundary of abundant sandy gravel and noting the position of shoals, Pratt and Schlee (1969) delineated the maximum limit of glaciation as shown in Fig. I-85. It is possible that some of the reworked outwash and till may be older than the Late Wisconsin. Holocene and glacio-fluvial erosion have made it extremely hard to identify ice-contact deposits on the CS. In the Gulf of Maine, reworked "till" (loess of silt and clay) is recovered from banks and ledges while deeper areas sometimes contain till-like materials from beneath Holocene deposits of silt and clay (Pratt and Schlee 1969; Tucholke and Hollister 1973). Fig. I-86 is a schematic representation based on the texture of surface sediments along the northeastern CS (Pratt and Schlee 1969; Schlee 1973). The figure shows the inferred Late Pleistocene geomorphic units responsible for shelf sediments in the Northeast. Fig. I-87 shows the same area after Holocene transgression, and the redistribution of sediment.

The distribution of bimodal sandy gravel (or gravelly sand) outlines those areas which have received considerable erosion and redistribution since the last transgression. Fig. I-87 shows these locations as well as areas which have received insignificant erosion (those with matrix-rich gravel). The areas with bimodal sandy gravel in our porject area correspond to shelf uplands (for example the regions along the New Jersey shelf, the Long Island Shelf Valley, Nantucket Shoals, and Georges Bank) and low areas which concentrate tidal and storm currents (such as the Great South Channel and Hudson Shelf Valley). Such areas may be useful for making surface collections of artifacts and mega-

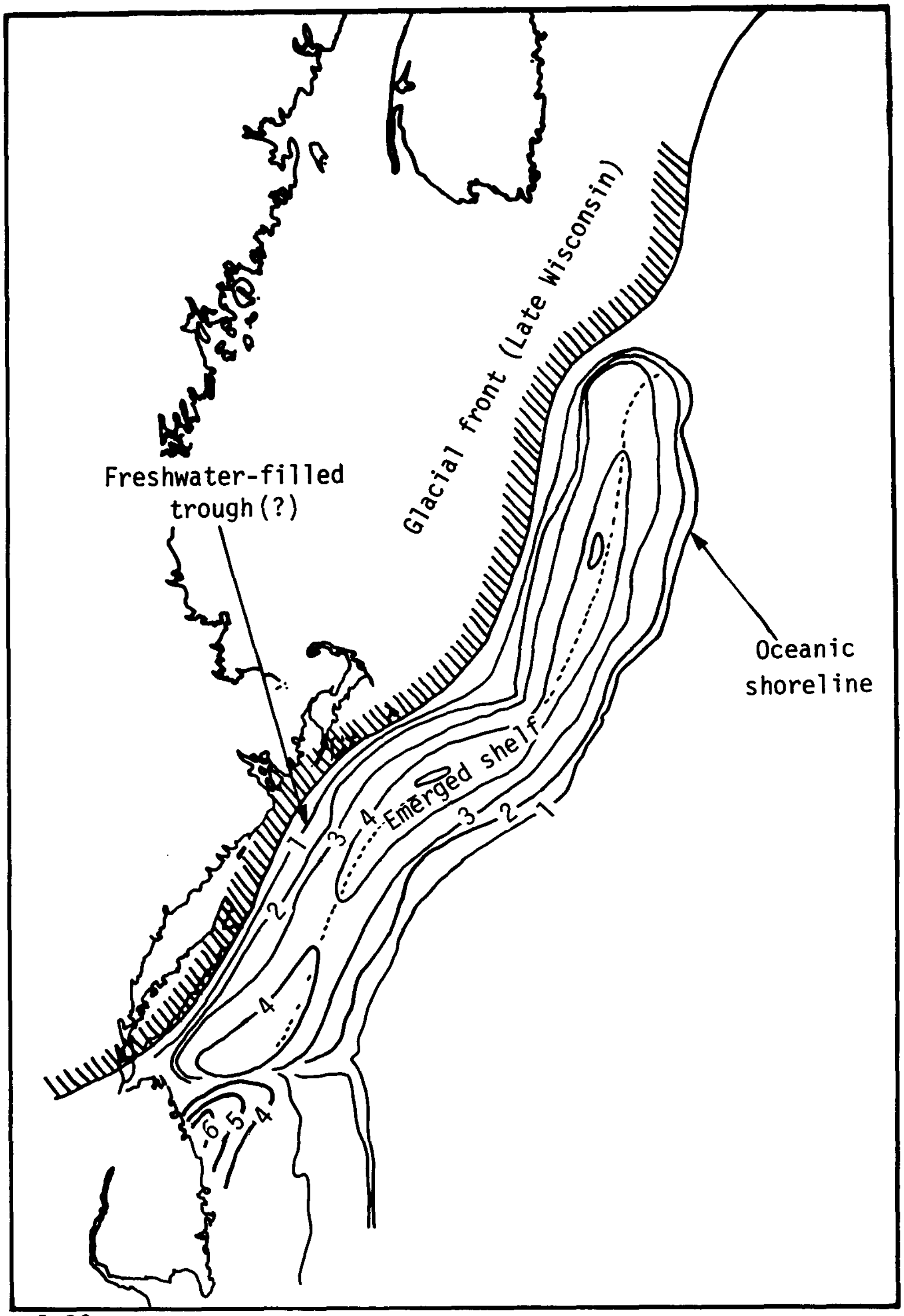


Fig. I-83 Hypothesized isolines for the amount of forebulge uplift during maximum Late Wisconsin glaciation (Edwards and Merrill 1977). Greatest uplift hypothetically occurred slightly landward of dotted line. The northern New Jersey shelf would have received the greatest uplift according to this diagramatic model. The trough created by the forebulge along the glacial front was probably occupied by freshwater although it was open to the sea. After Edwards and Merrill (1977).

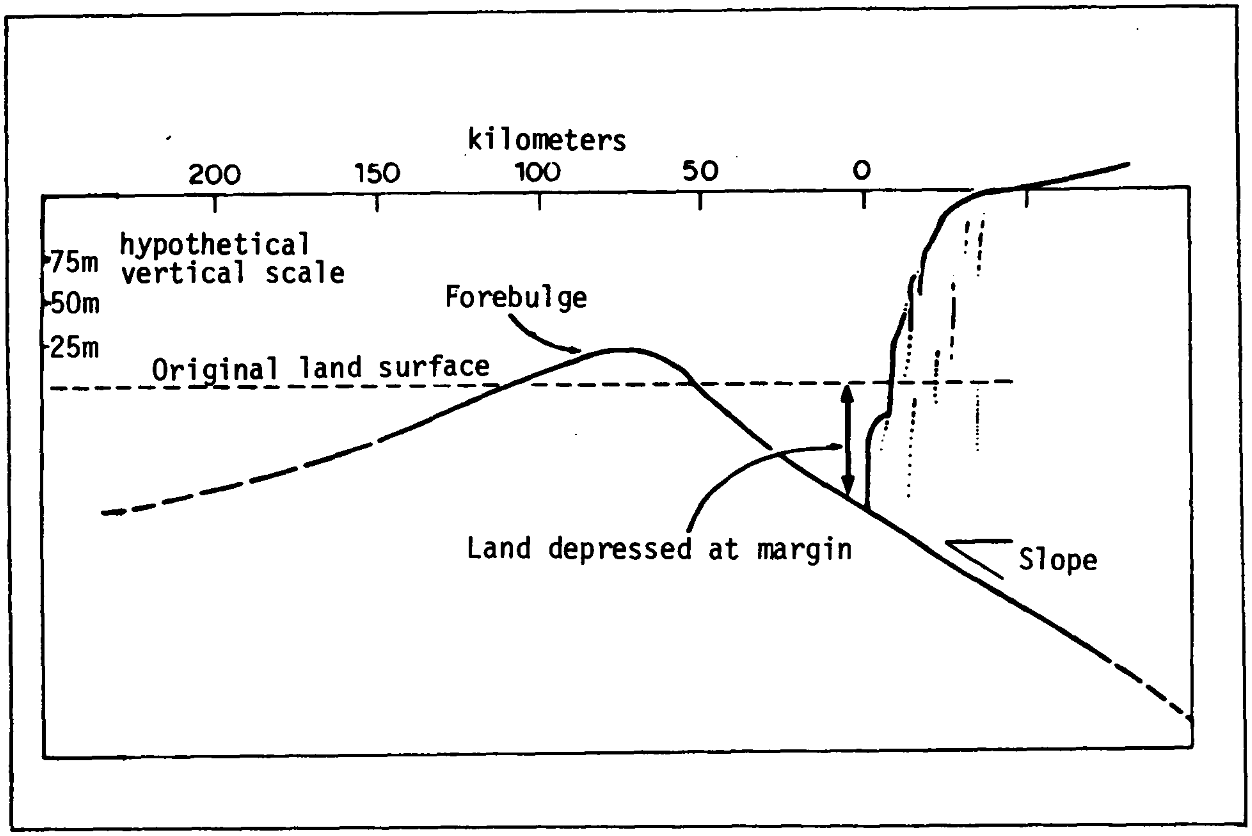


Fig. I-84

Forebulge expression based on theoretical models (Edwards and Merrill 1977). Reconstruction of land surface relationships requires knowledge of ice thickness, amount of isostatic depression, and sea-level positions which can only be crudely estimated at present. The vertical scale shown at the left is based on a .3% forebulge slope. After Edwards and Merrill (1977).

faunal remains since these items, if present, would become part of the gravel lag.

Drainage patterns along the subaerial shelf at the time of maximum glaciation would have appeared similar to those in the schematic representation given in Fig. I-86 (Schlee 1973). Except for the ancestral Hudson, Long Island, and Block Shelf Valleys, the location of all other streams and rivers has been inferred from canyon heads so that their shelf paths (valleys) are not accurately identified. Georges Bank would have consisted of an outwash plain containing numerous streams and rivers rivers. North of the Late Wisconsin terminal moraine, some drainage rearrangement occurred as early valleys became blocked with glacial debris. Along Long Island Sound and Rhode Island Sound, and in Cape Cod Bay, glacio-lacustrine sediments were deposited. By 12,000 B.P., however, most of these lakes had drained (Oldale and others 1973).

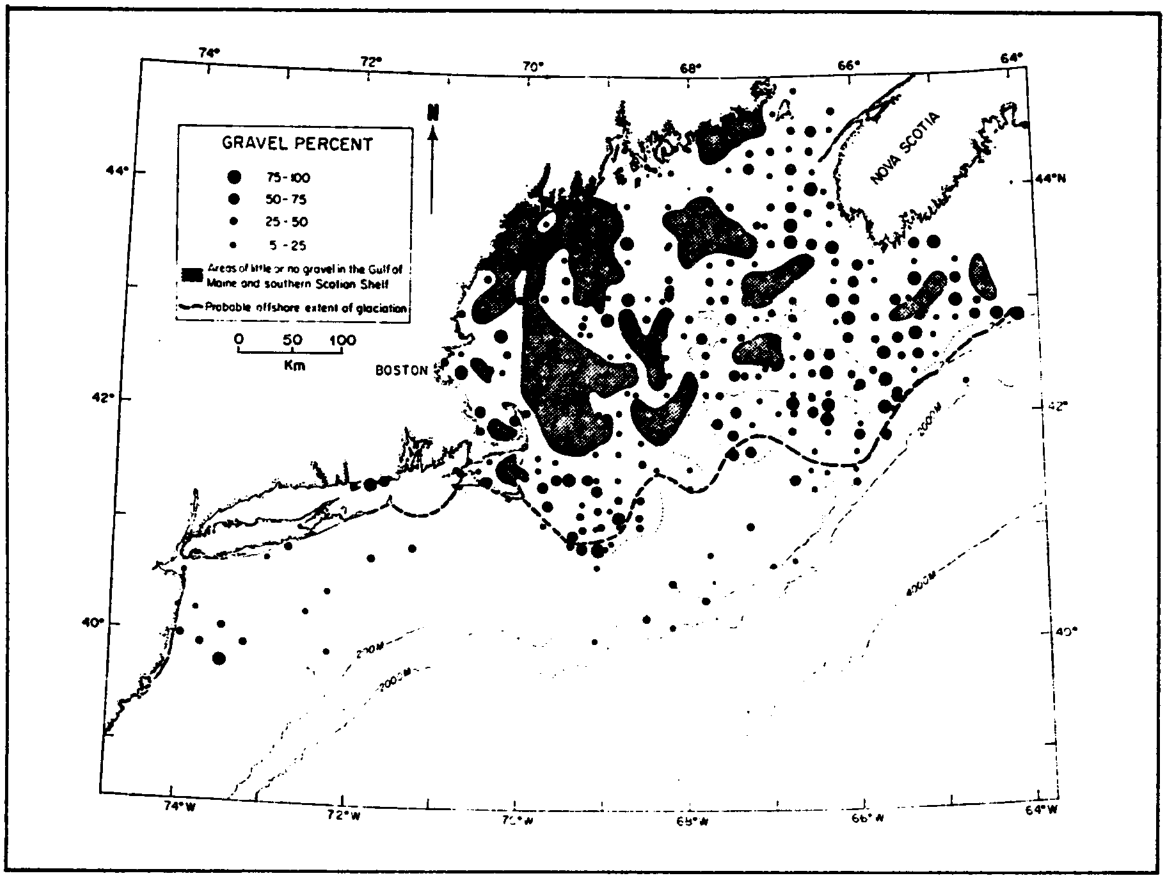


Fig. I-85a

Distribution of gravel on the continental shelf off New England and the probable offshore extent of Wisconsin glaciation. The gravel on the northern New Jersey shelf and the Long Island shelf is probably the result of current scour erosional shoreface retreat. After Pratt and Schlee (1969).

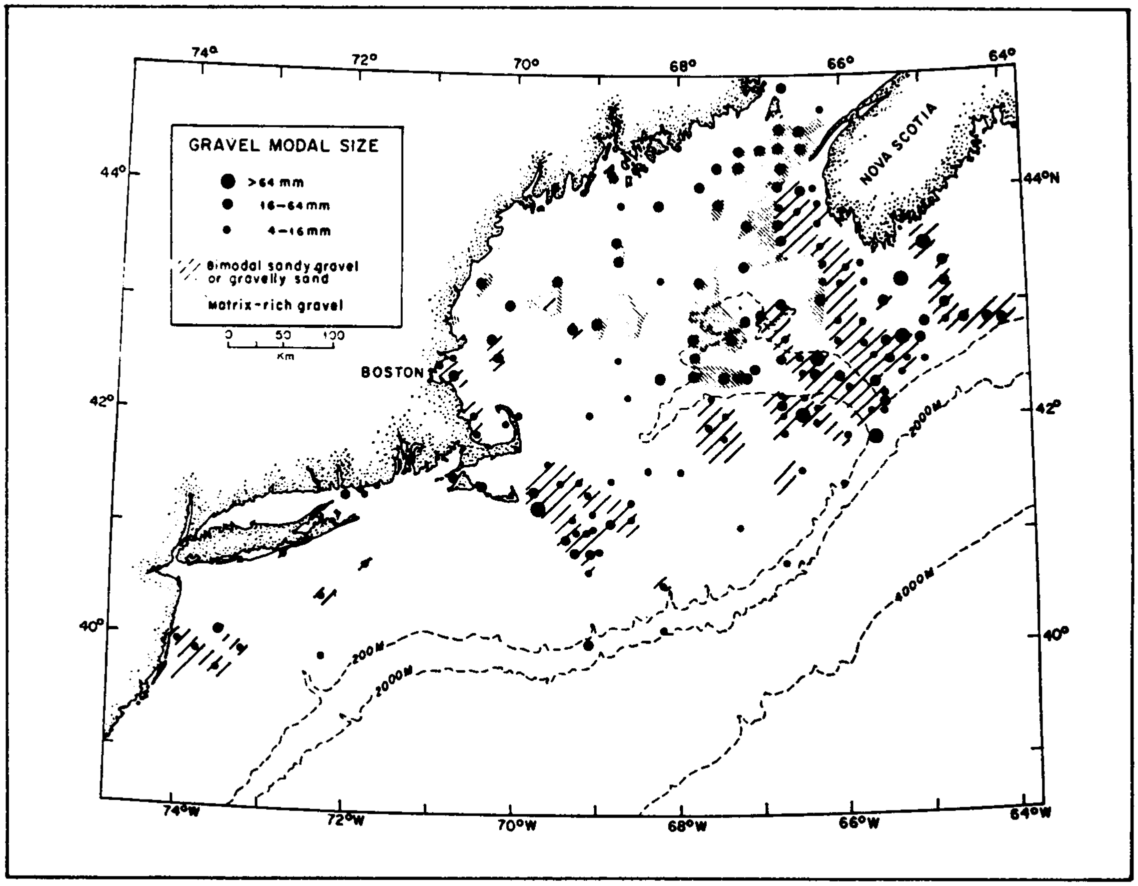


Fig. I-85b

Distribution of bimodal gravels (sandy gravel or gravelly sand) and matrix-rich gravel on the northeastern United States continental shelf. The bimodal gravel in general represents areas either scoured by storm and tidal currents (for example the Great South Channel, Georges Bank) or truncated during erosional shoreface retreat (coastal Massachusetts or the inner shelf of New Jersey). Matrix-rich gravel represents glacial till deposited with little erosion or transport. After Pratt and Schlee (1969).

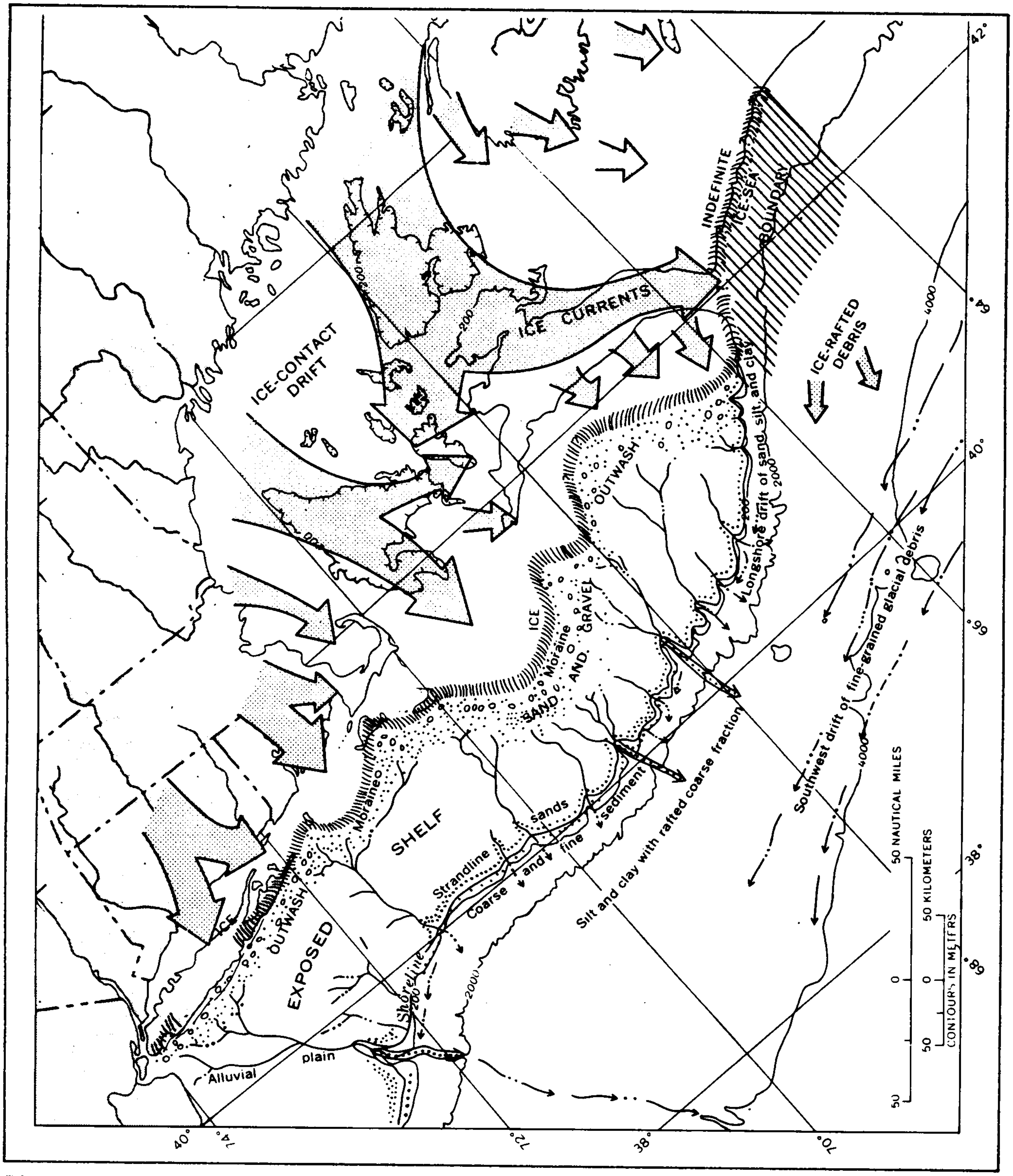


Fig. I-86

Schematic representation of the northeastern continental margin illustrating the geologic processes responsible for modifying its topography during the Late Pleistocene. After Schlee (1973).

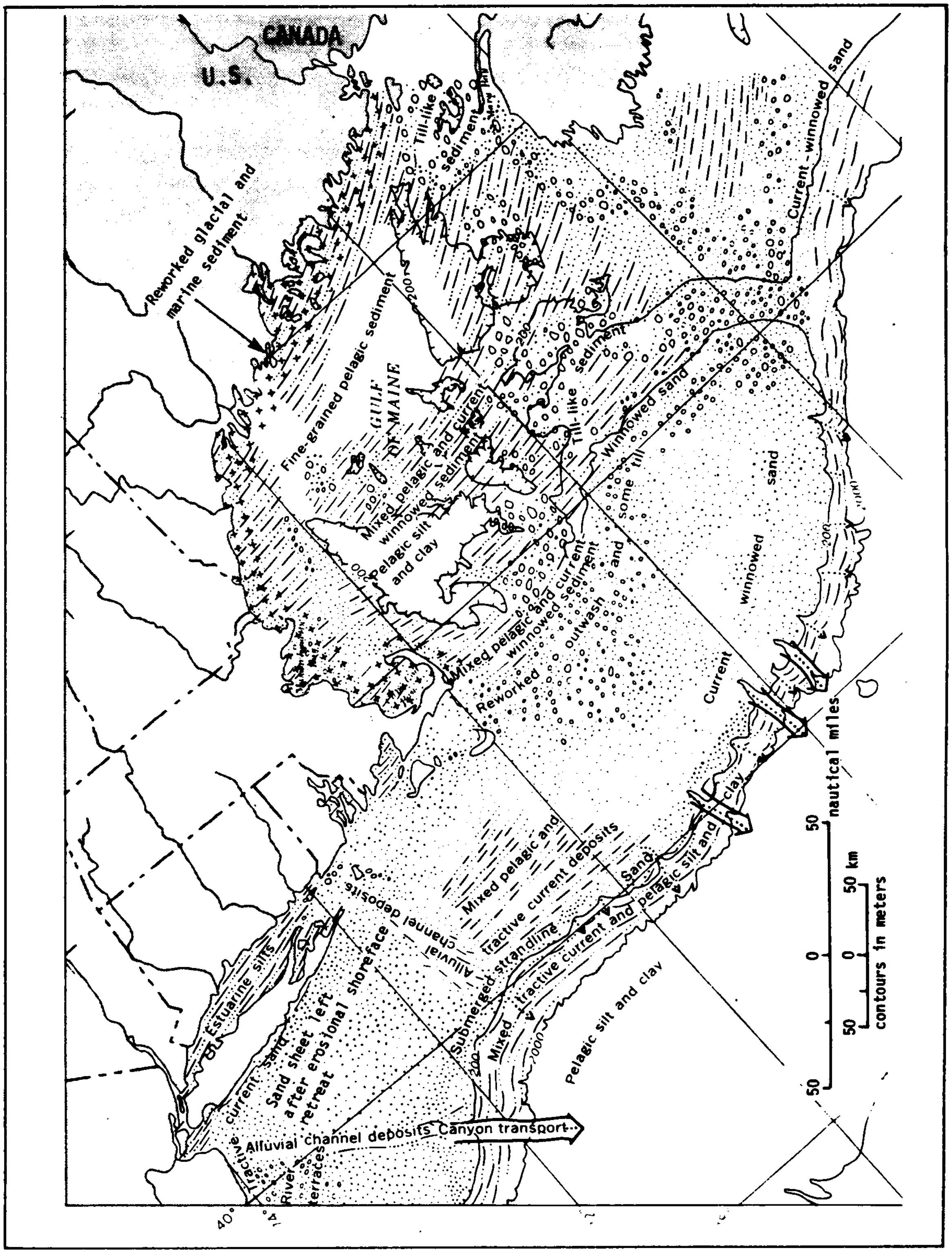


Fig. I-87

Schematic representation of the northeastern continental margin illustrating sediment texture after deglaciation and Holocene transgression. Adapted from Schlee (1973).

10.0 LONG ISLAND SHELF

Boundaries for the Long Island shelf compartment are the Hudson Shelf Valley to the south and the Block Shelf Valley to the north (Fig. I-88). Long Island Sound is included in this shelf compartment. The Holocene evolution of the Sound was quite different than that of the CS south of it. Considerable research has been devoted to these areas because of interest in the effects of urbanization and pollution (CNA 1977; TRIGOM 1976). Despite these investigations, the Late Quaternary evolution of the area is only poorly known. Isostacy and other glacially induced changes have added to the geological complexity of the area, making it hard to correlate and interpret many features. These processes were discussed in detail in the previous section.

Briefly, in review, this area was depressed by glacial ice-loading during the Late Wisconsin glacial maximum. A forebulge, however, raised a portion of the Middle and Outer Shelves, forming a body of fresh or low-salinity water between it and the glacial front. The maximum extent of glaciation during the Late Wisconsin is marked by the Ronkonkoma moraine (Borns 1973; Flint and Gilbert 1976; Newman 1977). Glacial retreat was followed by inundation until the ice front lay approximately along the present shoreline of Connecticut. At about 12,000 B.P., glacial rebound began to affect this area, restoring land elevation relationships nearly to those found today. Although the elevation of the forebulge is unknown, it is considered to have disappeared during the Early Holocene. The deflation of the forebulge in this region acted as a local factor in decreasing the rate of marine transgression over the subaerial surface. The apparent "preservation" of several drainage systems on the Middle and Outer Shelves along this region may be a direct outcome of rapid marine transgression. Additional evidence concerning important local events during the Early Holocene is discussed further on in this subsection.

Lagoonal deposits have been encountered along this shelf compartment (Sanders and Kumar 1975a, 1975 b), whose bathymetry, place names, and other important features are shown on Fig. I-88. Sanders and Kumar (1975a, 1975b) have investigated a portion of the Inner Shelf near Fire Island (Fig. I-89). They used coring and seismic profiling to study the effects of transgression on barrier islands. The authors were particularly interested in whether transgression caused the barriers to migrate slowly inland or whether the sea "jumped" the barriers, leaving them intact as submarine ridges. Evidence is presented for both processes on the basis of the relationship of several units consisting of lagoonal deposits and the overlying sand sheet. Where lagoonal deposits could be found submerged intact, the shoreline is inferred to have jumped. If lagoonal muds were absent, the barriers are viewed as migrating landward in tank-tread fashion with erosional shoreface retreat destroying the lagoonal sediment sequence.

On the basis of the evidence from the other shelf compartments reviewed so far, Sanders and Kumar's criteria for inferring shoreline jumping are judged to be inadequate (see for example Sheridan and others 1974). There is little support for their view that preservation of lagoonal sediments requires shoreline jumping. Similarly, their concept that continuous shoreline migration completely destroys back barrier deposits is also in error. As our model has illustrated time and time again, local rates of sea-level rise, sedimentation, pre-transgressive topography, storm climate, and other factors may combine during barrier migration to preserve lagoonal deposits. Consequently, the conclusions given by Sanders and Kumar (1975a, 1975b) need critical reexamination.

Further, the evidence Sanders and Kumar (1975a, 1975b) present for an intact shoreline barrier system 7 km seaward of the present barriers is open to other interpretations, as they themselves have indicated. Sanders and Kumar (1975a, 1975b) sometimes regard the bottom sediments as relict and neglect the effect of modern hydraulic processes. Their interpretation of the "shoe string" sands as representing the path of a coast-parallel-migrating tidal inlet is more reasonable, given the evidence they have at hand. Preservation of such a feature would be possible whether the shoreline migrated landward or "jumped" to a new position. Truncation of the upper portion of a barrier complex during barrier migration does not penetrate as deeply as tidal inlet scour and thus tidal scours may remain intact.

The truncated "lagoonal deposits" found in this area can generally be regarded as preserved portions of a low-energy environment most probably associated with back-barrier regions. Thin beds of salt-marsh peat dating 7750 B.P. and 7585 B.P. were encountered 18 to 19 m below sea level. This peat overlies a small section of lagoonal muds which themselves rest upon a greenish-gray silty sand. The characteristics of the silty sand directly beneath the lagoonal sediments (Sanders and Kumar 1975a, 1975b) seem to indicate possible subaerial exposure. These sediments may represent the pre-transgressive subaerial surface preserved beneath lagoonal muds. The absence of basal peat directly covering this surface may indicate some reworking of the lagoon shoreline before burial. Peat layers have been encountered in only 2 of about 50 cores taken in the area. Thus, shoreface erosion may have removed most of these deposits during transgression.

The barrier system forming the southern shore along Long Island today is interpreted by Rampino and Sanders (1977) as being superimposed upon older Late Pleistocene barriers. They regard the present barriers as welded upon these older features after following step-wise (that is, hopping) shifts in position from further out on the CS. So far, there is little substantive evidence for this type of "migration."

McKinney and Friedman (1970) have investigated a fairly large section across the Middle and Outer Shelves off Long Island (Fig. I-90). Although most of their research concentrates on surface sampling, some of their conclusions provide insight into the Holocene evolution of this

LEGEND

Delta

Shoal retreat massif and possible cuesta-like features.

Surface valley on shelf, indicating that buried estuarine and possibly river valley deposits probably exist within the general vicinity.

Fig. I-88
Major Late Pleistocene-Holocene features on the Long Island Continental Shelf.

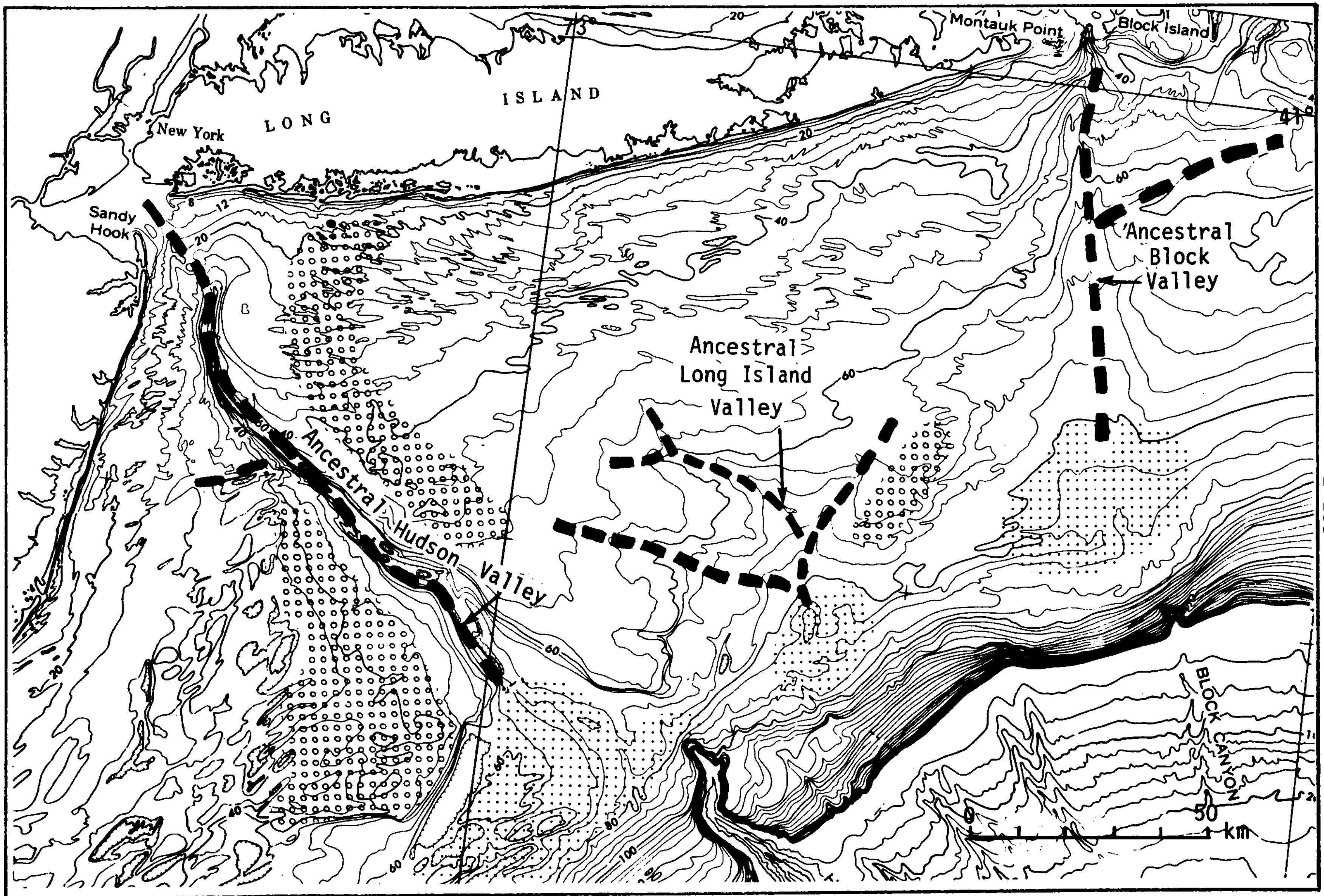


Fig. I-88 Major Late Pleistocene-Holocene features on the Long Island Continental Shelf. Features compiled from the following sources: Knott and Hoskins (1968); McKinney and Friedman (1970); McMaster and Ashraf (1973a, 1973b, 1973c); Pratt and Schlee (1969); Swift (1977); Swift and Sears (1974); Swift and others (1972). Contours in meters.

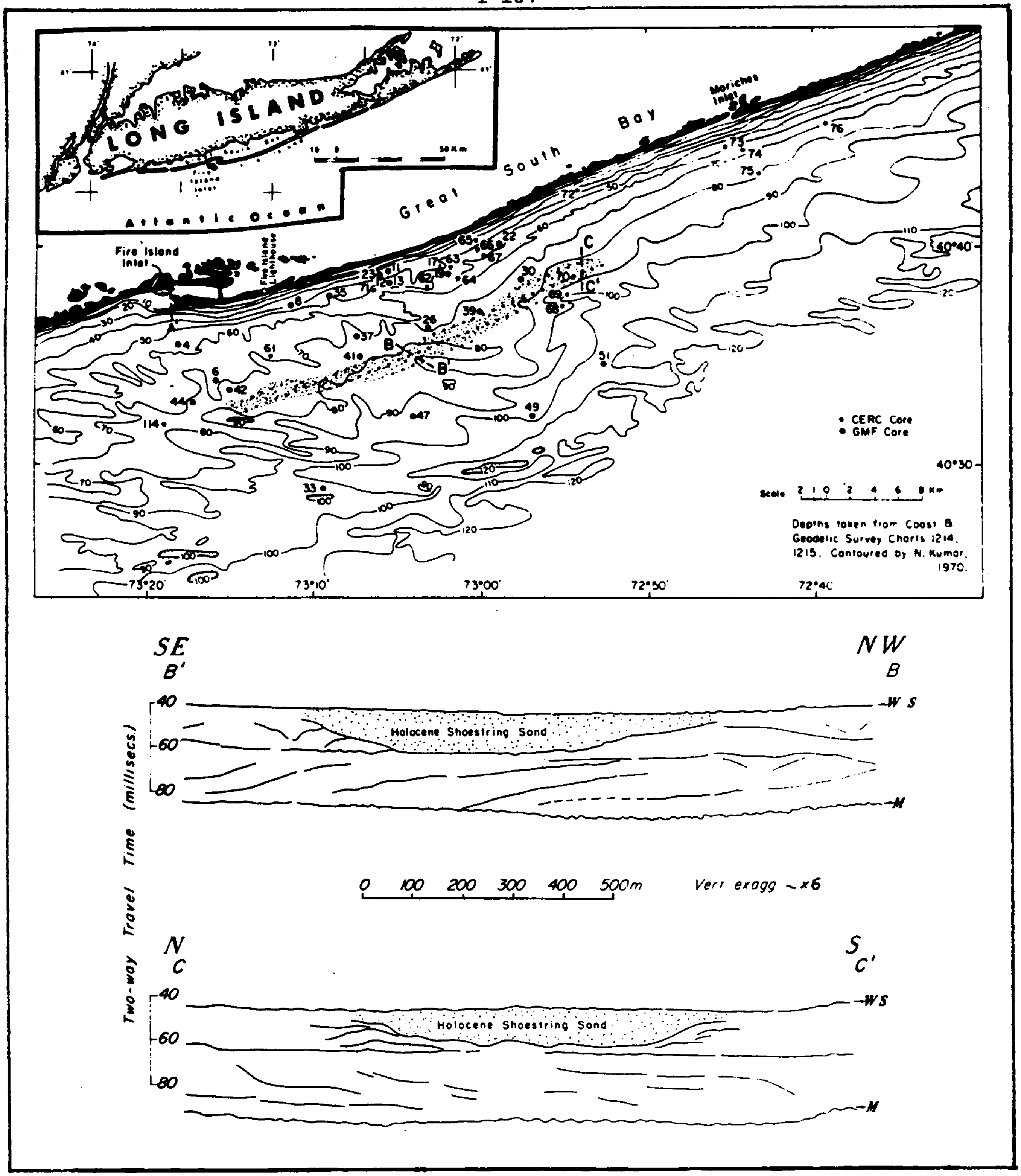


Fig. I-89

Area studied by Sanders and Kumar (1975a) along the inner shelf of Long Island. Stippled area probably represents scour trench left by a tidal inlet after some coast parallel migration. After Sanders and Kumar (1975a).

region. McKinney and Friedman identify the remains of three relict "drainage patterns" within their study region (Fig. I-90). The outermost (deepest) "drainage pattern" forms a trellis pattern and seems to be fairly well intact. It is not easy to understand the processes responsible for the preservation of this drainage system. Low sediment imput combined with rapid submergence (forebulge collapse?), stream capture, headland protection, and possible topographical controls by bedrock may be partly responsible. Not enough data are available to enable us to determine which factor acted to preserve this "drainage pattern," although some hypotheses have been given above.

According to McKinney and Friedman (1970:219), the innermost "drainage pattern" consists of well-developed west-to-east parallel channels which used to drain into Block Channel. Much of this inner "drainage pattern" is artificial and represents the erosion and redistribution of bottom sediments since transgression. To regard these bottom features as relict is to neglect the effect transgressive and submarine processes have had on the bathymetry of the area. It is important to note, however, that since McKinney and Friedman (197) wrote this article, they have realized that the Inner Shelf "drainage pattern" is not subaerial but has been formed by submarine processes (Swift, personal communication). Of the patterns recognized by McKinney and Friedman (1970), the middle "drainage pattern" (supposedly draining to the southeast), is the least easy to discern. Criticisms similar to those mentioned above may be leveled against it. Of the three patterns, only the outermost seems to be truly relict and most probably represents the remains of a Late Pleistocene-Early Holocene river system. This system has been designated the Long Island Shelf Valley system by Swift and others (1972).

McKinney and Friedman (1970) define two subregions based on textural parameters. Shoreward of a depth of about 46 m, bottom sediment consists generally of a clean sand facies. Seaward of this depth is a predominantly muddy sand facies on the Middle Shelf (46-64 m). The Inner Shelf sands (0-46m) are mineralogically more mature (orthoquartzose) but more angular than the Middle- and Outer-Shelf sands. Last of all, McKinney and Friedman note that the distribution of fine sediment more frequent in low areas. They also note an overall increase in fines beyond the shelf break. Fig. I-90 shows the distribution of mean grain size in the area studied by McKinney and Friedman (1970).

Off the eastern end of Long Island and around Block Island, several investigations identifying past drainage systems (Fig. I-91 and I-92) have been conducted (Garrison and McMaster 1966; McMaster and Ashraf 1973a, 1973b; McMaster and others 1968). These studies reveal seven post-Jurassic drainage patterns. Fig. I-92a shows the major features found along this area of the Long Island shelf compartment. Several Late Wisconsin glacial moraines are also shown.

This subregion is quite different from other shelf areas, especially those south of the Long Island shelf. It is characterized by highly irregular topography partially influenced by glacial and bedrock-

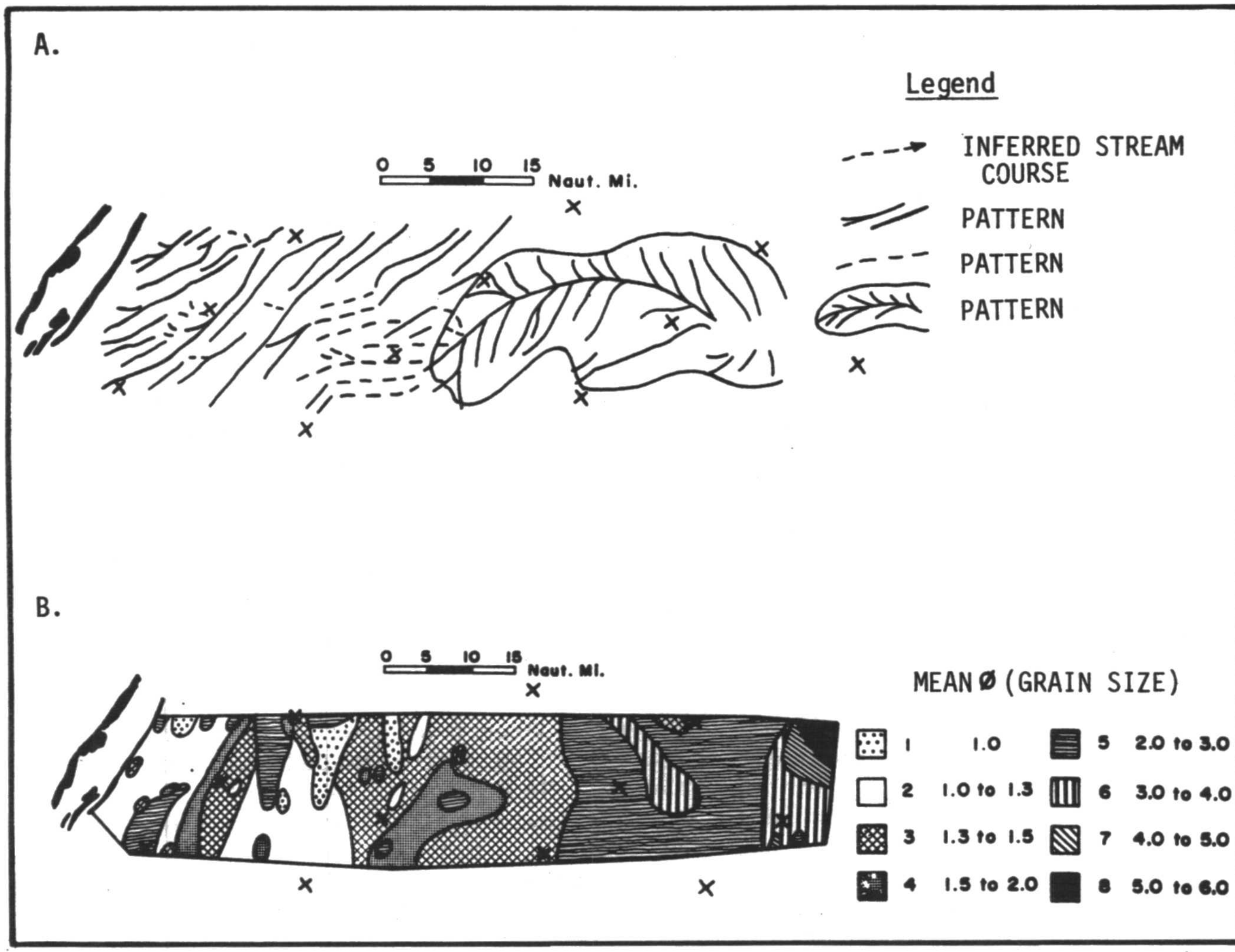


Fig. I-90

Inferred pre-transgressive drainage patterns (A) along a section of the continental shelf off Long Island. Drainage patterns based on present day bathymetry. Mean sediment size (B) illustrates variability of the "surficial sand sheet" along the northeastern Atlantic Continental Shelf. After McKinney and Friedman (1970).

structural elements. Bedrock outcrops and irregular topography have inhibited barrier formation and at the same time have protected some features from transgressive erosion. As a consequence of protection, a spit complex formed off Block Island (McMaster and Garrison 1967). Because of the local topography, barrier-island retreat was a less important process along this portion of the Inner Shelf off Rhode Island.

The dominant ancient features of this region are a complicated series of ancient drainage patterns which have been mapped by McMaster and Ashraf (1973a, b). Of the seven post-Jurassic drainage patterns that have been mapped, only the upper system (Fig. I-91) falls within the Late Wisconsin period. Several generalizations, however, can be made after comparing all the ancient courses over time. In general, the valleys are superimposed on each other, indicating the existence of preferred fluvial pathways. On each of the pathways valleys are "stacked" on each other, suggesting that during various lowstands streams flowed along similar paths. Streams flowing along these paths deposited enough sediment to resist erosion by subsequent marine transgressions. That the coastline has been tectonically stable in this area is indicated by the low valley-height/valley-width ratios, which suggest little down-cutting. In the northern part of the area, however, larger height/width ratios are present as the result of glacial scouring.

In general, local structures created by erosion and glacial deposition have governed stream courses in this area since post-Tertiary times. Streams in Long Island Sound were diverted southeastwards by the "cuesta" in that area. Fig. I-92 shows several of the features found in this region.

Of importance to this study are the most recent fluvial systems formed during the last sea-level lowstand. These valleys form a very complex drainage system which sometimes corresponds with existing depressions (Figs. I-91 and I-92). Concerning this system, some general statements can be made. First, in Rhode Island Sound the valleys trend southward and on its eastern side the trends are southeast. Most of the valleys can be traced seaward from Narragansett Bay, the Sakonnet River, and from valleys in coastal Connecticut. Out on the CS, Block Channel formed the major pathway into which these smaller drainages flowed. A major valley trending eastward from northeastern Long Island across Block Island Sound is possibly an eastern reach of the "Sound River" believed to be Late Pleistocene in age. Other features in this area date from the period during the Late Wisconsin when the ice began to retreat (Fig. I-92). A terminal moraine impounded glacial meltwaters and deposited a non-fossilliferous concretionary clay (Frankel and Thomas 1965; Grim and others 1970). Eventually, the meltwaters breached the moraine between Long Island and Block Island and produced small distributary deltas.

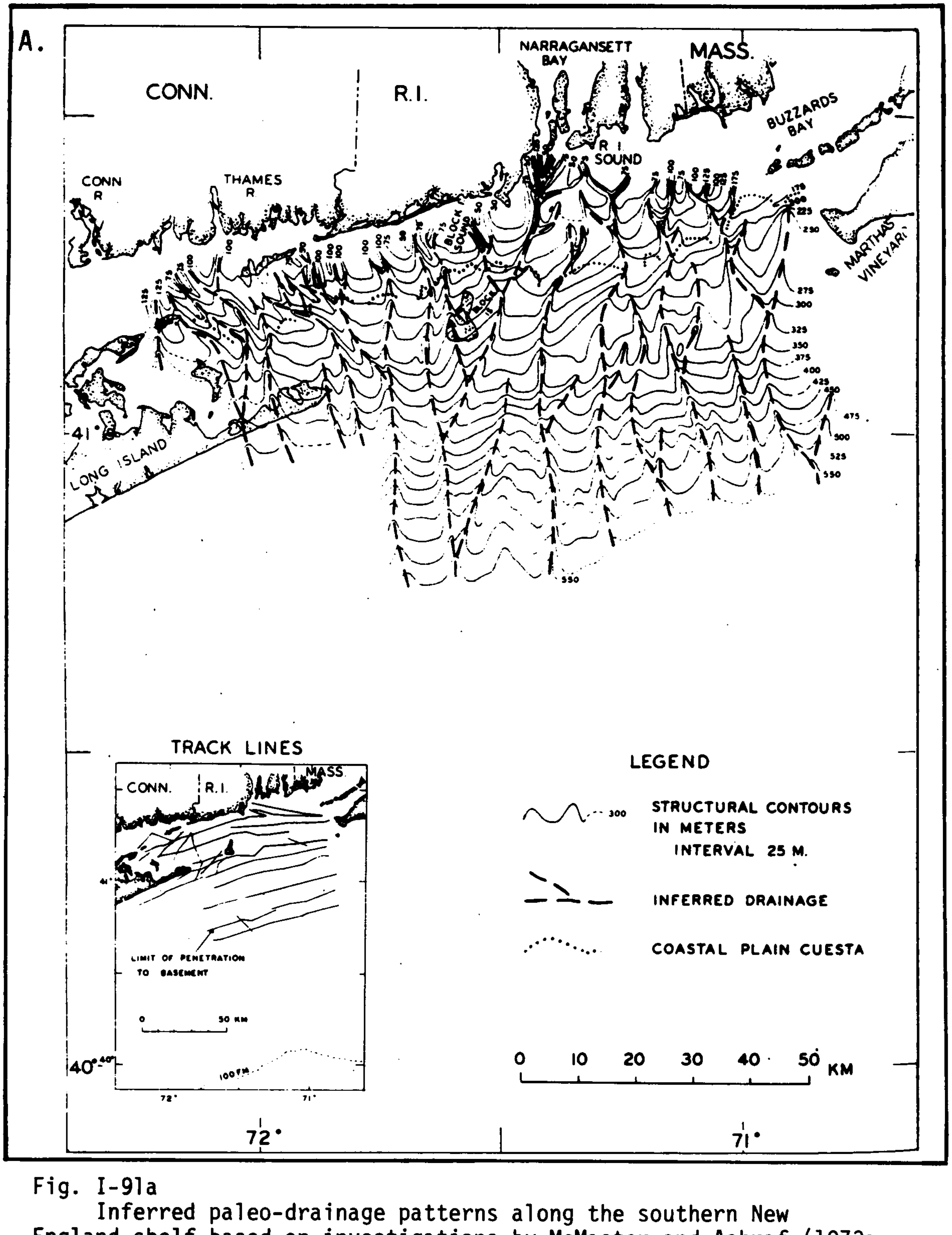


Fig. I-91a

Inferred paleo-drainage patterns along the southern New England shelf based on investigations by McMaster and Ashraf (1973a, b, and c). Although the drainage pattern of the basement surface predates the Pleistocene, it represents the preferred drainage corridors in which more recent fluvial systems (see Fig. I-91b) were confined. After McMaster and Ashraf (1973c).

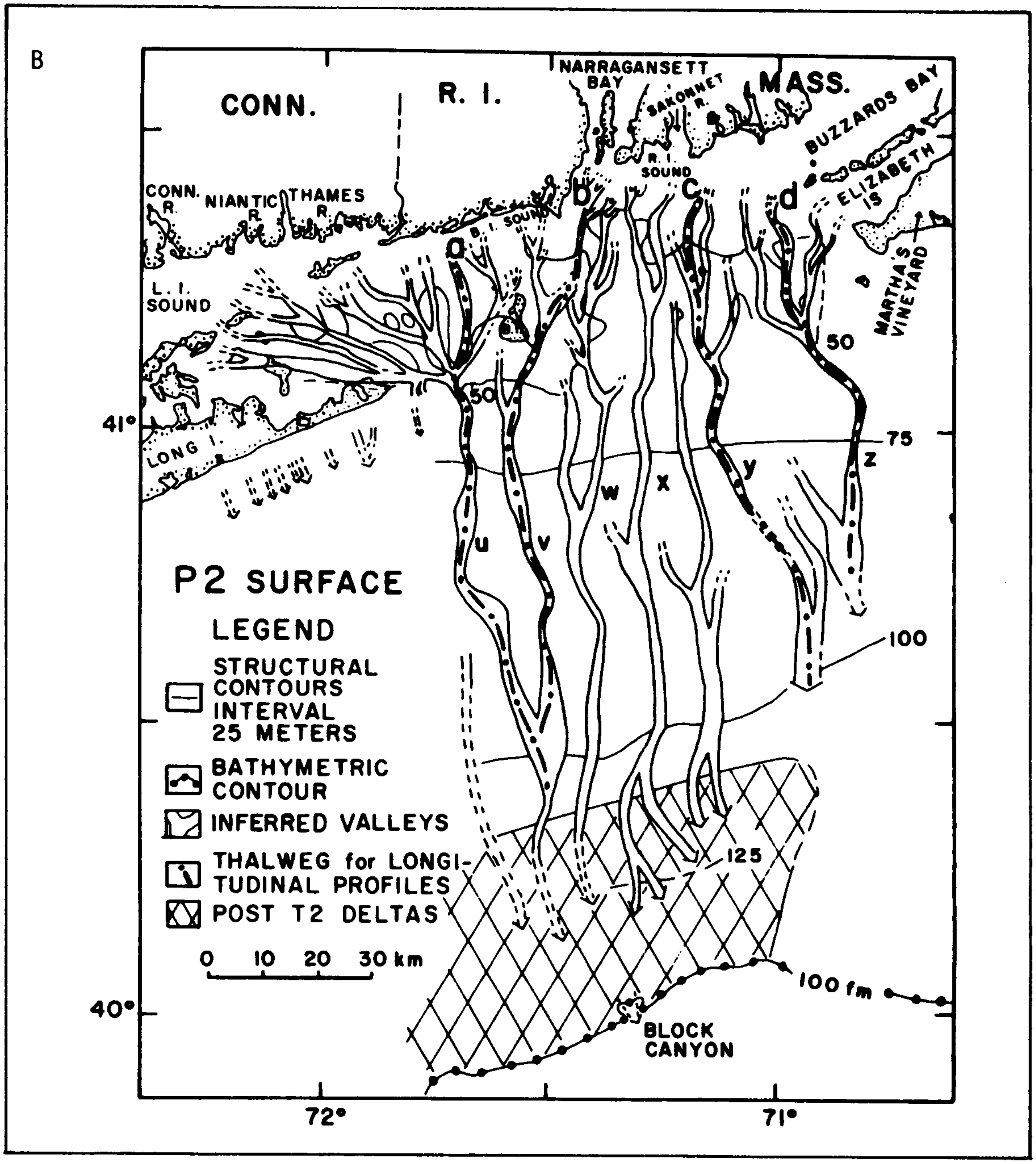


Fig. I-91b

Preferred drainage corridors used by Late Pleistocene rivers flowing across the southern New England shelf. After McMaster and Ashraf 1973a.

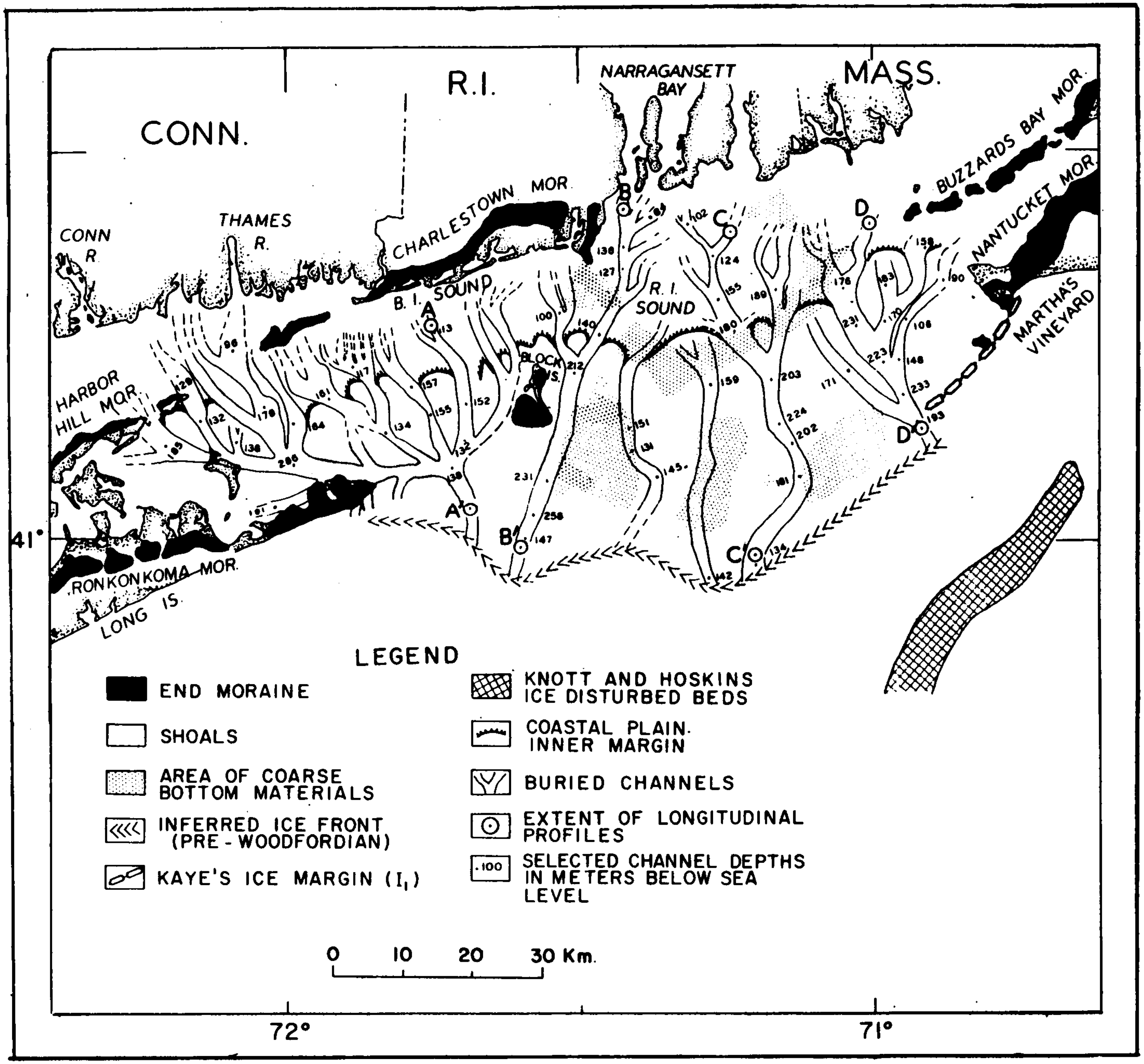


Fig. I-92a
Buried channels and other geologic features along the southern New England shelf. After McMaster and Ashraf (1973b).

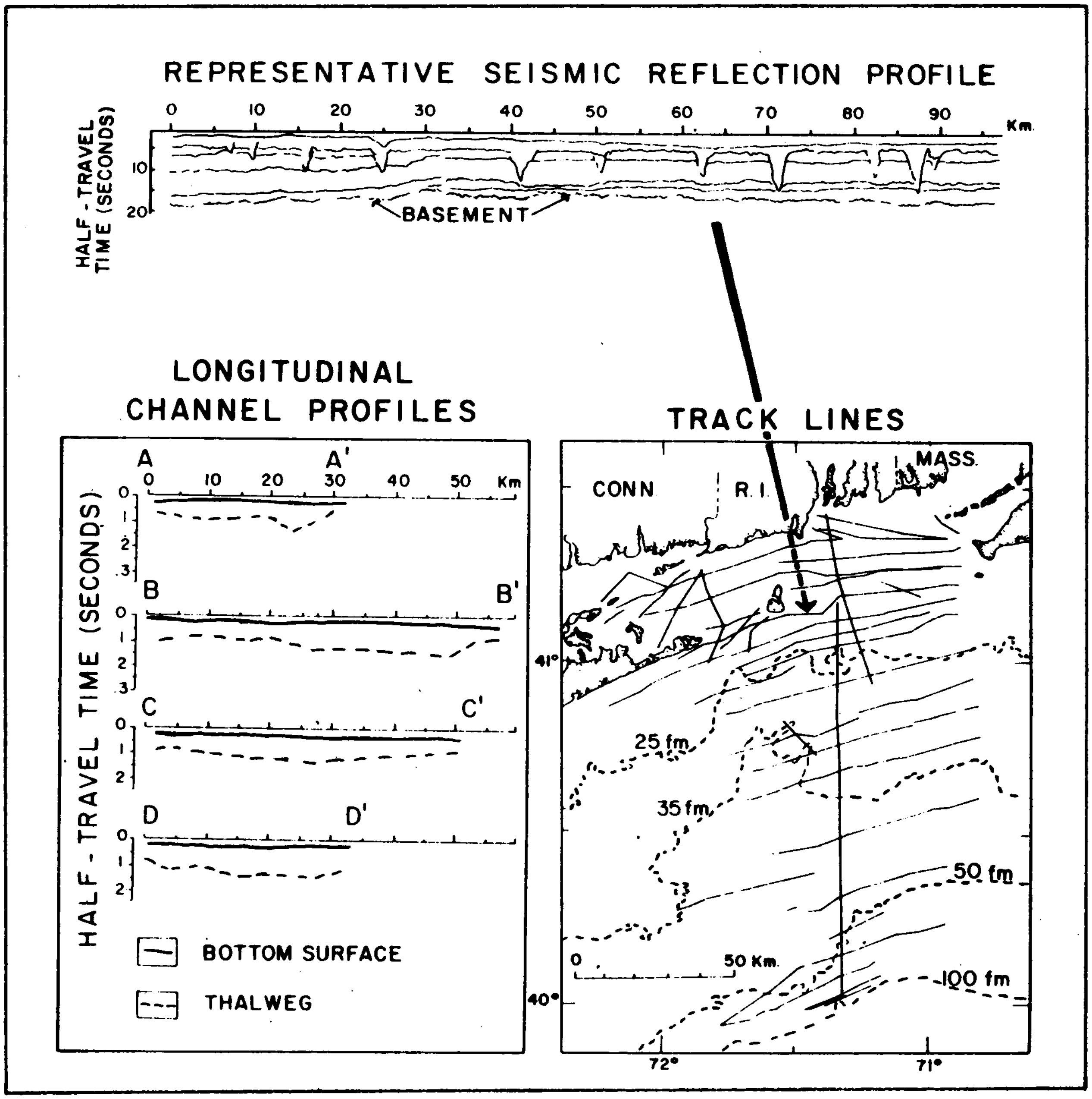


Fig. I-92b

Representative seismic profile from a transect along the Southern New England inner shelf. After McMaster and Ashraf 1973b.

Drainage rearrangement occurred when the "Sound River Channel" was filled with morainal deposits. Drainage continued to flow southward from Narragansett Bay but drainage from southeastern Massachusetts was diverted to the southwest. Seaward of the Late Wisconsin end moraine, the drainage followed the older patterns of Rhode Island Sound, funneling waters into Block Channel, which carried them toward the Block Delta.

Studies by McMaster and others (1968) have pointed out the major difference between the Block Island-Rhode Island Sound and the CS to the southeast. The southern New England shelf, like the Gulf of Maine, has a bedrock surface which consists mostly of metamorphic rocks (gneiss, schist of Paleozoic age) close to the sea floor. This surface acts as a structural control influencing the surficial geology in many cases. Between Long Island and Martha's Vineyard along the inshore boundary of the Upper Cretaceous formations is a well-defined fall line which trends northeast across the region (Fig. I-91). The bedrock surface is very irregular and has a maximum relief of 52 m. Interfluves north of the fall line are covered by thin (6-46 m) glacial deposits laid down during the Pleistocene. Thicker deposits are found in the valleys and may reach up to 100 m. After the last Wisconsin glacial advance and during the Early Holocene, the coastal region probably underwent some subaerial erosion. The subaerial surface in Block Island Sound is buried 18 m below the sea bottom, but in Rhode Island Sound the surface is exposed in some places. The surface is generally rough, exhibiting numerous gullies and small stream channels (McMaster and others 1968:473).

An example of the way in which the bedrock surface affects the surficial geology in this region is seen in the research of McMaster and Garrison (1967). Based upon detailed examination of bathymetric data, they have identified what they believe is an intact relict beach ridge and lagoonal system (Fig. I-93a). The survival of this feature is ascribed to the protective effects of the irregular topography formed by bedrock outcrops. In this case the barrier (that is, spit) was tied to a rocky headland and was unable to migrate landward as sea level rose. It presumably was engulfed intact owing to the protective action of numerous offshore ledges and islands which absorbed and reduced wave energy, thus preventing its erosion. The morphological cross section (Fig. I-93a) of this submerged feature is similar to that of modern spits nearby and supports their interpretation. No sediment samples of intact lagoonal deposits were found, although the "correct" sequence of surficial sediment types was observed from a submarine (McMaster and Garrison 1966). On the basis of local sea-level curves, this feature is estimated at 8300-9000 B.P. Small-scale features, such as ripple marks and scour patches, were observed on the submerged ridge during its investigation and are considered to be modern.

Irregular topography, as discussed above, has allowed portions of the subaerial surface to be preserved in this region. This nearshore environment was very different from those of areas to the south. Unlike them, it consisted of an irregular coastline with many rocky islands and headlands much like those found on the Maine coast today.

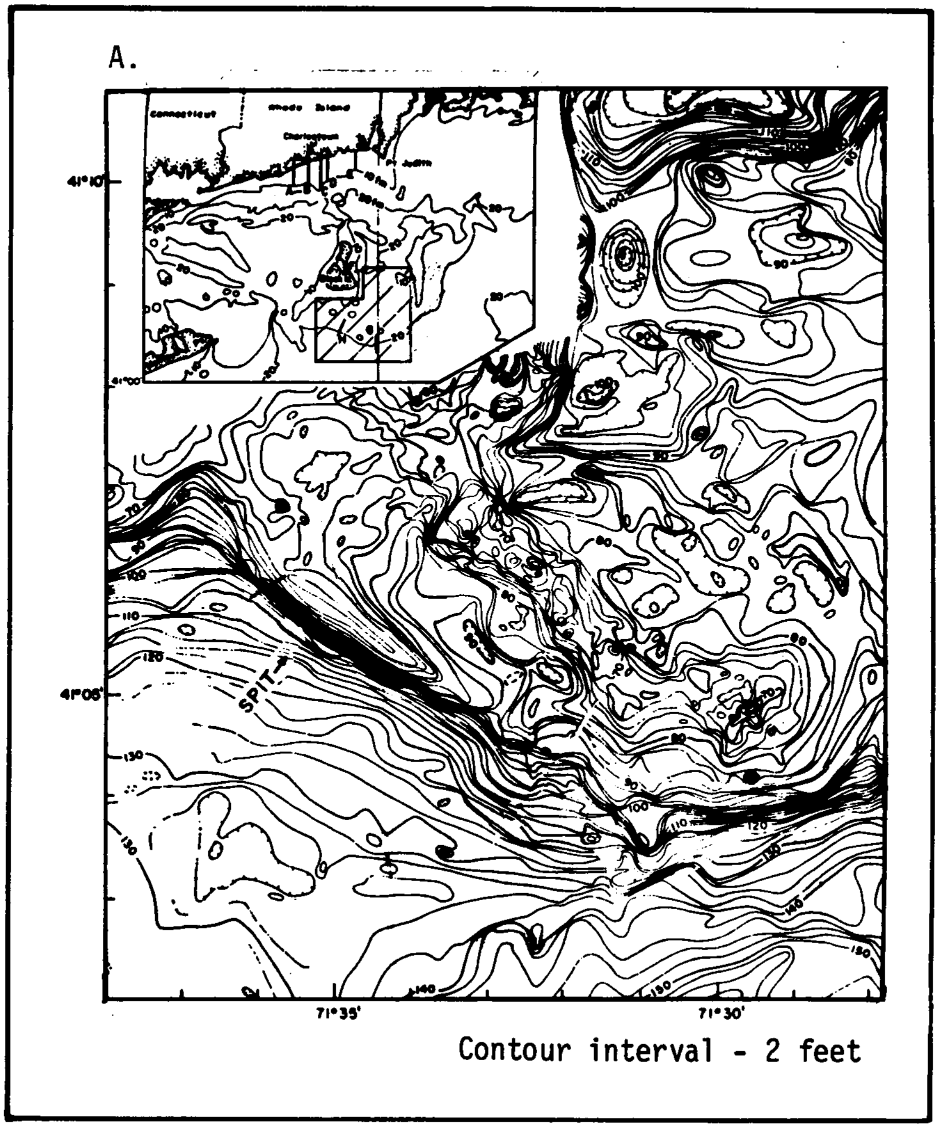


Fig. I-93a

(A) Bathymetry and location of probable submerged and truncated spit near Block Island, Rhode Island. (B) Profiles on following page show relief of present-day barriers. Also shown are profiles of the drowned shoreline near Block Island. Locations of the profiles are shown in the index map. After McMaster and Garrison (1967).

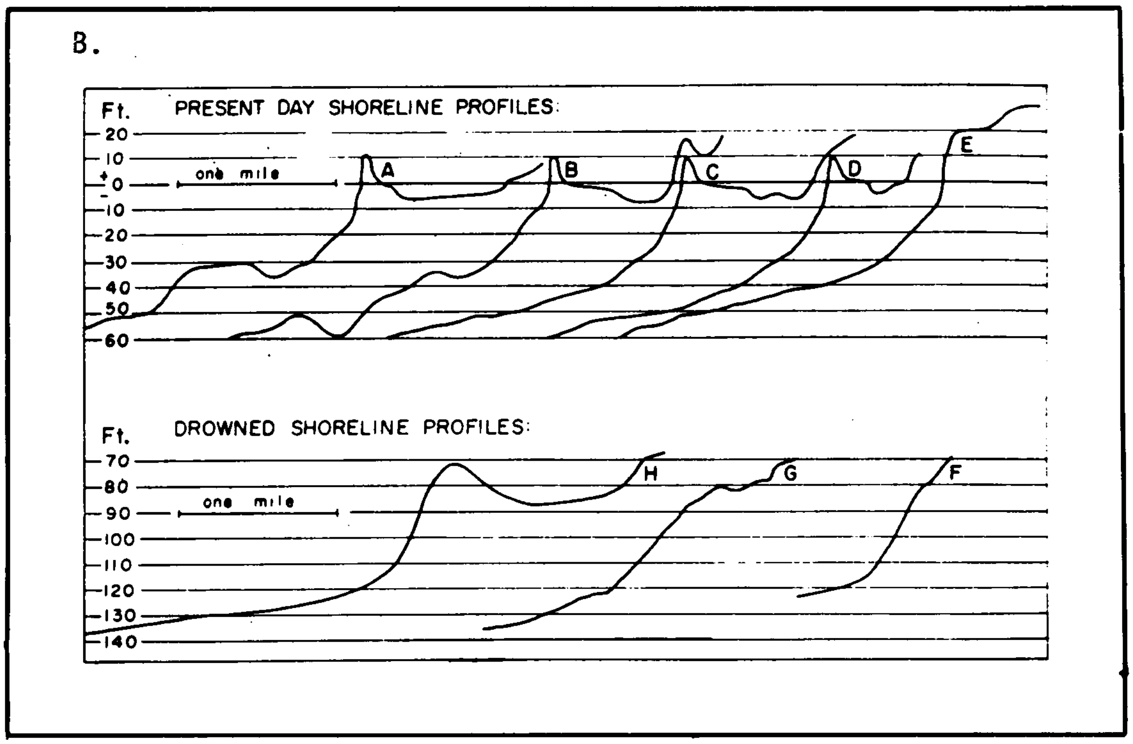


Fig. I-93b

B. Profiles along the present day coast of Rhode Island showing relief of barriers and profiles of drawn shoreline near Block Island. Locations of the profiles are shown in the index map. After McMaster and Garrison (1967).

It is important to realized that although the irregular topography helped to protect subaerial features from erosion during transgression, some erosion was unavoidable. McMaster and Garrison (1967) observed active ripple formations at depths of 34 m off Block Island, meaning that their "submerged" spit is still in range of modern hydraulic processes.

A different type of barrier island has been investigated by Dillon (1970) along the coast of Rhode Island (Fig. I-94). This barrier has been migrating landward by means of storm washover for at least several thousand years. Insufficient sediment supply has limited the size of the barrier and it has eroded as rapidly as its landward side has prograded. Dillon's (1970) research provides information on a mode of barrier-island migration which occurs when sediment supply is insufficient. Sand is supplied to the back of the barrier by washover and wind transport from the barrier front. This process leaves a thin lag deposit which is then reworked, leaving no evidence of the earlier barrier position except where lagoonal deposits have filled major depressions. Landward of the barrier, seismic profiling and coring have revealed glaciation features (kettles, outwash surfaces) and fluvial features (stream channels) (Fig. I-95) beneath the lagoonal muds. In this case, the subaerial surface is too shallow to escape erosion during transgression. As it now stands, only the most deeply buried portions of the subaerial surface remain intact on the seaward side of the barrier.

Several general investigations of this shelf compartment have been done using seismic profiling (Cousins and others 1977: Garrison and McMaster 1966). The most recent survey of this region suggests that Garrison's (1970) estimate of thin Pleistocene shelf deposits (30 M) is incorrect (Cousins and others 1977). Drilling data indicate that the seismic profiles of Cousins and others (1977) show only the Pleistocene-Holocene section. On these profiles the researchers have identified a reflector (reflector II) believed to represent a surface produced by the last marine regression. It is quite probable that this reflector consists of several features including the subaerial surface that is of interest to this project. This surface crops out on the sea floor in places. In the Hudson Shelf Valley, the reflector is covered by 10 to 20 m of sediment of presumed marine origin. In some areas, tidal scour within the early Hudson estuary may have removed or truncated the pre-transgressive subaerial surface. Thus, reflector II probably represents several different features. Its presence does not necessarily mean that an intact subaerial surface exists.

Along the Central and Inner Shelves of Long Island and Rhode Island (Cousins and others 1977) profiles have revealed very complex sedimentary structures and many discontinuous sub-bottom reflectors. These reflectors probably represent scouring, channeling, and cut-and-fill bedding from fluvial systems associated with glacial outwash. Some buried channels as small as 5-7 m deep and 5 km wide have been noted in

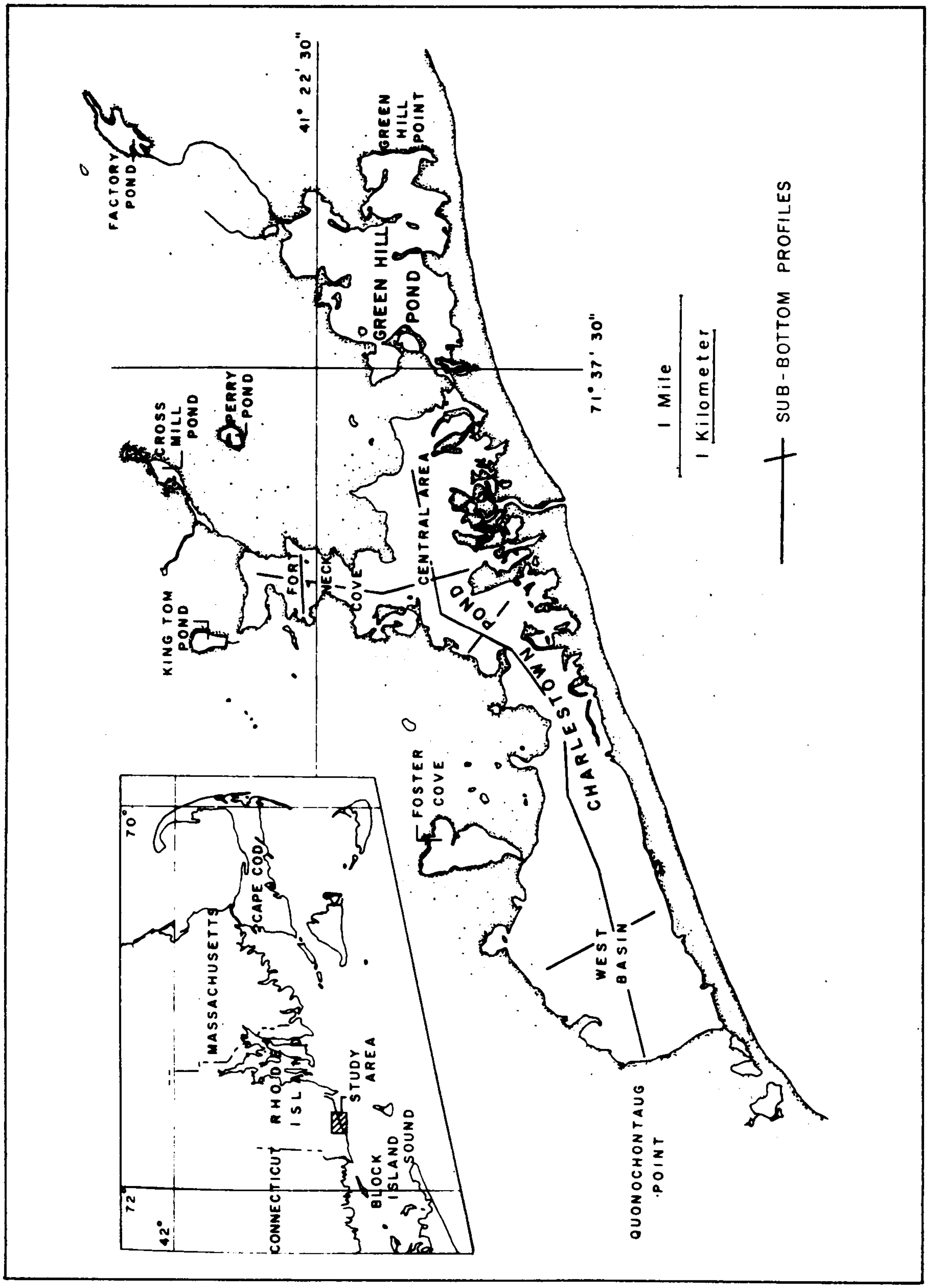


Fig. I-94

Section of the Rhode Island coast investigated by Dillon (1970) showing positions of seismic profiles (after Dillon 1970).

profiles (Cousins and others 1977). One such buried channel lies in the Long Island Shelf Valley.

Other forms on the Long Island Shelf are subtle scarps and terrace-like features (Cousins and others 1977; Ewing and others 1963; Garrison and McMaster 1966; Knott and Hoskins 1968). Reported scarps correspond well with the Nicholls (148 m), Franklin (125 m), and Fortune (85 m) shores (depths given are those near the northern flank of the Hudson Shelf Valley). The scarps are not horizontal but dip towards the north (Dillon and Oldale 1978). Each scarp is formed at the shoreward edge of a major terrace. The Nicholls scarp and its associated terrace are probably older than 30,000 B.P. based on a few radiocarbon dates (Donn and others, 1962; Emery and Uchupi 1972; Garrison and McMaster 1966). Ewing and others (1963) noted that the 146 m "terrace" was covered by about 19 m of sediment in the vicinity of the Hudson Valley. They noted further that the original surface cut into the Hudson Apron.

Also in this area, Knott and Hoskins (1968) located by seismic profiles a series of prograded beds 5 to 50 m below the sea floor and about 50 km southeast of Long Island. They suggested that these structures were deposited by glacial meltwaters at a time when sea level was well below -65 m.

East of Block Island, Knott and Hoskins (1968) found structures probably representing filled channels from glacial meltwater streams. Complex profiles interpreted as representing sediments folded by glacial ice were observed about 30 km south of Marth's Vineyard and are probably older than 20,000 B.P. Needless to say, coring is necessary to identify these features.

The evolution of the Outer Shelf along this compartment has been influenced significantly by local events. Glacial outwash contributed considerable sediment for prograding and upbuilding this portion of the Shelf. On the other hand, the Inner Shelf has been directly sculptured by glacial ice and remolded by glacial deposits. Finally, marine transgression has changed surface features as glacial and fluvial landforms were reworked.

The discussions so far have dealt with the Shelf in the Long Island shelf compartment. Some research which is of interest to archaeological studies has been conducted along the lower Hudson estuary and Long Island Sound, particularly that done by Newman (1977) in the vicinity of northwestern Long Island. Newman's (1977) research emphasizes the benefits of using a multi-disciplinary approach to paleoenvironmental reconstruction. The integration of several disciplines (such as geomorphology, pollen analysis, paleontology) have revealed several contradictions regarding previously published interpretations of the paleoenvironment (Newman 1977). Although Newman (1977) is not able to resolve all the conflicts, he provides a sharp focus on the issues that need addressing. He recognizes the discrepancy between his sea-level

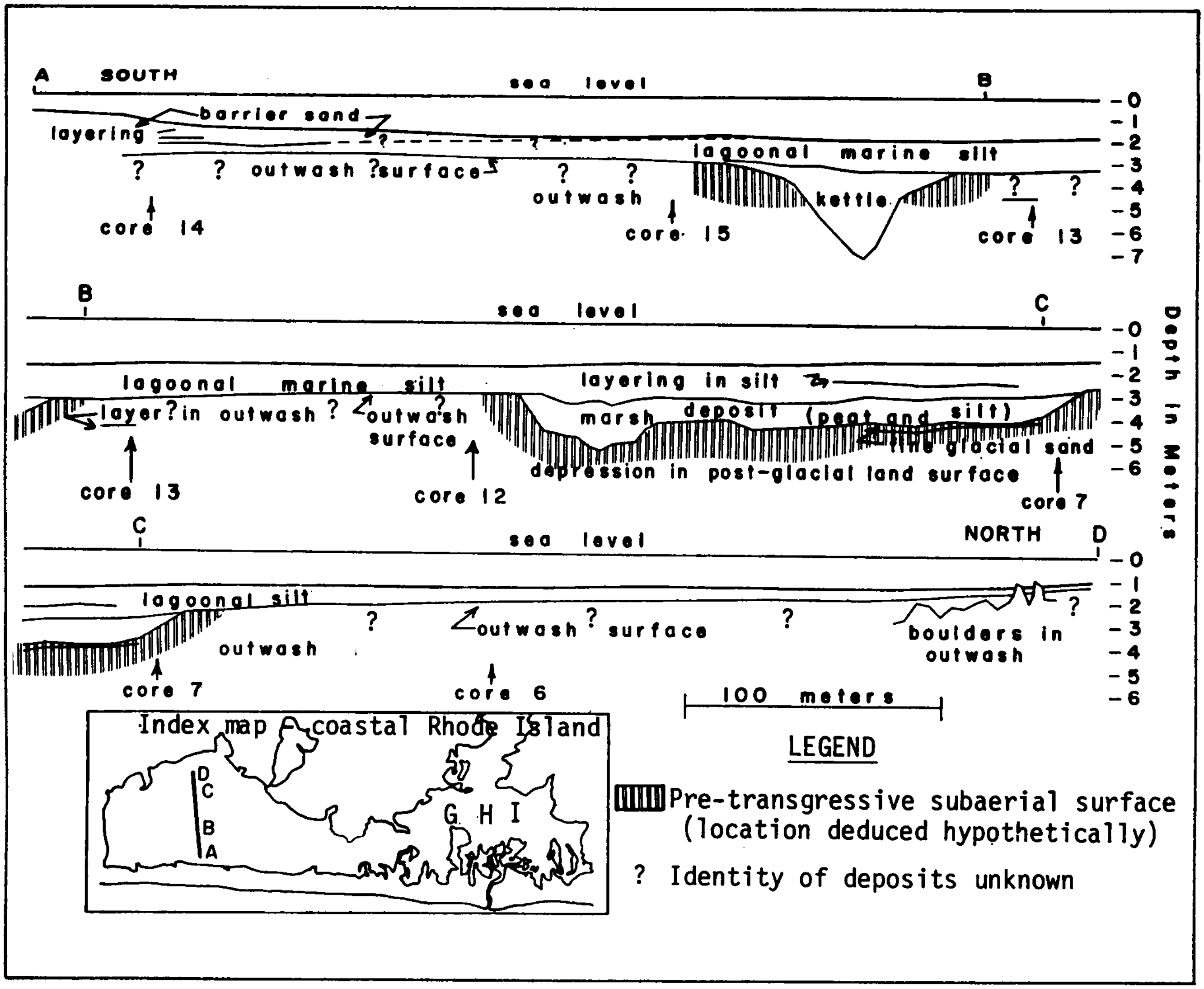


Fig. I-95

Interpretive profiles of a lagoon along coastal Rhode Island. The identification of the pre-transgressive subaerial surface is based on inference. As marine transgression continues in the future and erosional shoreface retreat passes over this area, only the buried subaerial surface in the stream valley between stations G-H may be deep enough to escape erosion and redistribution. Adapted from Dillon (1970).

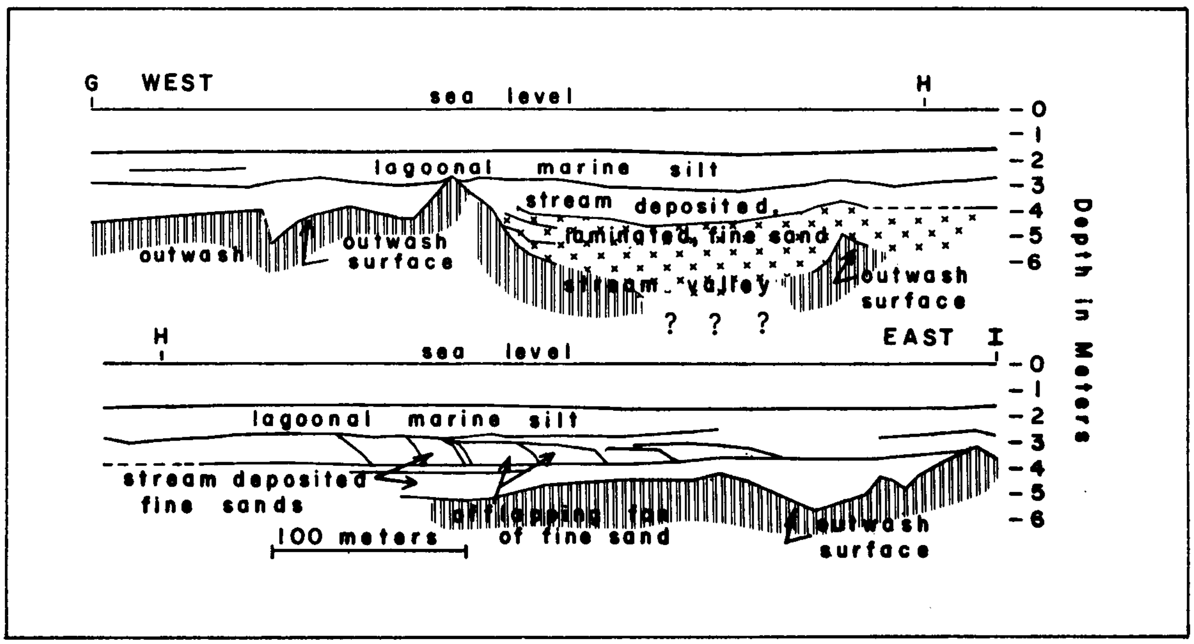


Fig. I-95 (continued)

data for this region (discussed at the end of this subsection) and those given by Milliman and Emery (1968).

The Holocene development of the lower Hudson River has been investigated by Weiss (1974) and Weiss and others (1976). Although most of his research emphasizes paleoecology, he also offers some data on the geomorphology of this estuary. Glacial Lake Hudson occupied this area at the end of the Pleistocene. The lake was formed by glacial meltwaters backed up behind a dam at the "Narrows" (Fig. I-96). The dam was formed by an extension of the Harbor Hill moraine across the Hudson Valley. The lake drained when erosion breached the dam. This occurred well before 12,000 B.P., since tidal conditions were present in the estuary well before that date. Salinity high enough to support foraminifera (approximately 32 parts per thousand) was established within the vicinity of today's lower Hudson River by 11,500 B.P. Weiss (1974), like Newman (1977), also notes that his data are in opposition to Milliman and Emery's (1968) sea-level curve. Glacially depressed topography and other factors seem to be at work making different sea-level curves for separate sections of the New York Bight and inland region.

Last of all, Weiss (1974) also noted that the Hudson estuary has been undergoing additional changes in the last 1,500 to 3,000 years. From his data, he observed a decrease in the salinity of the estuary, possibly representing the effect of continued sediment infilling on the river.

Sediments from Pleistocene lakes have been encountered along Long Island and Block Island Sounds (Athearn 1957; Bokuniewicz and others 1964; Frankel and Thomas 1966; Grim and others 1970; Newman 1977; Oldale and others 1973; Schafer and Hartshorn 1965). Stratigraphic correlation of these sediments over large areas has not been attempted. Speculation about these deposits has offered several possible correlations with varied clays outcropping on coastal areas (Grim and others 1970; Newman and Fairbridge 1960; Schafer and Hartshorn 1965; Tagg and Uchupi 1967).

The Pleistocene geology of Long Island Sound is not well understood. Most seismic profiling done within Long Island Sound can differentiate at least three major reflective surfaces (Grim and others 1970; Tagg and Uchupi 1967). Fig. I-96 gives bathymetry and place names for Long Island Sound.

Although Long Island Sound is at present rather shallow (generally less than 40 m) with low relief, seismic profiling and coring have indicated that its Pleistocene surfaces contain numerous valleys (Grim and others 1970; Tagg and Uchupi 1967). Upson and Spencer (1964) reported bedrock valleys with thalwegs between 24 and 8 m along the northern shore of the Sound. These thalwegs, however, represent features which are much older than the Late Pleistocene.

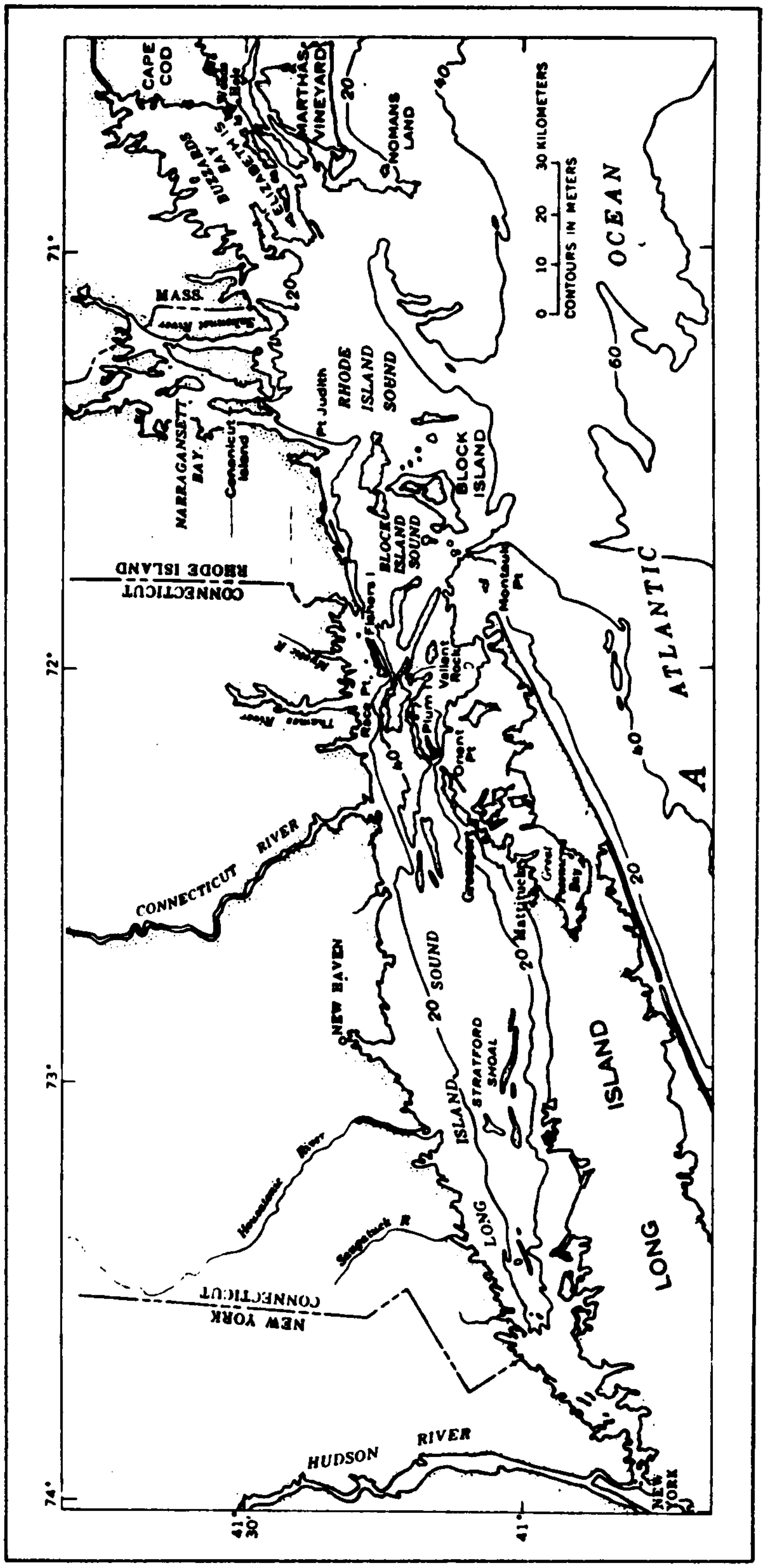


Fig. I-96

Bathymetry and place names for Long Island Sound and vicinity. After Tagg and Uchupi 1967.

Besides identifying bedrock thalweg depths, Upson and Spencer (1964) also identified the depth of thalwegs belonging to pre-estuarine valley conditions. These depths, integrated with local sea-level curves, provide some information on when estuary conditions were established along the major rivers of Long Island Sound. From their data, estuarine conditions were probably established first at the present mouth of the Thames River (-40 m) during the Early Holocene. Grim and others (1970) have followed the ancestral Connecticut River into Long Island Sound and found that its pre-estuarine (pre-transgressive) thalweg was -55 m below the surface 3 miles south of the mouth of the river; 5.5 miles south it was at about -58 m. These data also suggest that other rivers along the north shore of Long Island Sound used to extend out into the Sound. They also indicate the minimum depth to which rivers penetrated before sea-level rise submerged their channels.

The seismic profiling that has been done to date in Long Island Sound has identified only general sub-surface reflectors. For example, the work done by Grim and others (1970) identified three units above the basement surface (Fig. I-97). Tagg and Uchupi (1967) provided even less information of use to archaeological studies since they lumped all inferred Quaternary deposits into a single unit.

In general, the Pleistocene evolution of Long Island Sound and vicinity is complex (Fig. I-97). The Wisconsin is represented by the Manhasset formation, the Ronkonkoma and Harbor Hill moraines, and their associated outwashes. Overlying the Manhasset Formation are terminal moraines. From New York Bay to a point south of Manhasset Bay the moraine forms a single ridge. At about longitude 73° 42' W the ridge divides into two branches. The southern branch, which runs diagonally across Long Island to Montauk Point, is referred to as the Ronkonkoma moraine (Fig. I-78). The northern branch, which runs along the north shore to Orient Point makes up the Harbor Hill moraine (Fig. I-78). Both of these moraines extend across western Long Island Sound, eventually converging with moraine systems of Rhode Island and southwestern Massachusetts (Fig. I-78).

Whether or not these two moraines belong to two different Wisconsin stadials has been debated. Shafer and Hartshorn (1965) believe that the Manhasset Formation is probably pre-classical Wisconsin in age, and that the Montauk till is separate from the moraines. The same types of unsolved stratigraphic problems that pertain on land apply also to Long Island Sound. Within the Sound, the effects of marine transgression have further complicated the Pleistocene stratigraphic record.

The bottom of Long Island Sound consists of a "Holocene" mud-silt layer 0 to 12.5 m thick (Grim and others 1970; McCrone 1966). The former reported encountering three main types of acoustic sediments within this "Holocene" unit. One type of deposit usually found within 4 m of the floor produces poor reflections similar to those of peat or gas-filled organic silt. They refer to this type of acoustic characteristic as designating "soft bottom" deposits. A second type of deposit

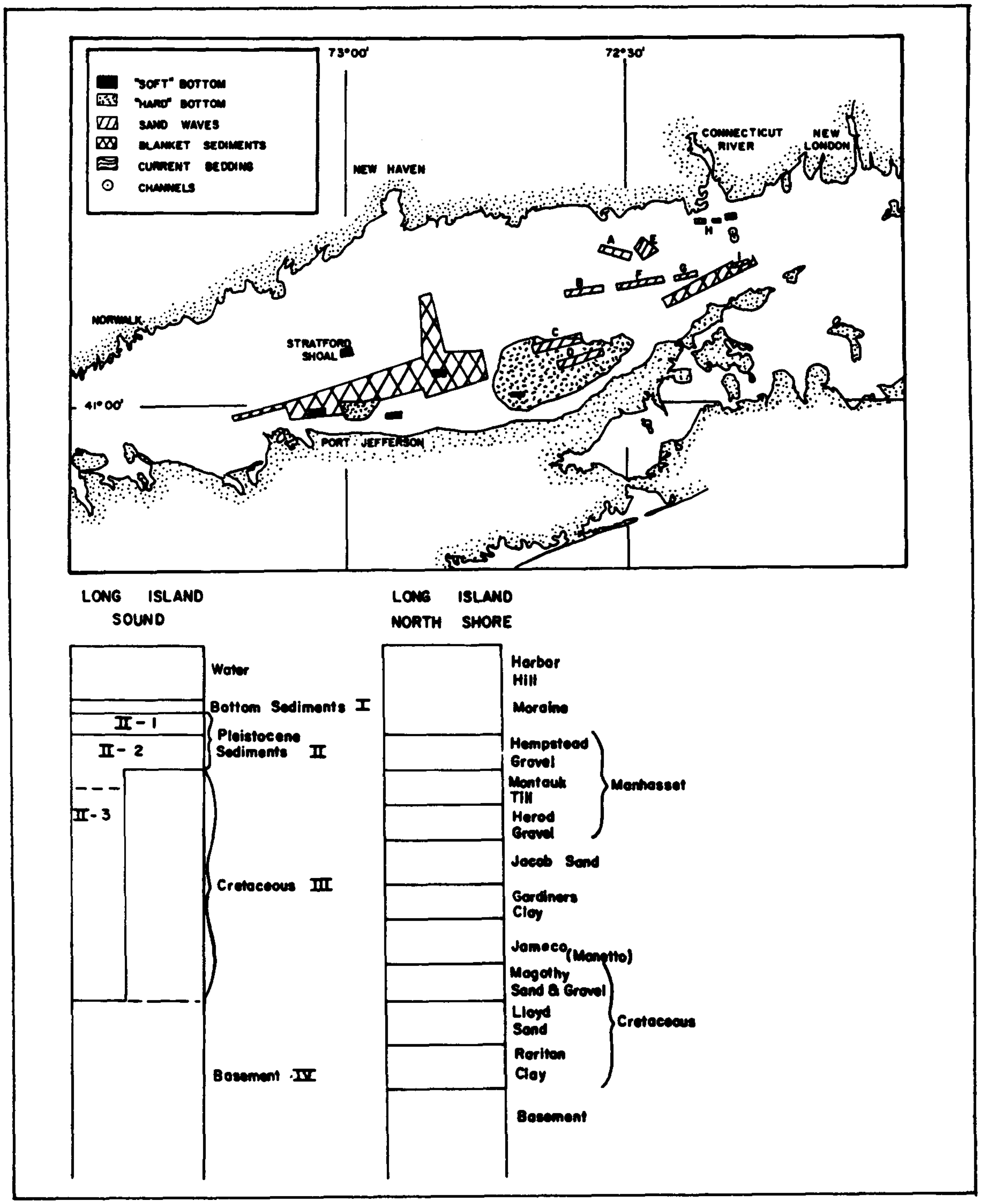


Fig. I-97

Seismic profile data collected from Long Island Sound by Grim and others (1970). "Hard" and "soft" bottom terms reflect the kind of seismic profiles collected in the region. South of the Connecticut River mouth 6 km of a buried river channel have been traced. After Grim and others (1970).

Grim and others identify within the mud-silt horizon produces strong multiple reflectors. They refer to this second type of deposit as "hard bottom" and believe that it represents compacted sand or clay. Fig. I-97 shows the distribution of these deposits within the Sound.

Beneath the "mud-silt layer" (Horizon I), Grim and others (1970) identified what they believe is a Pleistocene unit with at least two subdivisions (Fig. I-97). The upper sub-horizon (Horizon II-1) represents the surface that is of interest to this project. In all probability, it represents a composite horizon containing reworked subaerial deposits as well as fluvial and marine transgressive features. South of the Connecticut River, the three channel-like depressions mentioned previously were observed within this sub-horizon. The pre-estuarine channels described by Upson and Spencer (1964) along coastal Connecticut also may be regarded as extending into this horizon. Until subsurface sampling is done within the Sound, positive identification and correlation of different Pleistocene deposits is not possible. The Late Pleistocene development of Long Island Sound is known in general terms. After deposition of the Harbor Hill moraine, Long Island Sound formed an enclosed basin and received glacio-lacustrine deposits. Continued melting and a minor re-advance filled the basin with both outwash material and water. From Long Island Sound eastward towards Cape Cod, several lakes existed within the basins formed by glacial scouring and dammed by moraines.

In Long Island Sound, silting in these lakes continued until 13,000 to 14,000 B.P. Grim and others (1970) believe that their Horizon II-2 represents these lacustrine deposits. Above this horizon are sediments from the Middletown glacial re-advance which they believe also make up Horizon II-1.

The glacial lakes which extended toward Cape Cod drained during the Late Pleistocene and Early Holocene. The major river system which drained most of the Long Island shelf, as well as Long Island and Rhode Island Sounds and much of the surrounding mainland, was the Block River System of the CS (Fig. I-88). The shelf portion of this valley is easily discernible from bathymetric charts. Several streams can be traced into the Block Valley from its eastern and western flanks (Fig. I-96). Block Valley is 2-4 mi wide along the Inner and Middle Shelves. South of latitude 40° 10', it broadens to over 10 mi in width and ends in a distinctly lobate feature designated the Block Delta (Garrison and McMaster 1966; Knott and Hoskins 1968). Sediment within this valley is a mixture of fluvial and estuarine deposits. Tidal scour associated with the estuary has reworked some of the upper fluvial deposits laid down during marine transgression. Heavy minerals from the Block Shelf Valley indicate this type of admixture (Garrison and McMaster 1966). Beneath the estuarine and reworked fluvial sediments are probably sections of intact flood plain and tidal marsh deposits. A 40 cm core in the Block Delta contained finely laminated medium sand considered by Garrison and McMaster (1966) to be fluvial in origin. On the basis of this core, they suggest that the depth of estuarine and post-transgressive reworked deposits is extremely small.

Some information is available for past sea-level positions along the Long Island shelf from the research done by Bloom and Stuiver (1963), Dillon and Oldale (1978), and Newman (1977). Table I-9 lists the approximate position of sea level at intervals since the Late Pleistocene. Glacial depression and forebulge uplift make sea-level relationships extremely complex and very difficult if not impossible to reconstruct accurately at this time.

Table I-9: Approximate sea-level positions at 3,000-year intervals for the Long Island Shelf.

	Range	Best Estimate	Source
3,000 B.P.	3-4	4	1, 2, 3
6,000 B.P.	9-13	12	1, 2, 3
9,000 B.P.	19-22	21	1, 4
12,000 B.P.	?	?	1, 4
15,000 B.P.	?	?	4
18,000 B.P.	?	?	4

Sources

1. Newman (1977)
2. Bloom and Stuiver (1963)
3. Redfield and Rubin (1962)
4. Dillon and Oldale (1978)

11.0 SOUTHERN NEW ENGLAND SHELF

The region between the Block Shelf Valley and the inner portion of Cape Cod has been covered for the most part in the previous two sections. The work on paleo-drainage patterns (McMaster and Ashraf 1973a, 1973b, 1973c) was discussed in detail on pages 165 to 167. Glaciation in this area was reviewed in the appropriate section. The major region not treated so far in this paper is the Nantucket Shoals, which will be focused on in the following discussion.

The Nantucket Shoals represent the reworked surface of a Late Pleistocene-Early Holocene peninsula. The area is characterized by shoals and sand waves resulting from strong tidal and storm-generated flows (Fig. I-98). Shifting shoals are sometimes deposited within 1 m of the surface, making the area a severe navigational hazard.

Most of the shoals consist of reworked Pleistocene sediment. Some Late Pleistocene channels have probably been buried by these sediments as the shoals have expanded toward the southeast. Emery and Uchupi (1972) show several Late Tertiary and Early quaternary channels buried in the area of Martha's Vineyard and Nantucket Island. Some Late Pleistocene channels may correlate with some of these older drainage systems (McMaster and Ashraf 1973a, 1973b).

Sediment along the eastern edge of Nantucket Shoals consists of medium to fine sands with traces of silt and gravel (Zeigler and others 1964). Beneath these sands is a silt bed with the upper meter or so showing considerable reworking. A shell (Crepidula fornicata) was retrieved from this reworked silt and gave a radiocarbon date of 11,565 ± 400 B.P. There is no evidence of a buried soil horizon or lagoonal deposits in the analysis of a core retrieved from this area (Groot and Groot 1964). The silt seems to be of marine rather than lagoonal origin.

Reworking of the subaerial surface in this area has removed many of the finer sediments, leaving behind sands and gravels (Schlee 1973). An area of fine sediment found along the Middle and Outer Shelves off Martha's Vineyard probably received the sediments carried from the Nantucket Shoals during the Middle Holocene. Removal of these fines has left behind a distinctive heavy-mineral assemblage and older fossils as a result of extensive reworking (Stanley and others 1967). Although some sand has been transported into this area from Cape Cod, a significant amount has been derived from reworking older surfaces. Consequently, only the deeper sections of Early Holocene valleys have escaped the destruction that resulted from transgression. Sand-ridge clusters shown in Fig. I-98 probably indicate those areas where erosion has removed the subaerial surface. At present, our knowledge of shallow structure is limited by poor stratigraphic control.

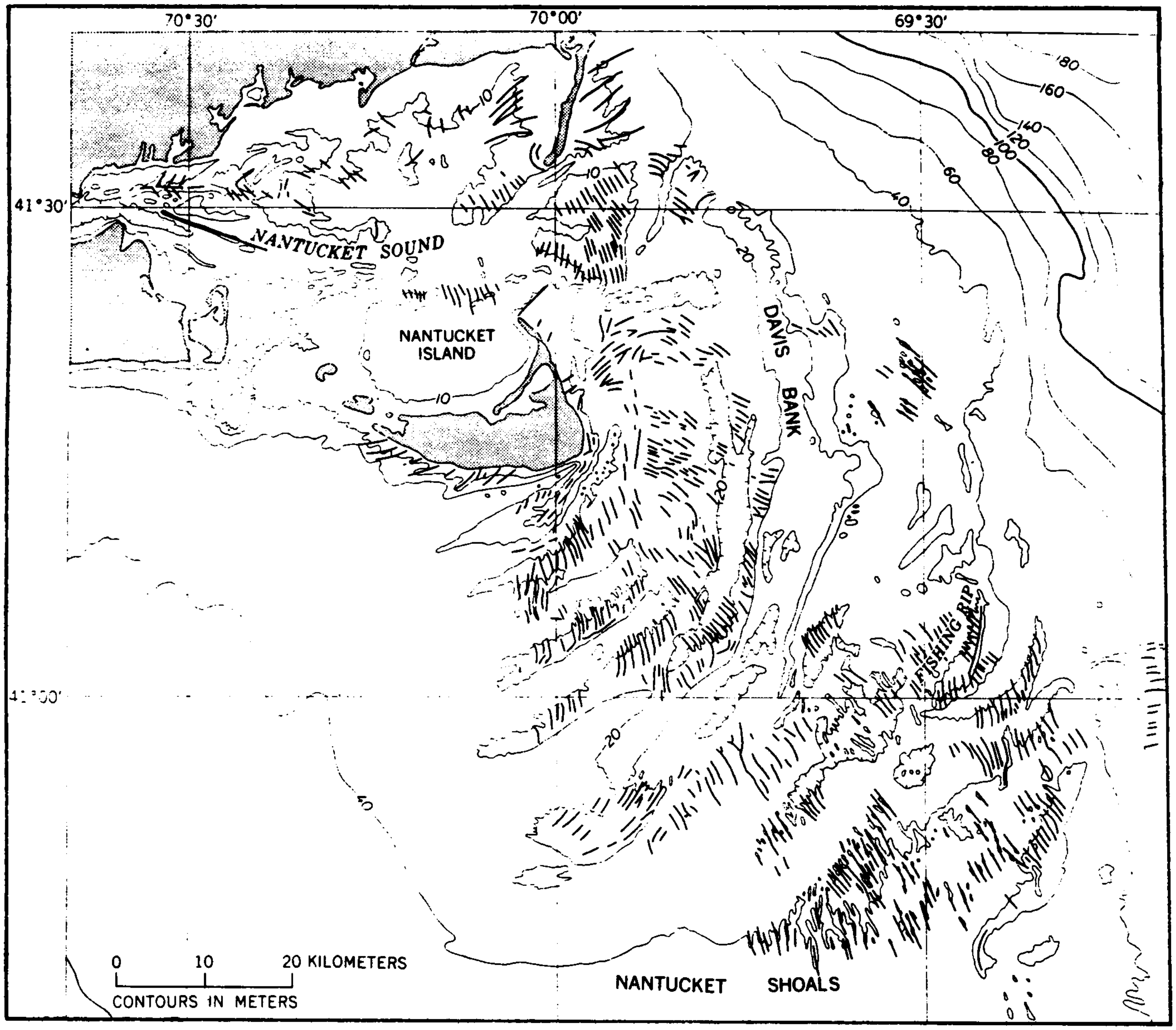


Fig. I-98

Sand shoals and sand waves in the vicinity of Nantucket Island, Massachusetts. These features have been formed by submarine hydraulic processes and not as a result of subaerial erosion. Nineteenth-century soundings indicate that many shoals have migrated and do not represent fixed bottom features. Heavy lines crossing bathymetry indicate the crest of sand waves. After Uchupi (1968).

Southwest of Nantucket Island is a belt of disturbed sediment (Fig. I-92). The exact age of these sediments is unknown, but it is believed that they were distributed by Wisconsin glaciation (Knott and Hoskins 1968).

12.0 GEORGES BANK

Georges Bank may be regarded as a submerged "peninsula-like" body separating the Gulf of Maine from the open Atlantic Ocean (Fig. I-13). When it was exposed to subaerial processes during the Late Wisconsin, the Great South Channel separated this "peninsula" from the mainland CS of southern New England. During the Late Wisconsin maximum, a lobe of the glacier extended partly down this channel (Fig. I-86). In many ways, Georges Bank, during its brief period of subaerial exposure, was probably very similar to the portion of the CS off Nantucket and Martha's Vineyard.

Georges Bank became exposed as sea level dropped during the Wisconsin glaciation. The approximate position of the shoreline along the southern border of Georges Bank is hard to delineate. As the Late Wisconsin glacial maximum was reached, Georges Bank was affected by glacial ice-loading as well as forebulge "uplift" (Edwards and Merrill 1977). Several estimated values for glacially depressed topography about 250 km northwest of Georges Bank range from 50 to 150 m (Kaye and Barghoorn 1964; Tucholke and Hollister 1973). Estimates for forebulge values are harder to make for this area. Edwards and Merrill (1977), however, suggest isolines for the glacial forebulge along the ancestral New England coast. They also suggest that the maximum amount of forebulge uplift was between 40 to 60 m. The tilt of scarps east of New England as well as the "concave-like" profile of the Hudson Shelf Valley support the concept of a glacially induced forebulge (Dillon and Oldale 1978; Edwards and Merrill 1977). The latter believe that by 12,500 B.P. land-surface relationships were essentially restored to today's situation along the southern New England region. In the vicinity of Georges Bank, this would mean that some releveling would have occurred soon after deglaciation. The northwestern edge of the Bank would have undergone some rebound while the southeastern side experienced subsidence of the forebulge.

Some high-resolution seismic profiling has been conducted along Georges Bank in order to determine its Late Quaternary structure (Knott and Hoskins 1968; Uchupi 1970). At least five erosional surfaces were recognized from profiles of the upper 70 m of the Bank. Each surface onlaps the one immediately beneath it. The surfaces pinch out toward the south. The northern flank of the Bank exhibits disturbed sedimentary structures, probably as a result of overriding glacial ice.

Knott and Hoskins (1968) believe that 70 m of sediment were deposited during the Pleistocene. Most of the sediment deposited during the Late Wisconsin has been subsequently reworked into sand ridges, sand waves, and shoals (Emery and Uchupi 1973; Uchupi 1968). Fig. I-99 illustrates the surface morphology found on the Bank today. The shallower portions (that is, 40 m) of the Bank are still being severely reworked by tidal and storm-generated currents (Uchupi 1968, 1970). Sand waves are reported to move up to 12 m annually, and movement of up to 600 m has been documented in historic records (Emery and Uchupi 1972). Georges Shoal (a north central portion of the Bank) is covered by 4 m of water today but is believed to be the site of a legendary ball game held on exposed sand in 1796 (Emery and Uchupi 1972).

The southern half of the Bank is a smooth plain, similar to the surface of the East Coast shelf to the west. To the north is an area of rough topography essentially enclosed within the -60 m contour. The area consists of a series of parallel northwest-trending shoals and troughs. This is a region of strong currents, some averaging more than 4 km per hour (Uchupi 1968). The shoals appear to be migrating westward, burying earlier structures beneath their sands. Gravels are found within some depressions in the area and represent lag material.

The chance of preservation of a subaerial surface that was exposed on the Bank between 15,000 and about 10,000 B.P. is low, given the amount of reworking of the surface of the Bank. In areas shallower than 40 m, any intact subaerial surfaces would have to exist along truncated portions of deeply incised Late Wisconsin stream valleys. Upland soil horizons between interfluves in all probability have been severely reworked since submergence. Those surfaces along the highest region of Georges Bank, while being exposed for the longest period, have been within range of submarine erosion and reworking for nearly 10,000 years.

The Early Holocene landscape along the exposed portion of Georges Bank would have consisted of major river channels with numerous coalescing systems of smaller stream channels. Knott and Hoskins (1968) observed numerous well developed channels within the upper 70 m of sediment covering the Bank. Many of these belong to earlier glacial periods, and some probably represent the truncated basal portions of Late Wisconsin drainages. After the glacier had retreated towards the inner coast of New England, glacial runoff ceased, substantially reducing the flow of rivers and streams along Georges Bank. Surface runoff and the draining of small ponds contributed to these fluvial systems' leaving larger incised valleys to serve smaller streams. As sea level rose, these valleys formed estuary retreat paths before complete submergence of the Bank occurred. Fig. I-99 suggests, on the basis of bathymetry and sand-ridge patterns, where some major river systems may have existed. Some Late Pleistocene and Early Holocene organic materials (peat, shell, large mammal remains) have been recovered from the Bank (Emery and others 1967; Uchupi 1964, 1968; Whitmore and others 1967). Whether any of this material still exists in undisturbed contexts remains to be determined. Emery and others (1965, 1967) mention that there is some evidence of a large area of intact peat along the northeastern side of the Bank.

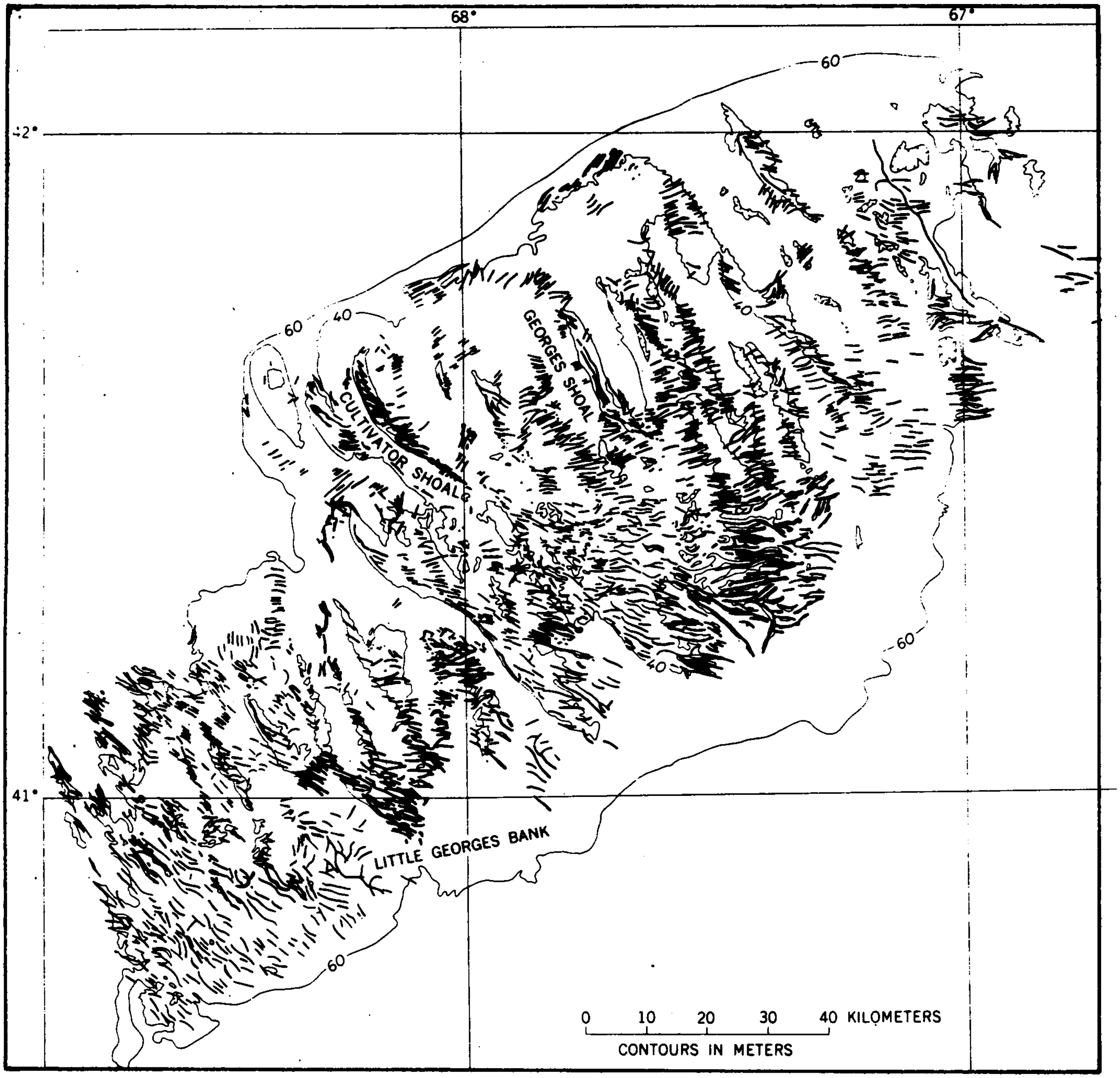


Fig. I-99

Sand waves and shoals on Georges Bank. Heavy lines indicate the crests of sand waves. This type of bottom configuration indicates that significant submarine erosion and redistribution has taken place during the Holocene and after the area was transgressed. After Uchupi (1968).

13.0 GULF OF MAINE

The shallow structure and morphology of the Gulf of Maine are characteristic of a glaciated Continental Shelf. The Gulf enclosed 21 basins with depths ranging up to 311 m. Its bathymetry is highly irregular and includes many bedrock outcrops, moderately deep basins, flat-topped banks and ledges, and sills. For this project, the Gulf has been divided into two subregions. The first is designated as the "Mainland Shelf" (Fig. I-100). It is characterized by glacially scoured valleys, irregular and hummocky bathymetry, scattered "pools" of sediment, and generally thin Quaternary deposits. Some narrow basins, banks, ledges, and bedrock pinnacles are present. Outcrops of igneous rock and boulder beds make up portions of the bottom. The Mainland Shelf is the most important region of the Gulf of Maine, given the objectives of this project. As a consequence of the interaction between deglaciation and sea level, this nearshore region is the only area to have been subaerially exposed (Bloom 1963; Grant 1970; Kaye and Barghoorn 1964; Oldale and others 1973; Schnitker 1974; Stuiver and Borns 1975; Tucholke and Hollister 1973).

Seaward of the "mainland Shelf" is an area that will be called the Central Gulf of Maine (Fig. I-100) in this report. This area consists of numerous irregular sediment-filled basins. Deposits of glacial drift, as well as banks of sand and gravel are found throughout this region. To date, there has been no evidence collected which suggests that the Central Gulf of Maine was subaerially exposed during the Late Pleistocene or Holocene (Oldale and others 1973; Pratt and Schlee 1969; Schlee 1973; Tucholke and Hollister 1973). Glaciation covered this area during the Late Pleistocene. Deglaciation (around 15,000 to 13,000 B.P.) was coincident with marine inundation throughout all of the Gulf and much of the coastal region. As a consequence, access was denied to early human groups (Bloom 1963; Borns 1973; Kaye and Barghoorn 1964; Schnitker 1974; Tucholke and Hollister 1973). Soon after 12,000 B.P. glacial rebound surpassed sea-level rise and present-day land relationships were re-established. In other words, rebound had returned depressed areas essentially to their pre-glacial elevations between 12,000 and 10,000 B.P.

Isostatic and eustatic relationships were such that the western shoreline of the Gulf of Maine transgressed across low areas of coastal New England from Boston northward. This transgression took place between 14,000 and 12,000 B.P., later as one moves northward through New England (Bloom 1963; Kaye and Barghoorn 1964; Schnitker 1974; Stuiver and Borns 1975). In the Boston area, marine deposits are found at least 20 m above sea level (Kaye and Barghoorn 1964) while in central Maine marine sediments were deposited up to 91 m (Bloom 1963) above sea level. Fig. I-79 shows the inland penetration of marine transgression in New England during Late Wisconsin deglaciation.

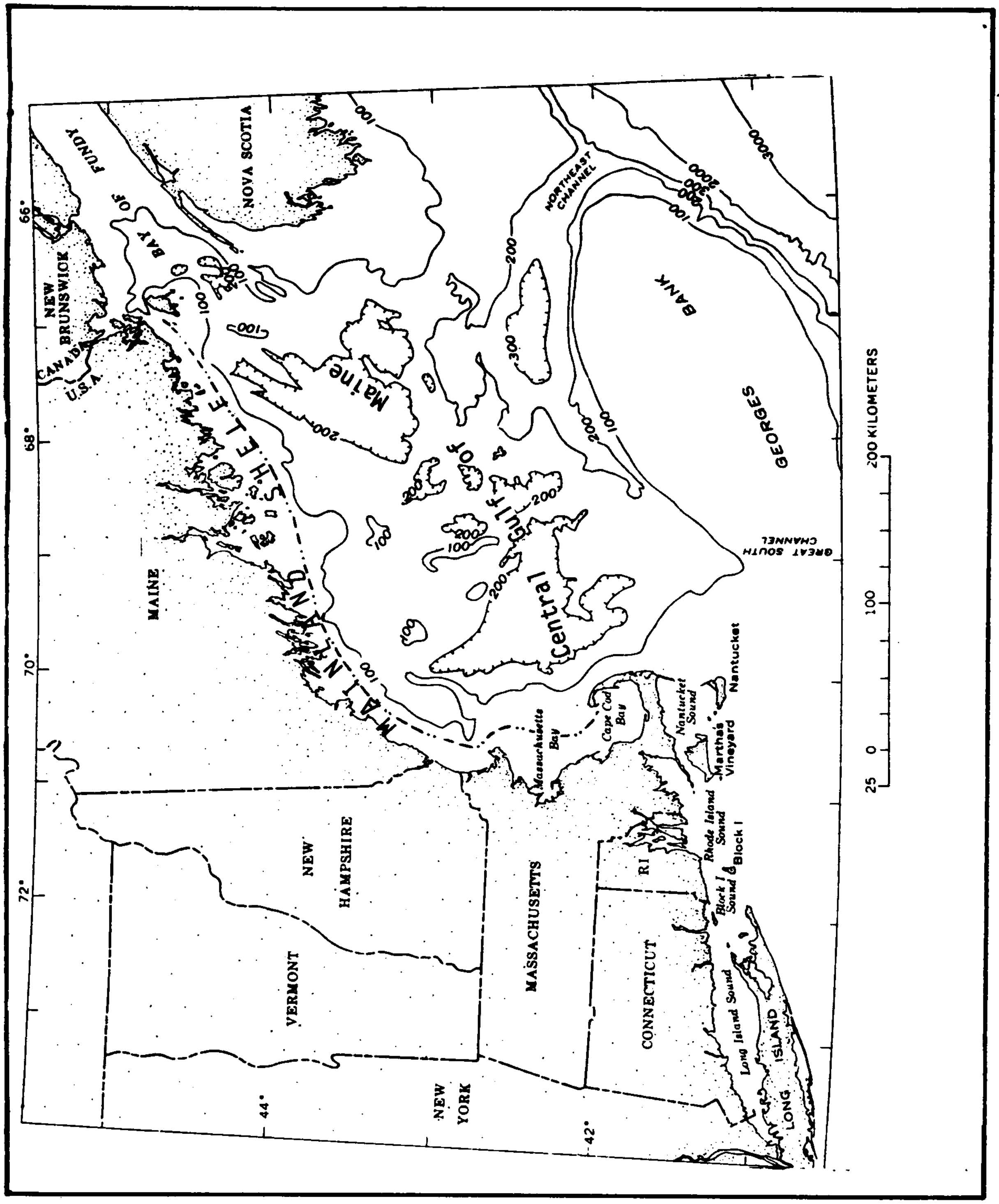


Fig. I-100

Bathymetry of the Gulf of Maine and major sub-divisions discussed in the text. The mainland shelf and Georges Bank are the major areas available for use to prehistoric people during the Early Holocene after the retreat of the Late Wisconsin ice sheet. Adapted from Oldale and Uchupi (1970).

Glacial rebound soon by-passed sea-level rise and transgression was brought to a halt. Between 13,000 and 12,000 B.P., emergence of the coastal region took place. This localized marine regression exposed from 20 to 64 m of the Mainland Shelf portion of the Gulf of Maine (Kaye and Barghoorn 1964; Schnitker 1974). A reconstruction of the shoreline from about 9000 B.P. to the present is given in Chart I-1a. Bloom's early research (1963) which suggests a much longer period of submergence is not substantiated by more recent investigations (see for example Amos 1978; Grant 1970; Schnitker 1974; Stuiver and Borns 1975). In response to the data collected so far, the following discussion focuses on the nearshore region of the Gulf of Maine "Mainland Shelf."

The "Mainland Shelf" may be further divided into the following three physiographic subregions. The southernmost subregion consists of Cape Cod Bay and Outer Cape Cod. The coast along this region is a submerged strand-plain shoreline of wave-reworked stratified and unstratified Pleistocene drift. North of this subregion, the coast is characterized by arcuate bays containing barrier islands and spits. This subregion extends from about Marshfield, Massachusetts, to Portland, Maine. From Portland northward, the coast is characterized by indented embayments with numerous islands. Each of these subdivisions of the Mainland Shelf (Gulf of Maine) are particularly useful since they also typify the types of coastlines present during most of the Middle and Late Holocene.

Cape Cod Bay was initially a glacial lake upon deglaciation (Oldale and others 1973). Once deglaciation extended north of Stellwagen Basin (before 14,000 B.P.) the proglacial lake drained and was later replaced by marine water. Data from the Boston area indicate that sea level dropped only to a maximum low of about 20 to 25 m soon after rebound brought the area back to equilibrium (Kaye and Barghoorn 1964). This exposed a narrow strip of the Mainland Shelf and Cape Cod Bay. Oldale and others (1973) review extensive seismic profiles from Cape Cod Bay and northward along the Mainland Shelf to Portland, Maine. They do not specifically identify any evidence of a buried subaerial surface within the presumably exposed portion of the Shelf.

There is the possibility that a greater section of the Mainland Shelf was exposed along the western Gulf of Maine. The work done by Bloom (1963) and Kaye and Barghoorn (1964) relies heavily upon eustatic curves for inferring sea-level relationships. Only the work done by Schnitker (1974) in central Maine and Tucholke and Hollister (1973) near Stellwagen Basin uses direct evidence. The work by Schnitker (1974) in particular, indicates that there is good evidence that subaerially eroded sediments are encountered to depths of about 65 m off central Maine. Until more sea-level data are collected, we can only infer approximate shoreline positions from the evidence at hand.

The Holocene-Late Pleistocene record north of Marshfield, Massachusetts becomes extremely complex. Oldale and others (1973) have mapped surface sediments and offered interpretive cross sections for the southwestern Gulf of Maine. Fig. I-101 has been taken from their work and offers a simplified version of the area's shallow structure. In reality, relationships between upper Pleistocene and Holocene deposits are complex. Some of the near-shore moraine deposits represent subaerial facies corresponding to glaciomarine deposits (Oldale and others 1977). Marine deposits of Holocene age grade laterally into reworked glacial drift on topographic highs (Oldale and others 1973). In other areas, especially northward of latitude 43°, more relief is present and many nearshore topographical highs have little or no unconsolidated sediment covering them. Fig. I-101 provides some information useful for estimating the probabilities that the preservation of the subaerial surface has been preserved. Areas shown on these figures as having no upper Pleistocene-Holocene deposits (that is, those defined by the 0 isopach) would also be areas where shoreface erosion and post-transgressive current reworking would have been most severe. Many of these areas also represent submerged headlands.

Between the sediment-poor areas delineated by Oldale and others (1973) are regions which acted as local sediment sinks. They are shown as containing marine deposits on Fig. I-101. Tracing low areas filled with Holocene marine deposits shoreward frequently brings one to the mouth of a present-day coastal river. The Pistaque and Saco Rivers are two examples which illustrate this correlation. The paths traced across the nearshore shelf probably represent deposits containing both estuarine and marine sediments.

Along the coast of New Hampshire, Tuttle (1960) investigated the evolution of about 15 km of shoreline. Fig. I-102 illustrates rather vividly many of the concepts behind transgression along the New England shoreline. Wave action, concentrated on headlands, erodes soft glacial deposits and redistributes the material across the mouths of shallow bays. Boulder pavements remain behind where hills of glacial till were eroded. Cliffs of unconsolidated glacial debris reach up to 12 m in height along this section of coast. Shoreface erosion is essentially halted along a headland whenever a bedrock ledge is exposed (Tuttle 1960) 1960). Fig. I-102 also shows the extent of marsh development along this section of coast. These low areas correspond to river valleys, estuaries and local lowlands between headlands. Preservation of the subaerial surface will be patchy in this region. A short distance from the beach, preservation of large sections of the subaerial surface will most probably be restricted to major embayments and drowned river valleys (see for example Oldale and others 1973; Tuttle 1960).

A section of the large marsh in the bottom of Fig. I-102 was sampled by Keene (1971). He obtained basal peats which gave sea-level relationships back to 6850 ± 155 B.P. His samples were extracted about 4 km inland along a marsh-filled estuary. Closer to the beach older peats probably exist beneath the marsh. At the locations sampled by Keene

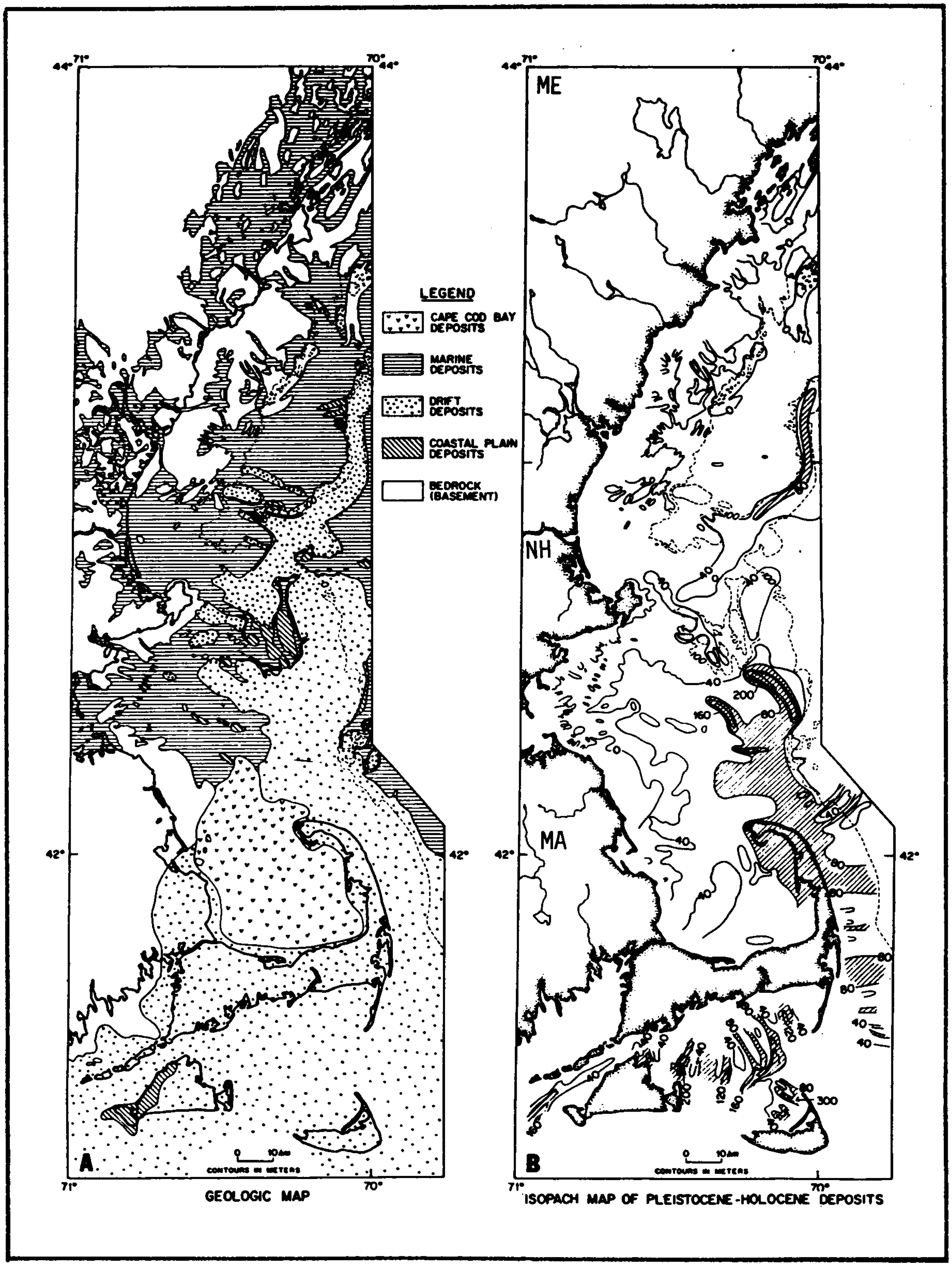


Fig. I-101

Geologic map (A) and thickness of Quaternary deposits in the Western Gulf of Maine based on the study done by Oldale and others 1973. After Emery and Uchupi 1972.

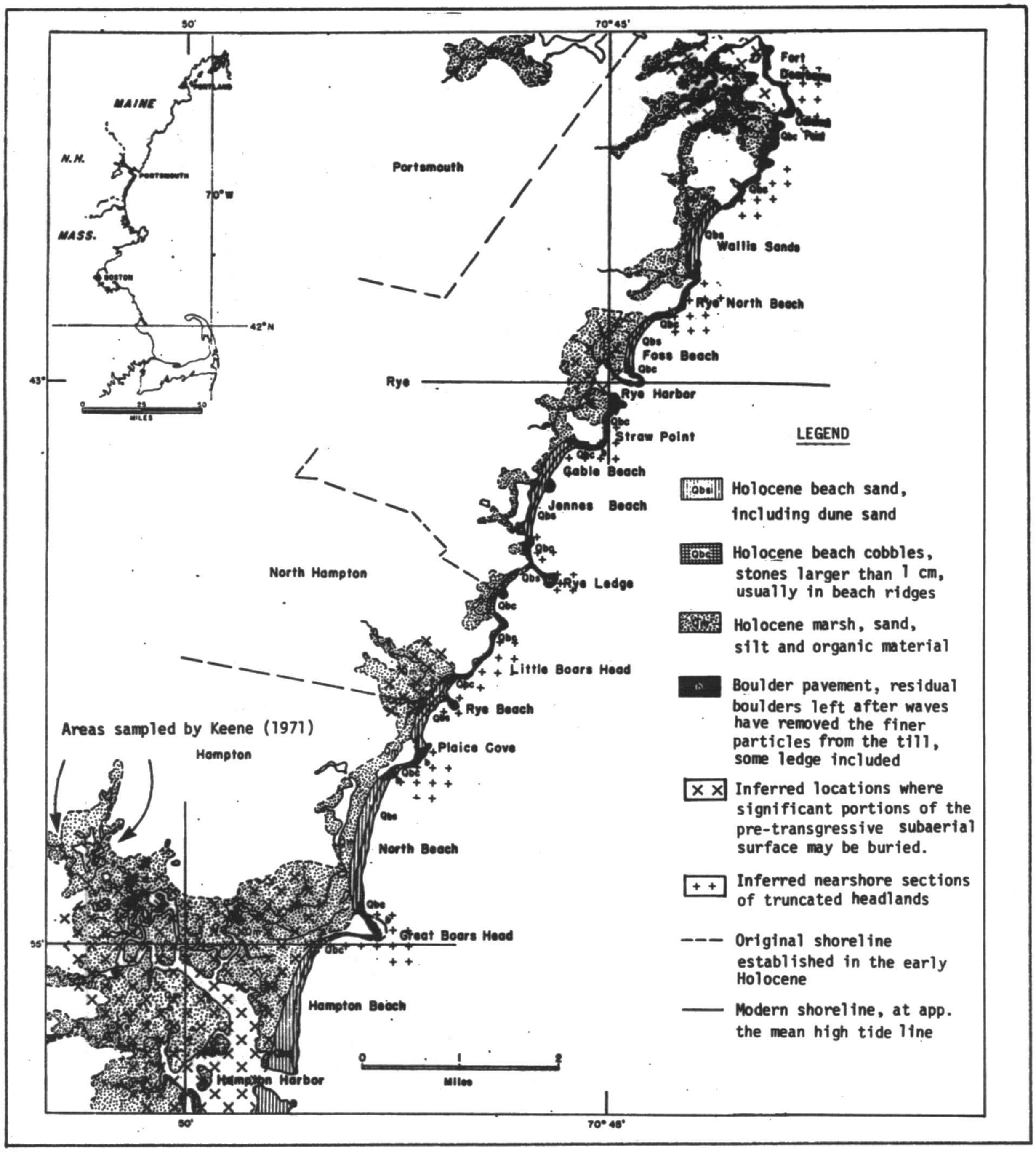


Fig. I-102

New Hampshire shoreline and areas of potential preservation of the pre-transgressive subaerial surface as inferred from the data given above (Tuttle 1960). Truncated headlands located between marshes and estuaries are the least likely areas where the pre-transgressive subaerial surface may be retained after erosional shoreface retreat. The estuaries and marsh filled valleys and lowlands, on the other hand, have the greatest chance for preserving a buried portion of the pre-transgressive subaerial surface. Adapted from Tuttle (1960).

(1971), up to 11.2 m of organic sediment, silts and clays had accumulated over Late Wisconsin glacial till or bedrock. Further seaward, Keene notes sequences in excess of 15 m thick, offering a good chance for preservation of the Holocene transgressive subaerial surface.

Central and northern Maine produce a similar picture of the preservation of the Holocene subaerial surface. Except for sea-level research (Schnitker 1974; Stuiver and Borns 1975) and the tentative identification of morainal deposits (Harbison 1969), there exists little research of use to archaeologists for determining the distribution of the buried subaerial surface.

14.0 SUMMARY OF THE LATE PLEISTOCENE AND HOLOCENE EVOLUTION OF THE SUBAERIAL CONTINENTAL SHELF

Archaeologists are very interested in knowing the physical environment of the Continental Shelf in order to reconstruct prehistoric land use and settlement patterns. The exposed CS did not offer a physically homogeneous or uniform environment. We know that Shelf topography was much more varied than that shown today on bathymetric charts, although we can only reconstruct the physical environment on exposed portions of the CS in a very general way because of the absence of pertinent data. Chart I-1b offers a summary of the major features associated with the CS during the Late Quaternary. Chart I-1b is a compilation of the data discussed in each section of the CS.

In order to understand the environment of the Shelf at about the time of earliest known occupation in the eastern United States, it is necessary to go back even further, to the period about 30,000-36,000 B.P. The surface morphology of the subaerial CS was initially established during the last marine regression (about 36,000 to 18,000 B.P. — Dillon and Oldale 1978; Milliman and Emery 1968). As the ocean retreated during the Late Wisconsin, subaerial processes replaced submarine processes in eroding and reshaping the newly formed landscape. If the last marine transgression had not taken place, it would be much easier to reconstruct the physical environment of the Shelf. Unfortunately, this most recent transgression has eroded and redistributed or buried most all of the Late Quaternary subaerial surfaces of the CS. As the previous sections have pointed out repeatedly, we have only a vague understanding of the major subaerial geomorphic features (major river channels, valleys, headlands) at present.

Marine regression occurred sometime between 36,000 and 18,000 B.P., eventually producing the lowstand associated with the last world-wide glaciation.

If marine regression took place rather uniformly, the Shelf would probably have been a region characterized by low beach ridges, prograding deltas, and gentle, seaward-dipping interfluves covered with trellis-like drainage systems. If, on the other hand, marine regression was sporadic and intermixed with periods of minor transgression, then the topography of the newly exposed Shelf would have been quite different. Scarps, estuary and coastal-plain terraces, overlapping deltas and similar morphologic elements would have been much more frequent, making the terrain more varied and complex.

The last marine regression probably oscillated between these two patterns, being characterized by periods of uniform sea level lowering intermixed with phases of stillstand and possibly significant transgression.

Unfortunately, a reconstruction of shelf terrain depends heavily upon a reconstruction of sea-level change between 36,000 B.P. and 18,000 B.P., a period for which few data are available. Fig. I-103 suggests that a highly unstable climate was characteristic during the period of the last regression, a suggestion derived from the oxygen-isotope ratio found in material from an ice core taken at Camp Century, Greenland (Dansgaard and others 1970; Langway and others 1973). However, the data presented in Fig. I-103 should not be accepted blindly, since it is partly interpretive and based on some complex theoretical assumptions regarding glacial ice flow, ice compression, oxygen-isotope ratios, and sampling techniques.

The last marine transgression has removed or buried most of the landforms and surfaces that were part of the subaerial CS. Based upon the data presented in the earlier discussion of Shelf subregions, the following generalization may be made for the period between 18,000 B.P. and the present. As before, the areas north and south of the Hudson Shelf Valley will be discussed separately.

In general, four major periods of environmental change can be recognized for the Shelf area south of the Hudson Shelf Valley. The first deals with the period of the lowstand, when the shoreline was located on the upper edge of the Continental Slope at about 18,000 B.P. During this time, river and stream entrenchment was at its maximum on the Outer Shelf and alluvial fans and deltas would have been frequent along river valley flanks and at the coastline. Sometime between 18,000 and 14,000 B.P. rapid sea-level rise brought the shoreline across the Shelf break and part of the Outer Shelf. This period would have witnessed major environmental changes as deltas on the Outer Shelf evolved into cuspate forelands or barrier coastlines and were finally submerged. Also during this period river valleys evolved into medium-sized estuaries as sea water pushed up their axes.

Between about 14,000 B.P. and about 7000 B.P. sea-level rise and erosional shoreface retreat brought the shoreline across the rather broad and flat region of the Middle Shelf. During this time, estuaries reach their maximum size with the largest ones protruding well over 100 km inland from the coastland. Also during this period, barriers formed along much of the coast, protecting lagoons, saltmarsh, and swamps on their landward side. The areas of fresh and salt water marsh probably reached their maximum near the middle of this period as the transgressing sea pushed across the low uplands flanking the sides of former estuary retreat paths.

The last period, from about 7000 B.P. to the present, represents a time of slowed sea-level rise. It is a period of shrinking estuaries as river-valley infilling overtakes sea-level rise. It is also a period during which barriers are pushed landward faster than lagoon shorelines,

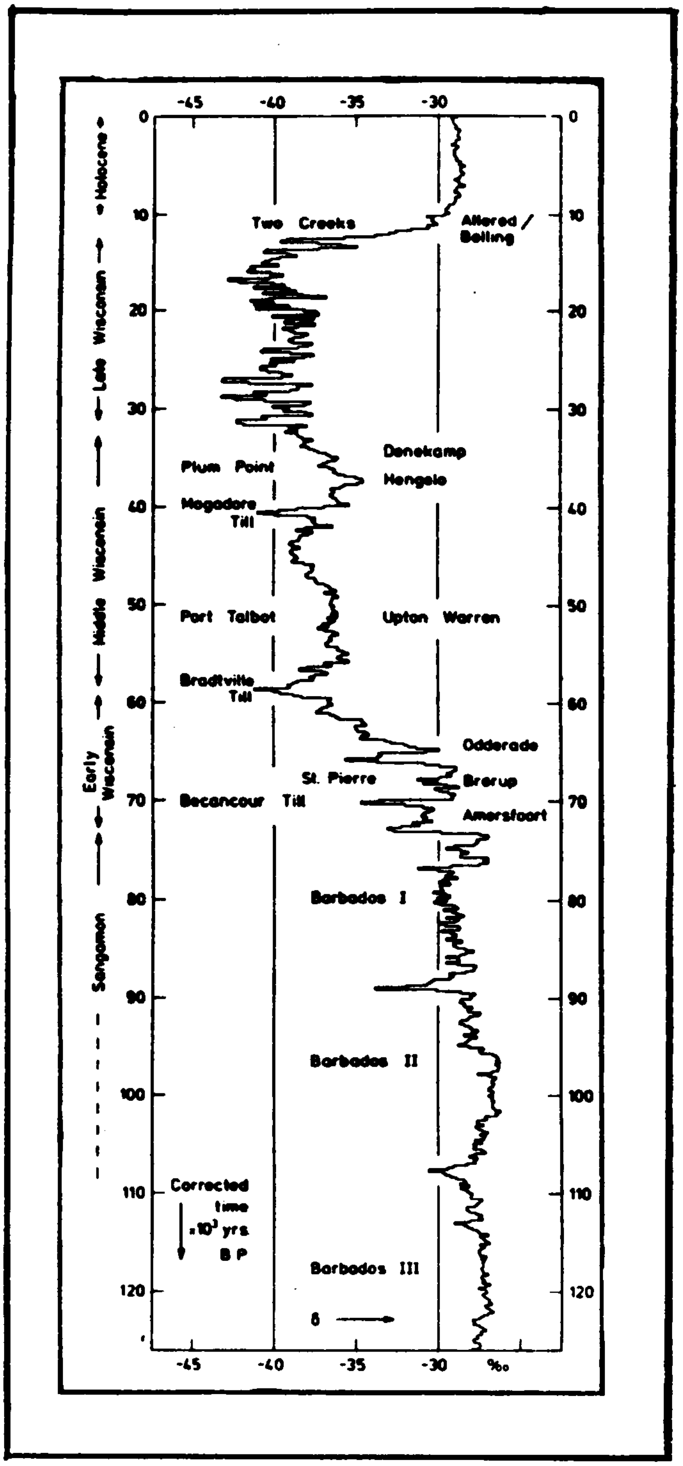


Fig. I-103

Climatic fluctuations during the Late Pleistocene, based on oxygen isotope ratios from the Camp Century Ice Core, Greenland. Nearly 7,000 samples were used to construct this graph. The interpretation of the graph is based on important theoretical assumption regarding glacial ice flow, oxygen isotope ratios, and sampling techniques. After Langway and others (1973).

resulting in a slow but steady shrinkage in lagoon size. Some lagoons also shrink in size because of salt-marsh progradation. Consequently, the last 7,000 years may be characterized by the reduction in size of some coastal environments, such as lagoons and estuaries, and the growth and expansion of coastal marshes.

North of the Hudson Shelf Valley, the following general environmental changes have occurred since 18,000 B.P. Between 18,000 and 15,000 B.P. Late Wisconsin glaciers were pushing toward their maximum advance. By 15,000 B.P. stagnation of the glacial front was in progress and deglaciation soon followed. Along the Inner Shelf of southern New England, Long Island Sound, and the western flank of Georges Bank, outwash plains flanked end moraines. North of these moraines, numerous late glacial and proglacial lakes were formed (Oldale and others 1973). Between 15,000 and 11,000 B.P. many of these lakes drained as moraines were breached and better drainage corridors established.

Environments on the Outer Shelf were also changing along the Long Island-Southern New England Shelf between 18,000 and 15,000 B.P. The shoreline advanced over the shelf break as sea-level rise pushed it landward. The large deltas were submerged after passing through several intermediate stages. Estuaries which had been small in size grew larger as ocean water invaded them. Between 15,000 and 11,000 B.P., deglaciation cleared the Gulf of Maine. Instead of exposing the mainland shelf along northern Massachusetts, New Hampshire, and Maine, isostatic depression caused marine inundation to follow the retreating glacial front.

Along the Long Island-Southern New England Shelf isostatic adjustment reestablished the crustal relationships found today. Transgression placed the shoreline along the Middle Shelf by 10,000 B.P. Estuaries increased in size and barrier coastlines were also established during this period. The greater relief found on this Shelf indicates that slightly different environments may have existed during Shelf submergence than those found south of the Hudson Shelf Valley. Large and medium-sized estuaries would have been flanked by slightly elevated uplands instead of the broad marsh and swamplands common south of the area during transgression. Barrier coastlines may have started at a slightly later date along the Long Island Shelf than those found along the Shelf to the south (9000 to 7000 B.P. rather than possibly as early as 13,00-14,000 B.P.).

By 9000 B.P., emergence of the coast of Maine, New Hampshire and northern Massachusetts exposed a thin strip 10 to 20 km in width. This newly exposed region would have consisted of many poorly drained areas. Over the next few thousand years, however, better drainage systems were developed and some poorly drained areas filled with organic material and sediment. In the last 9,000 years, transgression has filled many of the older river valleys as estuaries have retreated landward. For the last 5,000 years and possibly longer, the coastline of Maine, New Hampshire, and northern Massachusetts has probably remained relatively unchanged with the exception of infilling estuaries and broadening tidal marshes.

The last 7,000 years have witnessed some major changes along the coastline of Long Island and southern New England. Transgression has enlarged Long Island Sound, taking with it a part of the mainland and islands. Nantucket, Martha's Vineyard, and Long Island have been substantially reduced in size by sea-level rise and the shoreface erosion during this period. Lagoons along the mainland and flanking some islands have also been reduced in size by decreasing sea-level rise. Estuaries have continued to be filled and marshes have spread over many areas that once were covered with shallow water.

The above has been a brief discussion of major changes occurring to the physical environment during transgression. The major subaerial features that once existed on the Shelf are shown on Chart I-1b. Chart I-1a gives the approximate position of shorelines at 3,000-year intervals from 18,000 B.P.

At 18,000 B.P. four major river systems flowed across the exposed Shelf to meet the ocean near the shelf break (Chart I-1b). These major river systems were the Chesapeake, the Delaware, the Hudson, and the Block. Each drained a fairly large area which lay both inland of today's shoreline and on the exposed Shelf. The lowstand associated with the Late Wisconsin caused rivers to become entrenched along the shelf break. Tributaries also became entrenched along the sides of major river valleys. The best description of Late Pleistocene entrenched topography is to be found in several studies performed in the vicinity of Cape Henlopen, Delaware (Kraft 1971; 1977; Kraft and others 1978; Sheridan and others 1974).

Besides the four major river systems discussed above, several minor rivers were also in existence at 18,000 B.P. From south to north, they were the Roanoke River, Albemarle River, James River, Great Egg River, New Jersey Shelf River, and the Long Island Shelf River. Some smaller river systems drained the Nantucket Shoals-Georges Bank region.

Chart I-1b shows the major river systems and some of the smaller rivers that have been identified to date. During the last lowstand, these rivers helped to drain the exposed CS. As transgression began, sometime between 18,000 and 15,000 B.P., estuaries developed along each major river system. At first the size of the estuaries was limited by the slope of the Outer Shelf. Some deep entrenchment of rivers occurred at the shelf break but did not extend very far inland. Instead of deep entrenchment, the major rivers developed fairly broad and shallow river valleys along the Middle and Inner Shelves (Twichell and others 1977). Along the major valley flanks, however, stream entrenchment did occur as pointed out previously (Kraft 1971; Sheridan and others 1974).

The growth of large estuaries during the Early Holocene is an important change to recognize. As transgression brought the shoreline to the middle portion of the Continental Shelf, large estuaries developed along the Chesapeake, Delaware, Hudson and Block River systems. Slightly smaller estuaries developed along the other rivers shown on Chart I-1b. This development evidently began about 12,000 B.P. given sea level and shelf topography at that time. Sometime between 11,000 and 7000 B.P. maximum estuary size was reached. Since about 7000 B.P., estuaries have been dwindling in size as sea level rise decreased and infilling occurred.

The Gulf of Maine passed through a series of glacial and postglacial environments between 15,000 and about 11,000 B.P. Maximum shelf emergence along the Maine coast probably occurred around 10,000 B.P. Since that time, transgression has taken place. It is reasonable to speculate, on the basis of bathymetry, that the configuration of the coastlines of Maine, New Hampshire, and eastern Massachusetts has been quite similar during the Holocene in terms of estuaries and embayments.

Another important change which also occurred during the Early Holocene was the development of barrier protected coastlines along the Middle Atlantic Bight. Evidence collected so far has indicated that many buried lagoonal deposits along the Inner Shelf on the Bight were active between 9000 and 6000 B.P. (Sanders and Kumar 1975a, 1975b; Stahl and others 1974; Stubblefield and others 1975; Stubblefield and Swift 1976; Swift 1975a; Swift and others 1972).

It is not unrealistic to consider that barrier lagoon complexes may have developed as transgression took place along the Outer Shelf. No lagoon deposits have been identified on the Outer Shelf so far but then there have been few attempts to do so.

Aside from the shelf features shown on Chart I-1b and the approximate position of shorelines given in Chart I-1a, the available data permit few additional insights into the physical environment of the shelf and its subaerial component prior to transgression. Until research is directed towards identifying the pre-transgressive subaerial surface, little of substantive value may be said regarding the soils, drainage, and topography of the exposed Shelf. As this study has illustrated, geologists have not actively pursued the reconstruction of the Late Quaternary subaerial CS. It is anticipated that in the future much more information will be available as multidisciplinary research teams explore the Shelf and its Late Quaternary deposits. We hope that this study has clearly pointed out those areas which desperately need attention.

15.0 A PATTERN FOR THE PRESERVATION OF THE PRE-TRANSGRESSIVE SUBAERIAL SURFACE

The previous sections have reviewed in detail information about the evolution of the Continental Shelf since 18,000 B.P. It is obvious that to date most geologists have not been particularly interested in the fate of the subaerial surface during transgression. Few studies provide specific information about the presence or absence of a buried subaerial surface. For this reason, it is necessary to approach the problem of subaerial surface preservation both indirectly and hypothetically.

Throughout the last 18,000 years, the rate of sea level rise has not been great enough to drown coastal regions intact. Instead, erosional shoreface retreat has generally eroded and redistributed from 10 to 15 m of unconsolidated sediments during the process of transgression. Thus, for the subaerial surface to be preserved intact, it must be buried beneath enough sediment to protect it from being disrupted by erosional shoreface retreat.

The two most important landform continua capable of burying a subaerial surface beneath enough sediment to protect it are marsh-lagoon-barrier systems and flood plain-marsh-estuary systems. The subaerial surface is least likely to be preserved along unprotected oceanic shorelines or cliff-backed shorelines of unconsolidated material.

The United States Continental Shelf north of Cape Hatteras, North Carolina is mapped in Figs. I-104 to I-116 using the following classification system regarding the preservation of the pre-transgressive subaerial surface: considerable preservation, partial preservation, negligible preservation, and absent (subaerial surface not present since 18,000 B.P.). See below (p. 207) for fuller definition of these terms. This classification system is relative, since little information is available on the distribution or extent of buried flood plain, marsh, lagoon or estuary deposits or buried truncated headlands. In order to extract adequate data it would be necessary to institute a program of coring with ample radiocarbon dating and detailed seismic profiles performed systematically so as to allow several dozen observations per sq km on the sediment immediately beneath the "transgressive sand sheet." Such a program has not been undertaken, however, the only significant exception being the research done by Stubblefield and Swift (1976) off central New Jersey. Their use of vibracores, radiocarbon dates, and seismic profiling on the sediment just beneath the "sand sheet" enabled us to determine that preservation of the subaerial surface may be on the order of 92% per unit area along this portion of the Middle Shelf (see Stubblefield and Swift 1976). The work of Kraft (1971, 1977) and Sheridan and others (1974) also allows absolute values to be placed on the amount of buried subaerial surface preserved.

Figs. I-104 to I-116 have been compiled both directly and indirectly from the published literature. The classification scheme is not meant to be used to represent absolute preservation values since not enough data are available to assign exact values. In the future, it would be extremely useful to do detailed subsurface testing in several shelf regions to determine actual preservation values. For now, only a general idea of the magnitude of preservation may be suggested for each group in this classification.

Considerable preservation generally means that it is hypothesized that somewhere between 40 and 100% of the subaerial surface is preserved per unit area. Along major river valleys this preservation class would probably fall near 100% and not fluctuate. Along barrier island-lagoon shorelines this value could vary sharply upward or downward, depending upon the pre-transgressive topography and the impact of tidal-inlet scouring.

Partial preservation, of course, means less areal preservation of the subaerial surface than in the category mentioned above. In terms of percentages, it may be useful to assign subaerial surface preservation values of from 5 to 40% per unit area. Partial subaerial preservation would be more common along the flanks of major river valleys where incised streams and fairly steep interfluves are frequent. Partial preservation may also occur along shorelines protected by barriers but where erosional shoreface retreat and inlet scour have frequently penetrated lagoonal deposits.

Regions with hypothesized subaerial-surface-preservation values less than 5% per unit area have been assigned to the negligible preservation class. Into this class would fall most unprotected oceanic shorelines (that is, non-estuary or non-barrier environments). Uplands, islands, and plateau-like features of the Shelf would most probably fall into this class as well. Georges Bank, Nantucket Shoals, and the submerged uplands on the southern side of the Hudson Shelf Valley have received more intensive wave and current erosion than lower-lying adjacent areas of the CS.

The last class is not concerned with preservation but rather is used to delimit areas which have not been subaerial since about 20,000 B.P. Instead, these areas were covered by water and have remained covered up to the present. In the Gulf of Maine, some areas also fall into this group because a subaerial surface did not form after deglaciation. In other words, marine transgression followed the retreating glacial front and submerged the Shelf (Borns 1973; Tucholke and Hollister 1973). To the south and along the Delmarva Outer Shelf, sea level certainly did not fall below -180 m and possibly did not even fall below -110 m during the last 20,000 years (Dillon and Oldale 1978).

To the four classes of subaerial-surface preservation mentioned above, one modification has been added. On Figs. I-104 to I-116, each preservation class has been given two symbols, one to denote that the classi-

LEGEND

Considerable preservation
(based on published data)

Considerable preservation
(deduced hypothetically)

Partial preservation
(based on published data)

Partial preservation
(deduced hypothetically)

Negligible preservation
(based on published data)

Negligible preservation
(deduced hypothetically)

Continuously covered by water since
18,000 B.P.

Probably covered by water since
18,000 B.P.

Fig. I-104
Relative amount of pre-transgressive subaerial surface preserved on the northern North Carolina-southeastern Virginia shelf.

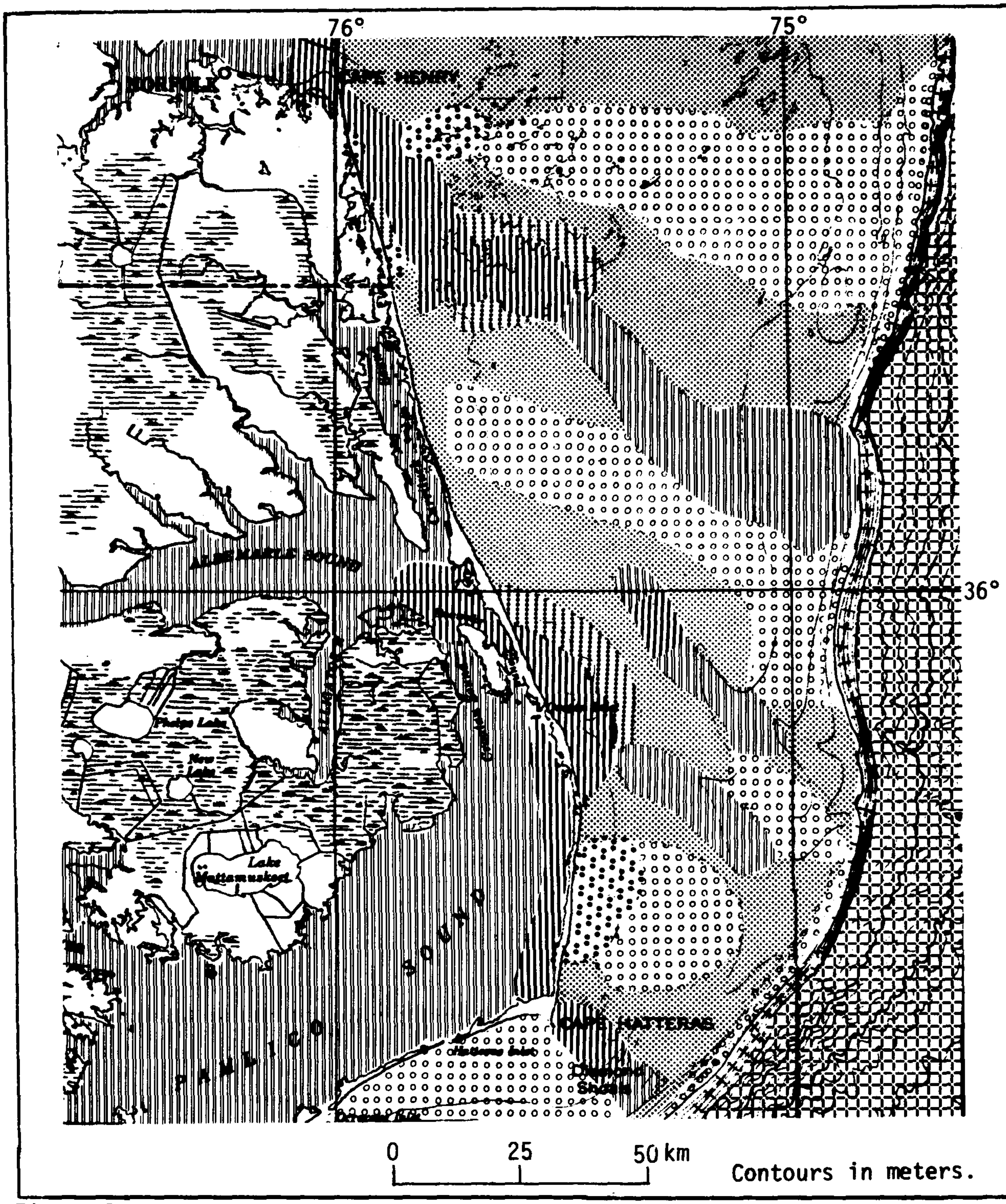


Fig. I-104 The amount of pretransgressive subaerial surface preserved on the northern North Carolina-southeastern Virginia shelf. Predictions based on information extracted from the following sources: Pierce and Colquhoun (1970); Shideler and others (1972, 1973); Shideler and Swift (1972); Swift (1975a); Swift and others (1972, 1977, 1978); Swift and Sears (1974).

LEGEND

Considerable preservation (based on published data)

Considerable preservation (deduced hypothetically)

Partial preservation (based on published data)

Partial preservation (deduced hypothetically)

Negligible preservation (based on published data)

Negligible preservation (deduced hypothetically)

Continuously covered by water since 18,000 B.P.

Probably covered by water since 18,000 B.P.

Fig. I-105
Relative amount of pre-transgressive subaerial surface preserved on the Delmarva shelf.

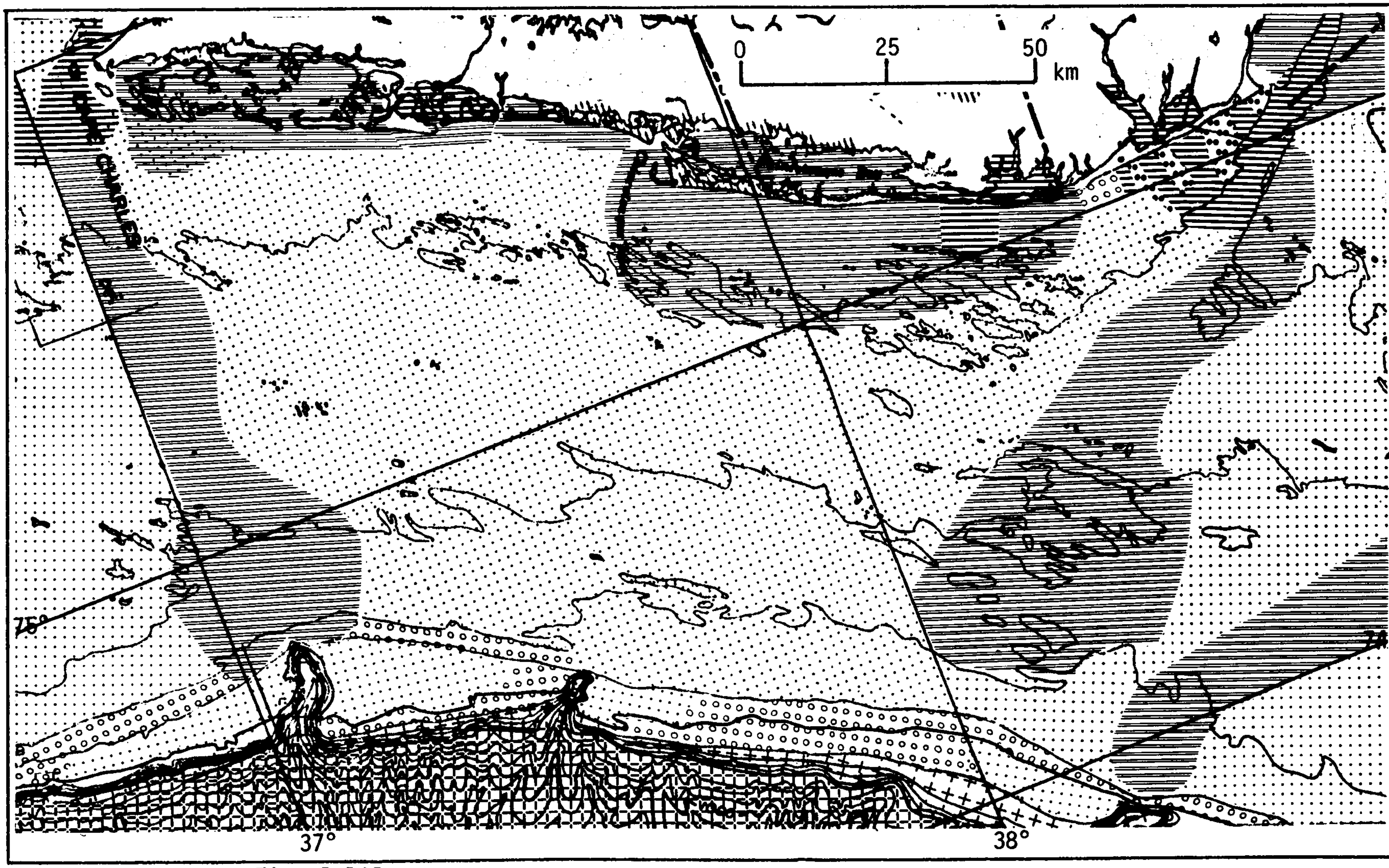


Fig. I-105

The amount of pretransgressive subaerial surface preserved on the Delmarva shelf. Predictions based on information extracted from the following sources: Harrison and others 1965; Kraft 1971, 1974, 1977; Meisberger 1972; Duane and others 1972; Swift and others 1972, 1978; Sheridan and others 1974, 1977; Swift and Sears 1974; Swift 1975a, 1976b; Field and Duane 1976; Twichell and others 1977; Belknap and Kraft 1977; Kraft and others 1978.Contours in meters.

LEGEND

Considerable preservation
(based on published data)

Considerable preservation
(deduced hypothetically)

Partial preservation
(based on published data)

Partial preservation
(deduced hypothetically)

Negligible preservation
(based on published data)

Negligible preservation
(deduced hypothetically)

Continuously covered by water since
18,000 B.P.

Probably covered by water since
18,000 B.P.

Fig. I-106
Relative amount of pre-transgressive subaerial surface preserved on the New Jersey shelf.

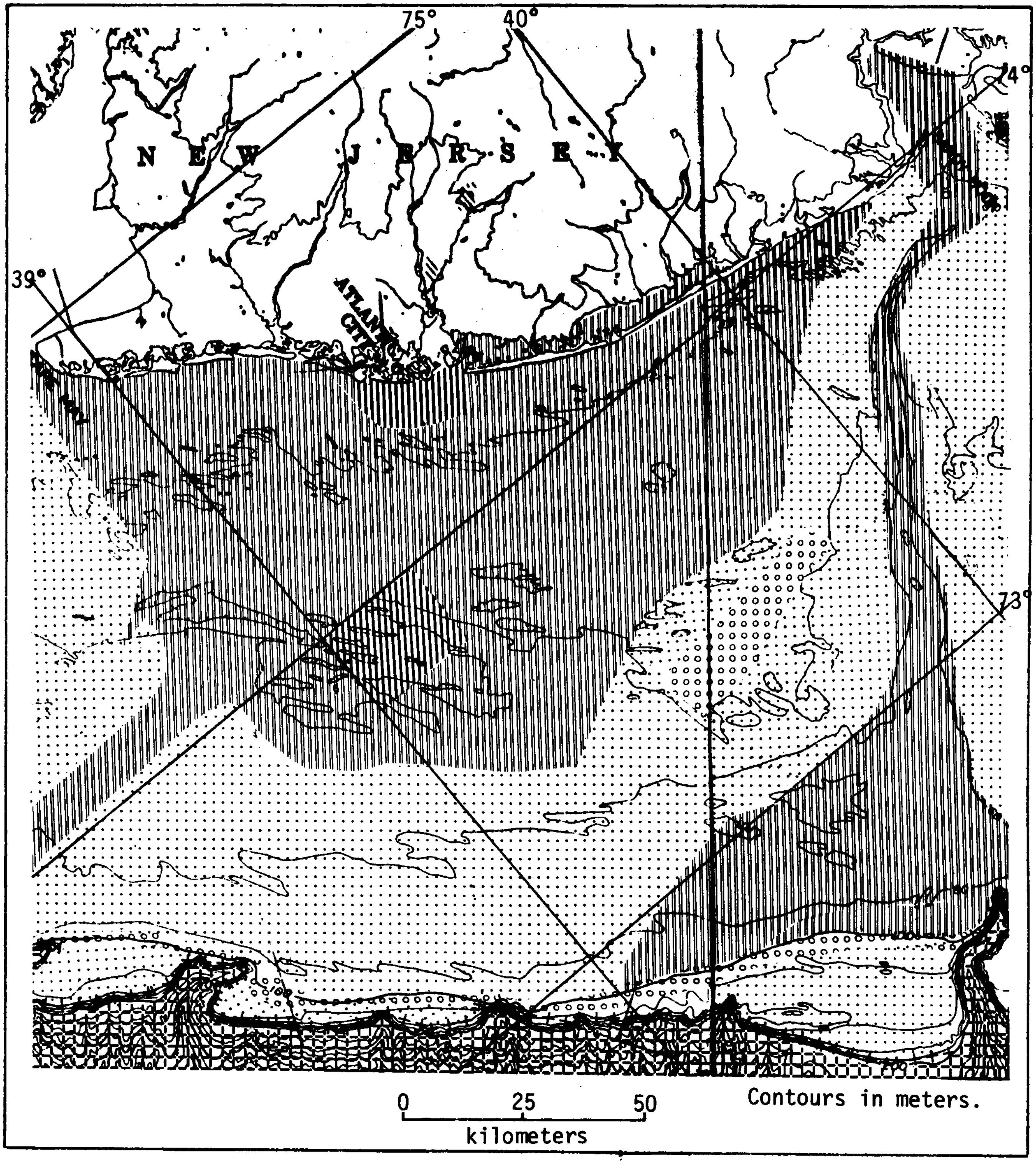


Fig. I-106

Amount of pre-transgressive subaerial surface preserved on the New Jersey Shelf. Assessments made using the following sources: Knott and Hoskins (1968); McClennan and McMaster (1971); Swift and others (1972, 1974); Swift (1973); Twichell and others (1977).

LEGEND

Considerable preservation
(based on published data)

Considerable preservation
(deduced hypothetically)

Partial preservation
(based on published data)

Partial preservation
(deduced hypothetically)

Negligible preservation
(based on published data)

Negligible preservation
(deduced hypothetically)

Continuously covered by water since
18,000 B.P.

Probably covered by water since
18,000 B.P.

Fig. I-107
Relative amount of pre-transgressive subaerial surface preserved on the Long Island shelf.

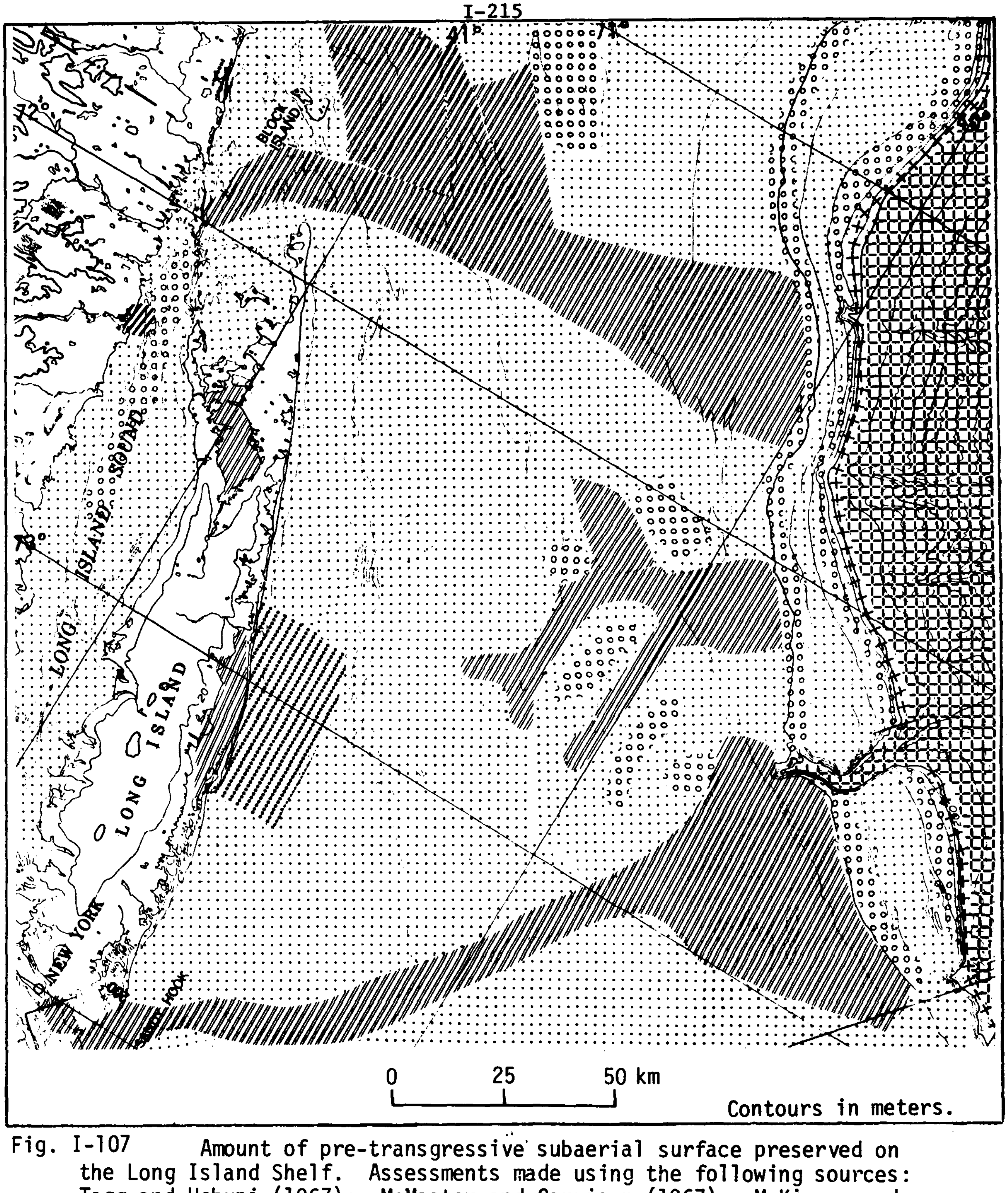


Fig. I-107 Amount of pre-transgressive subaerial surface preserved on the Long Island Shelf. Assessments made using the following sources: Tagg and Uchupi (1967); McMaster and Garrison (1967); McKinney and Friedman (1970); Grim and others (1970); Dillon (1970); Swift and others (1972); Schlee (1973); Caldwell and Sanders (1973); McMaster and Ashraf (1973a, 1973b, 1973c); Swift and Sears (1974); Sanders and Kumar (1975a); Swift (1977); Dillon and Oldale (1978).

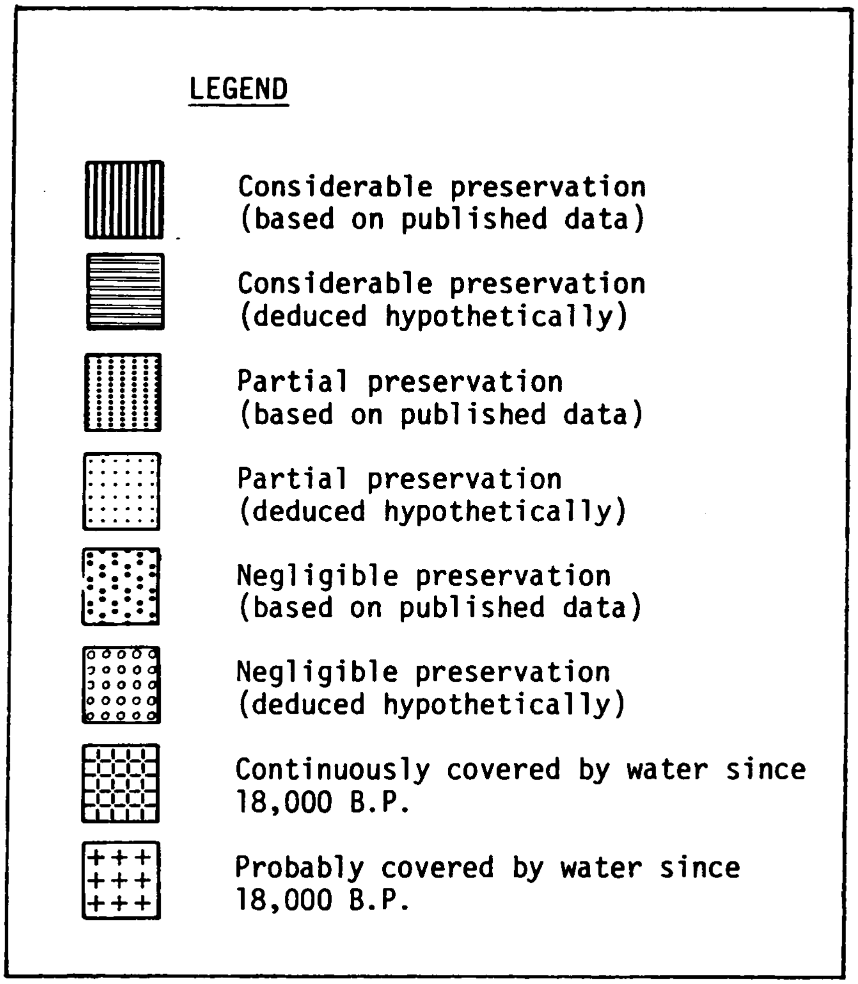


Fig. I-108
Relative amount of pre-transgressive subaerial surface preserved on the southeastern New England shelf.

Fig. I-108 Amount of pre-transgressive surface preserved on the southeastern New England Shelf. Assessments made using the following sources: Pratt and Schlee (1969); Schlee (1973); McMaster and Ashraf (1973a, b, c); Uchupi (1968); Oldale and Uchupi (1970); Garrison (1970); Garrison and McMaster (1966); Groot and Groot (1964).

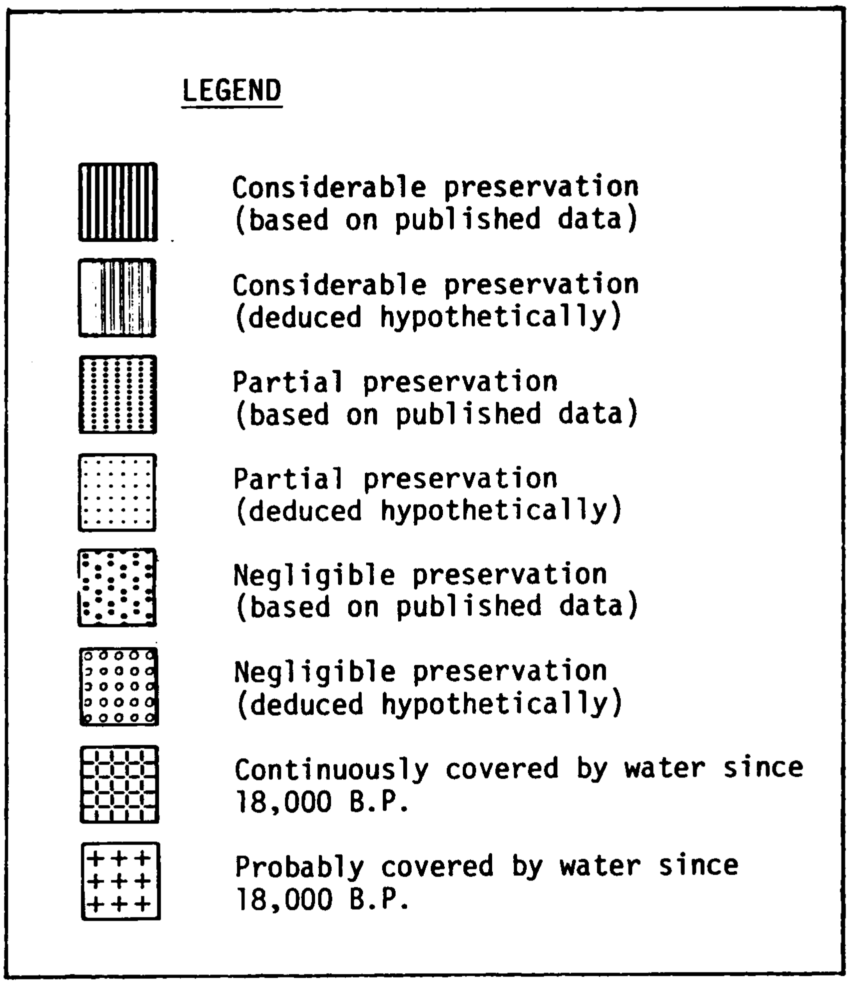


Fig. I-109
Relative amount of pre-transgressive subaerial surface preserved on Georges Bank.

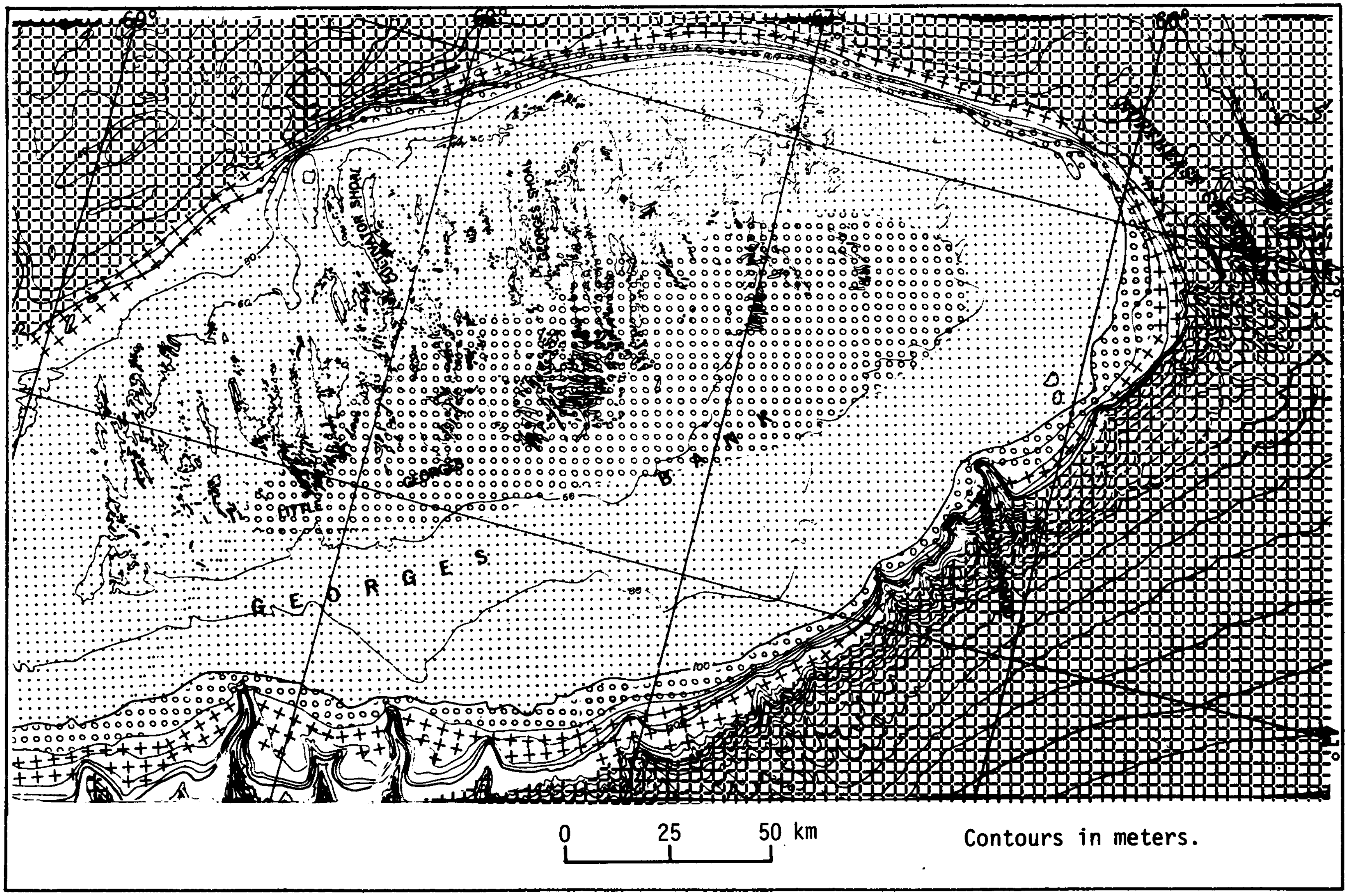


Fig. I-109
Amount of pre-transgressive surface preserved on Georges Bank. Assessment based on shelf bathymetry and the following sources: Emery and Uchupi (1965); Pratt and Schlee (1969); Uchupi 1968; Knott and Hoskins (1968).

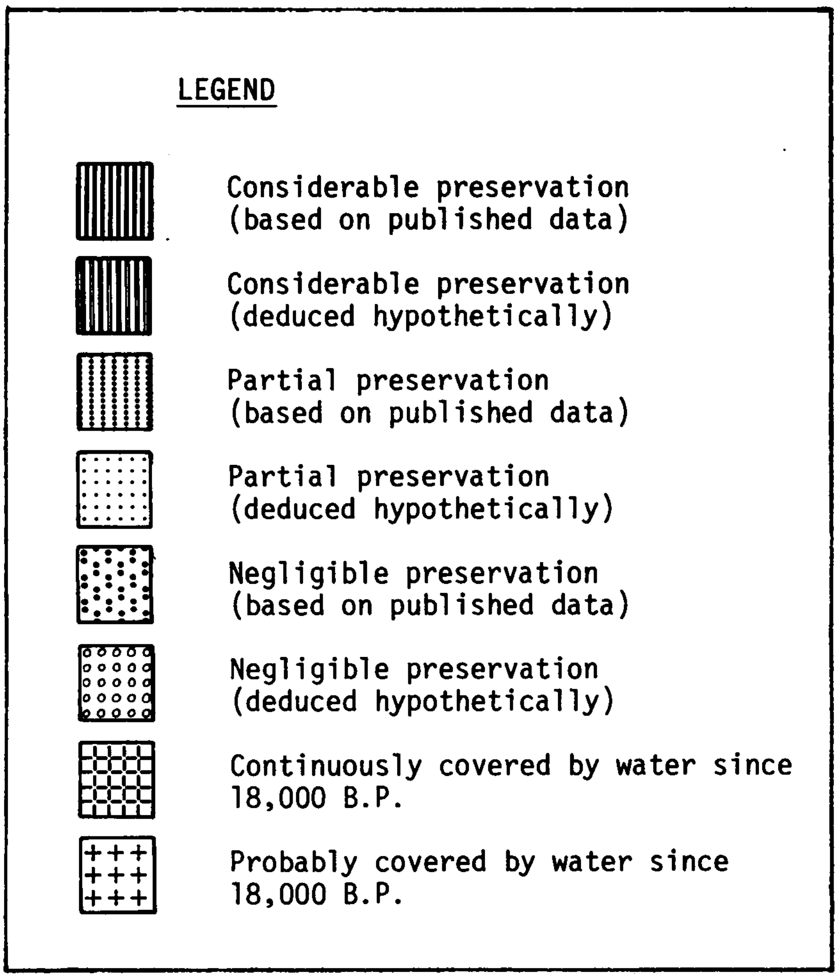


Fig. I-110
Relative amount of pre-transgressive subaerial surface preserved in the southern Gulf of Maine.

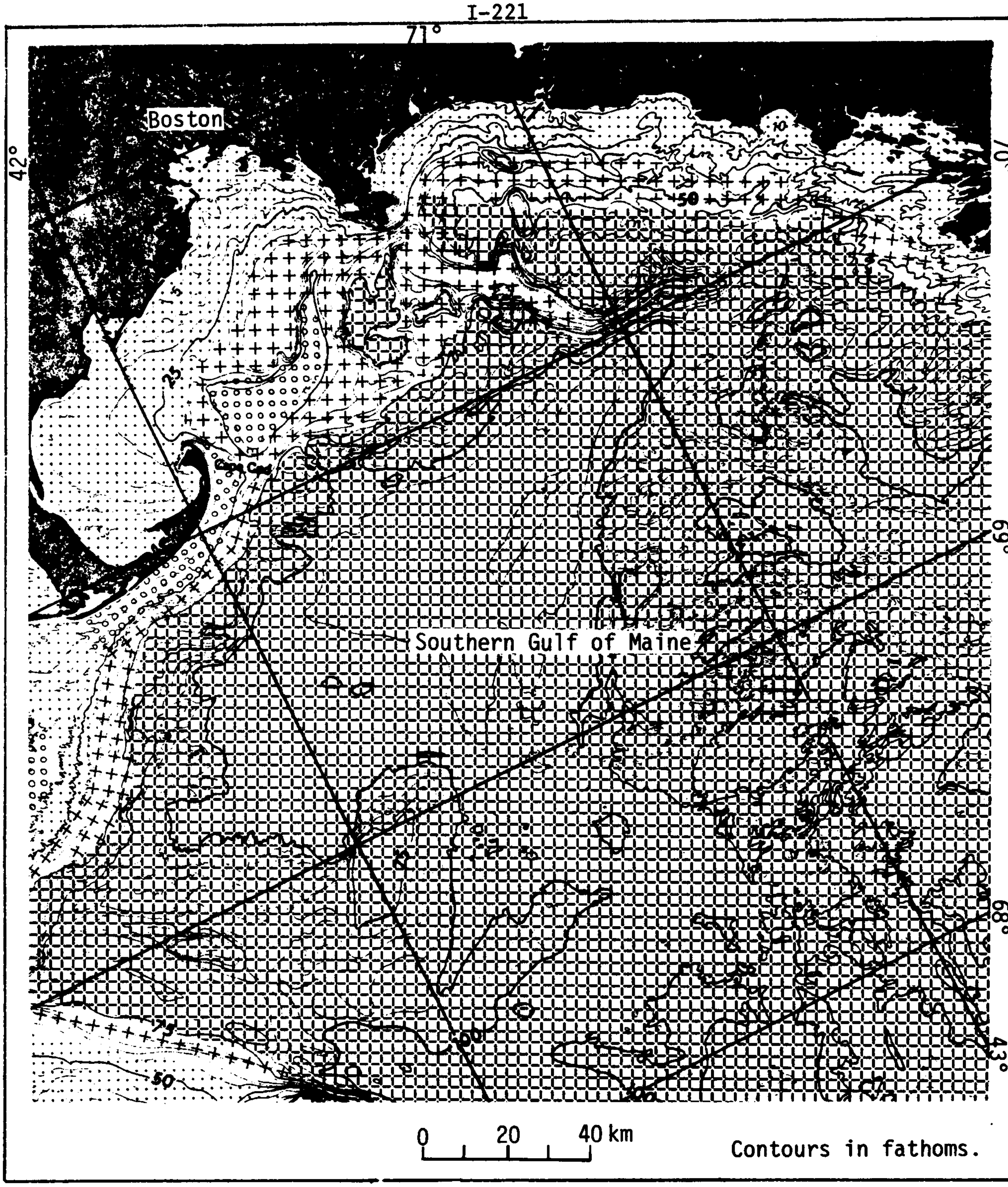


Fig. I-110

Amount of pre-transgressive subaerial surface preserved in the southern Gulf of Maine. Assessments based on sea level and deglaciation data (Tucholke and Hollister 1973 Dillon and Oldale 1978; Keene 1971; Born 1973; Flint and Gilbert 1976; Hoskins and Knott 1961; Kaye and Barghoorn 1964).

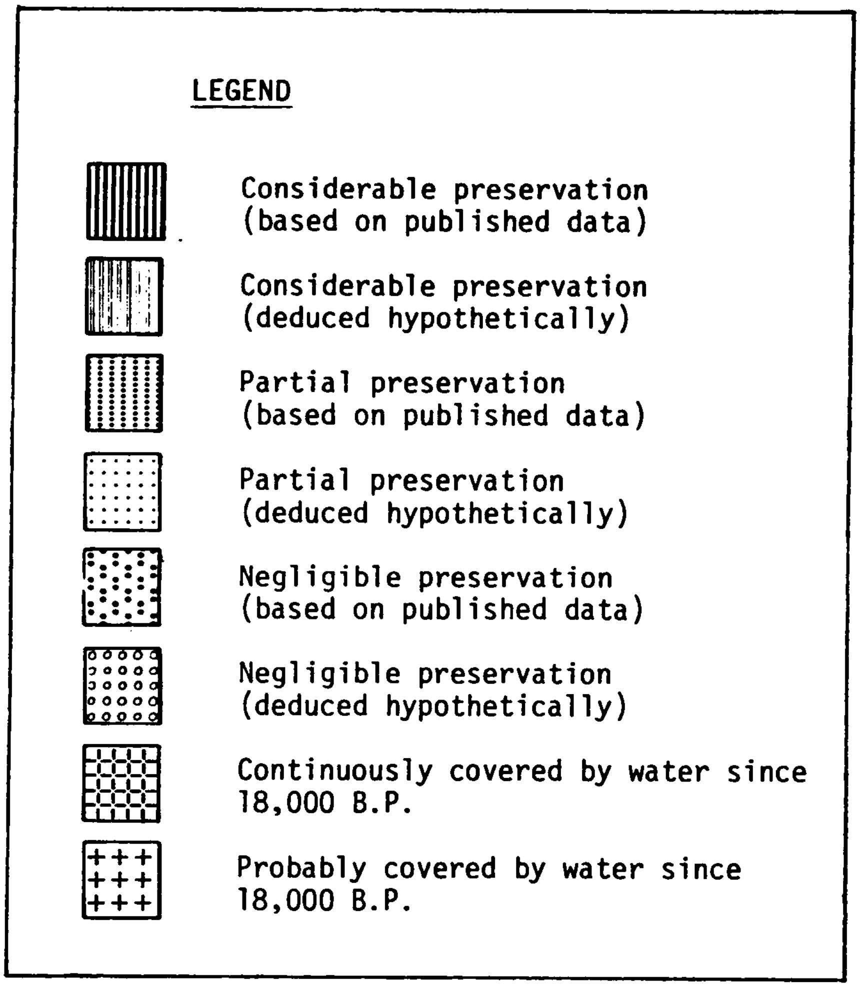


Fig. I-111
Relative amount of pre-transgressive subaerial surface preserved in northern Gulf of Maine.

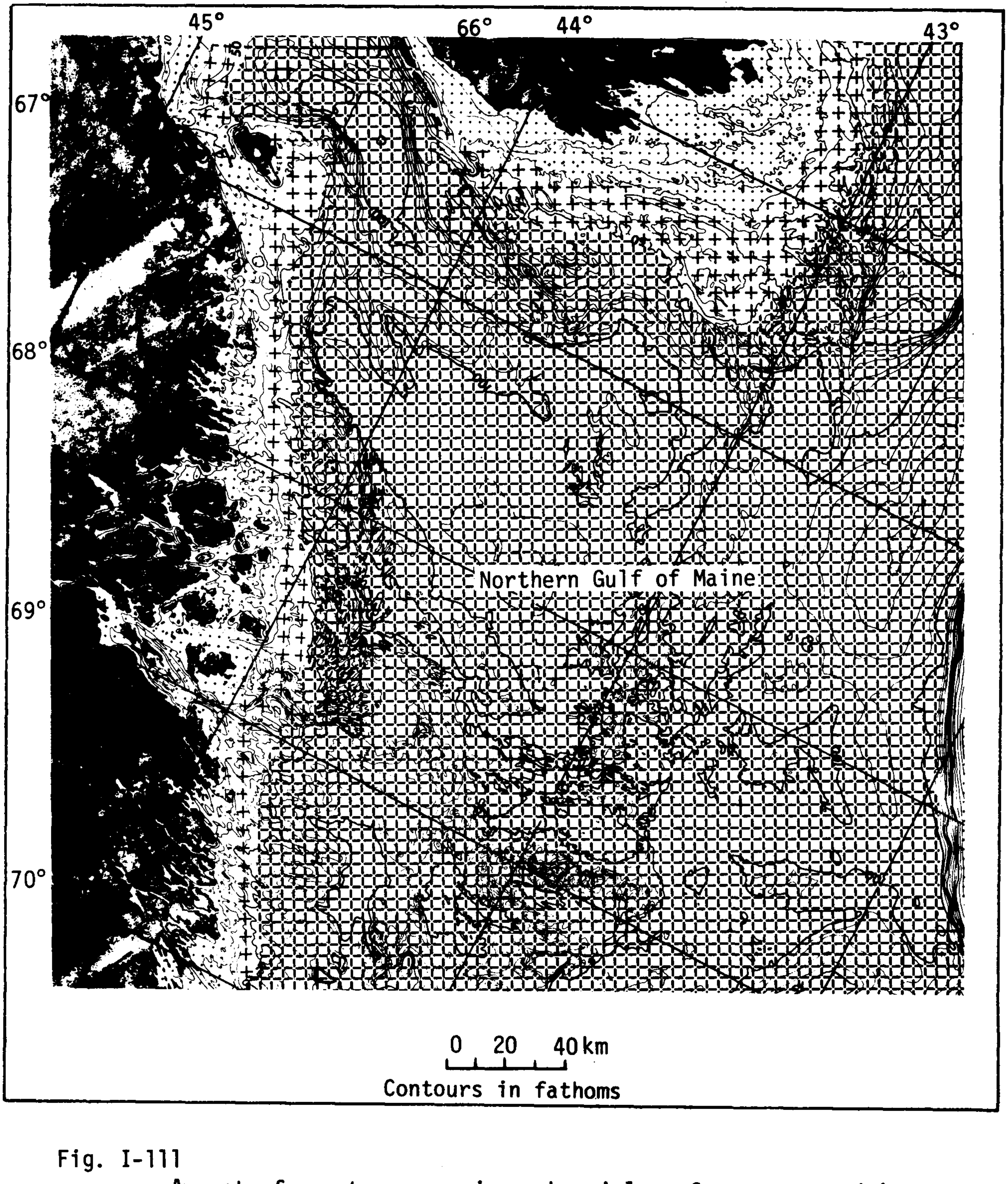


Fig. I-111

Amount of pre-transgressive subaerial surface preserved in northern Gulf of Maine. Assessments based on sea level and deglaciation data and the following sources: Grant (1970); Schnitker (1974); Tucholke and Hollister (1973).

LEGEND

Areas with adequate geological information

Areas with partially adequate information

Areas with little or no useful information for assessing the preservation of the subaerial surface

Fig. I-112

Amount of geological literature available for making an assessment of the preservation of the pretransgressive subaerial surface on the northern North Carolina-southeastern Virginia shelf.

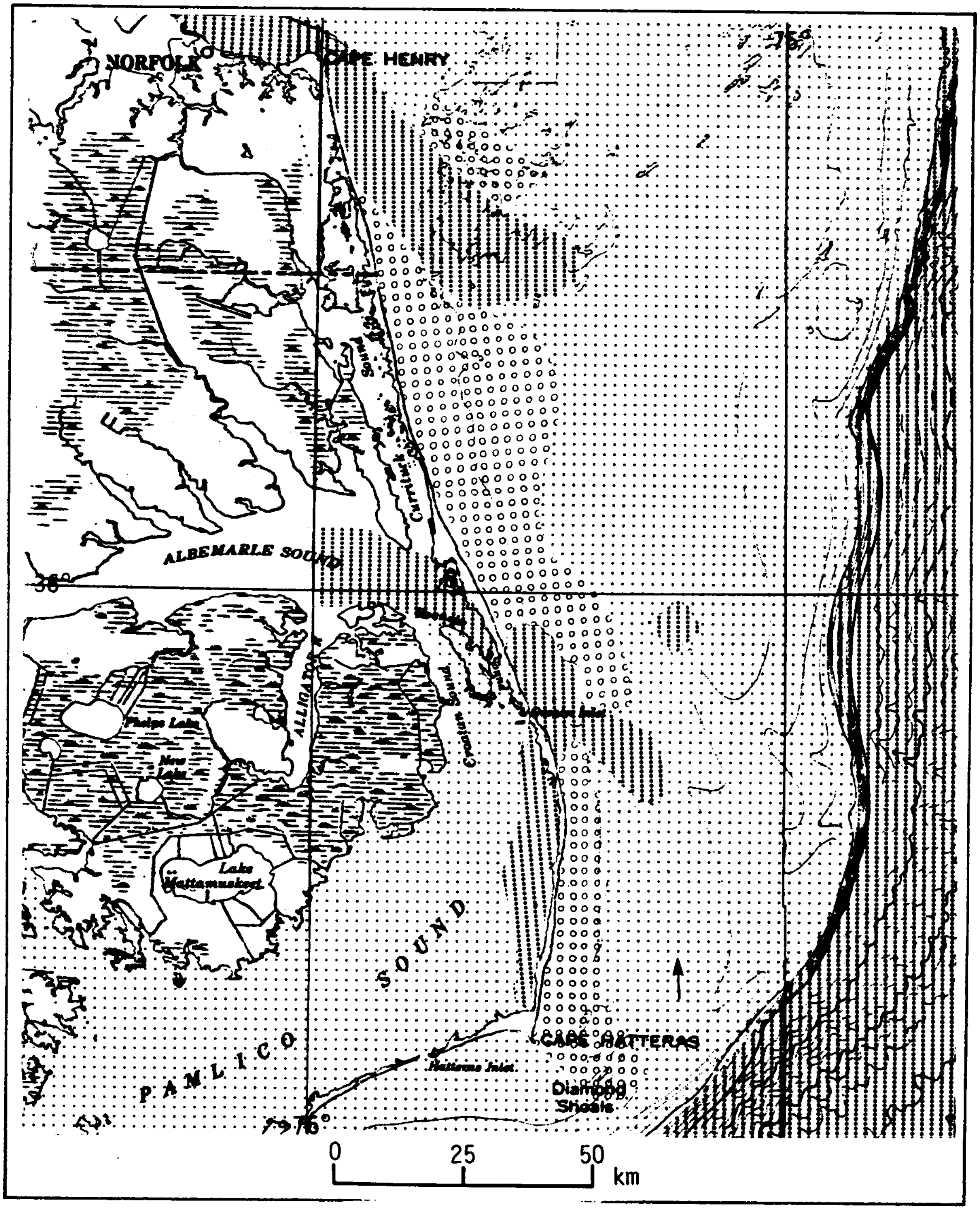


Fig. I-112

Contribution of literature for assessing the amount of pretransgressive subaerial surface preserved on the northern North Carolina-southeastern Virginia shelf. Information drawn from the following sources: Pierce and Colquhoun (1970); Shideler and others (1972, 1973); Swift (1972, 1975a); Swift and others (1972, 1977, 1978); Swift and Sears (1974).

LEGEND

Areas with adequate geological information

Areas with partially adequate information

Areas with little or no useful information for assessing the preservation of the subaerial surface.

Fig. I-113
Contribution of the literature for assessing the amount of pre-transgressive subaerial surface preserved on the Delmarva Continental Shelf. Sources used to compile this figure are: Dillon and Oldale (1978); Duane and others (1972); Field and Duane (1976); Kraft (1971, 1974, 1977); Kraft and others (1978); Sheridan and others (1974, 1977); Swift (1975a, 1976b); Swift and Sears (1974); Swift and others (1972, 1978); Twichell and others (1977).

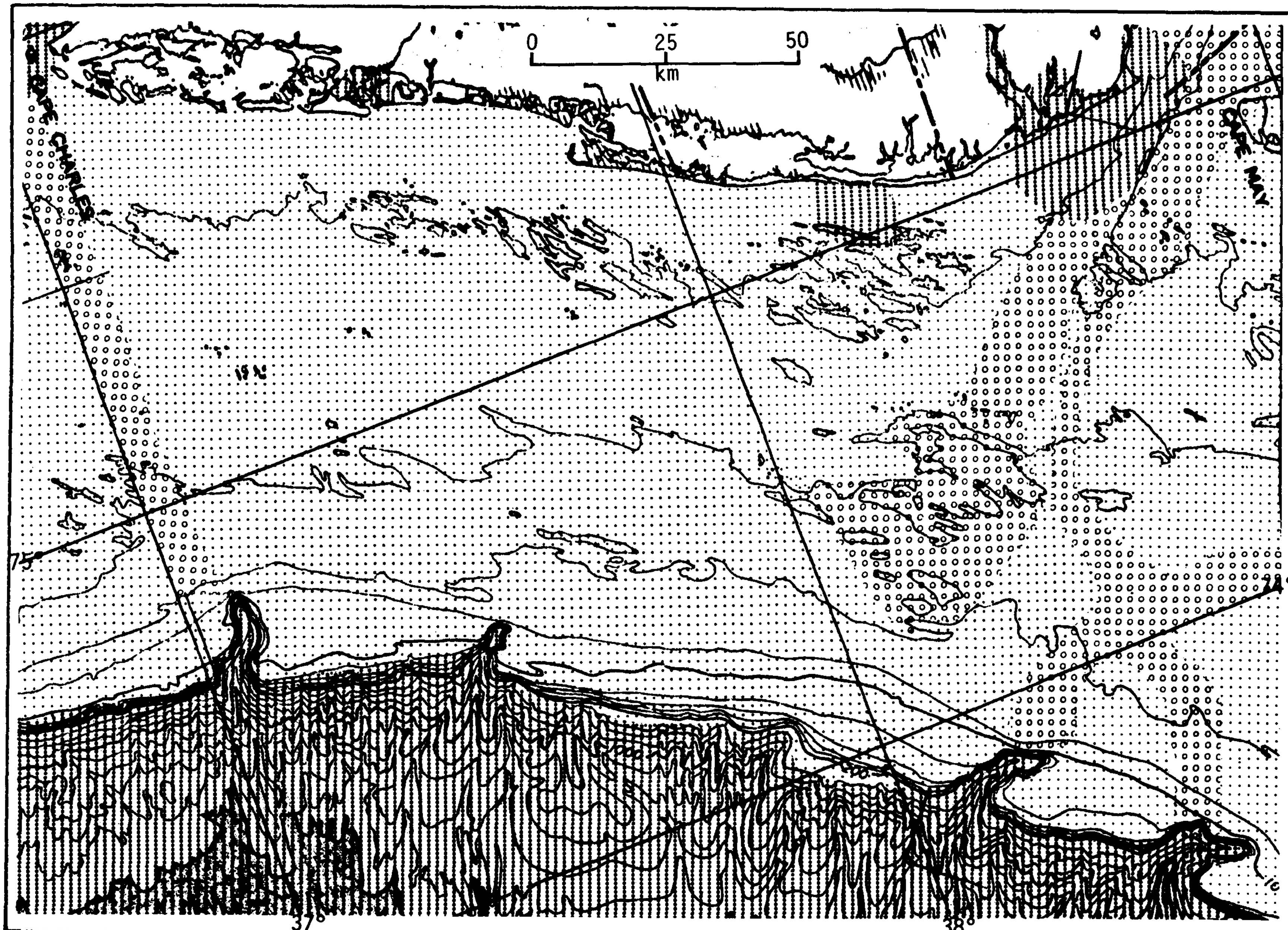


Fig. I-113

Amount of geological literature available for making an assessment of the preservation of the pre-transgressive subaerial surface on the Shelf. Contours in meters.

LEGEND

Areas with adequate geological information

Areas with partially adequate information

Areas with little or no useful information for assessing the preservation of the subaerial surface.

Fig. I-114

Amount of geological literature available for making an assessment of the preservation of the pre-transgressive subaerial surface on the Shelf.

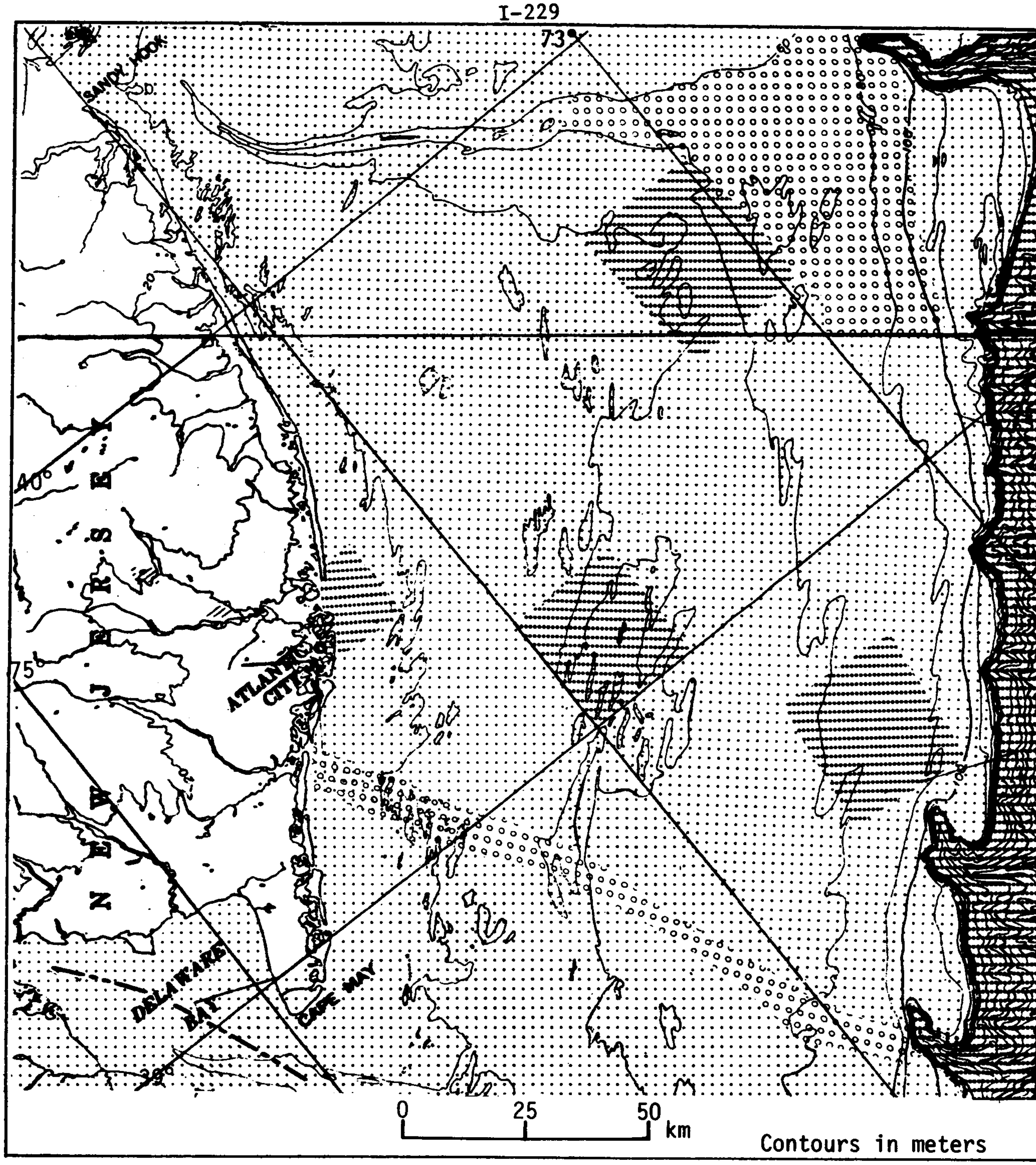


Fig. I-114

Contribution of the literature for assessing the amount of pre-transgressive subaerial surface preserved along the New Jersey Shelf. Sources used to compile this figure: Dillon and Oldale (1978); Knott and Hoskins (1968); McClennan and McMaster (1971); Swift (1973); Swift and Sears (1974); Swift and others (1972); Twichell and others (1977).

LEGEND

Areas with adequate geological information

Areas with partially adequate information

Areas with little or no useful information for assessing the preservation of the subaerial surface.

Fig. I-115

Amount of geological literature available for making an assessment of the preservation of the pre-transgressive subaerial surface on the Shelf.

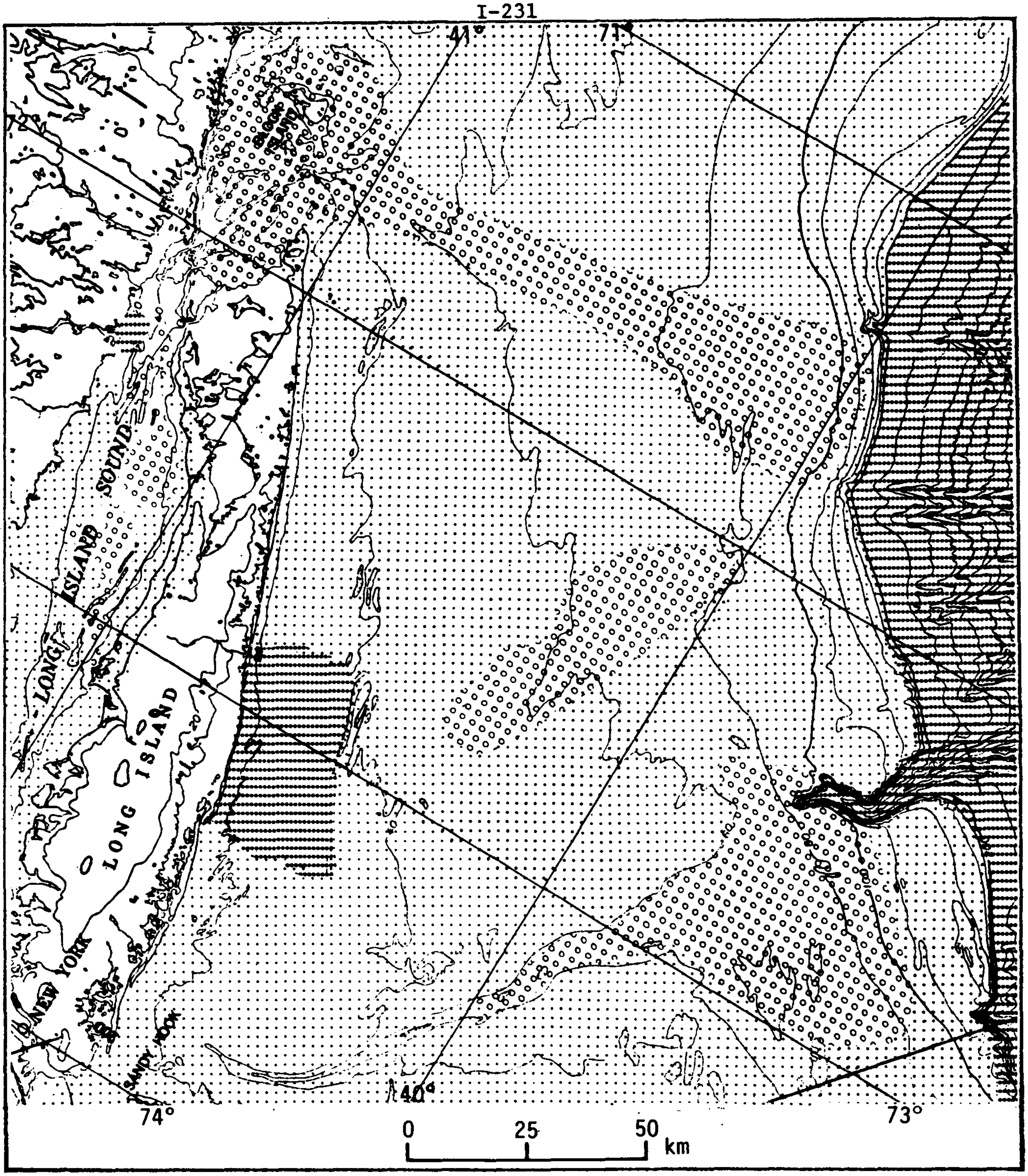


Fig. I-115 Contribution of literature for making an assessment of the amount of pre-transgressive subaerial surface preserved on the Long Island Continental Shelf. Sources used to compile this figure are: Grim and others (1970); Kraft and others (1968); McKinney and Friedman (1970); McMaster and Ashraf (1973a, 1973b, 1973c); Pratt and Schlee (1969); Sanders and Kumar (1975a, 1975b); Schlee (1973); Swift (1977); Swift and Sears (1974); Swift and others (1972); Tagg and Uchupi (1967).

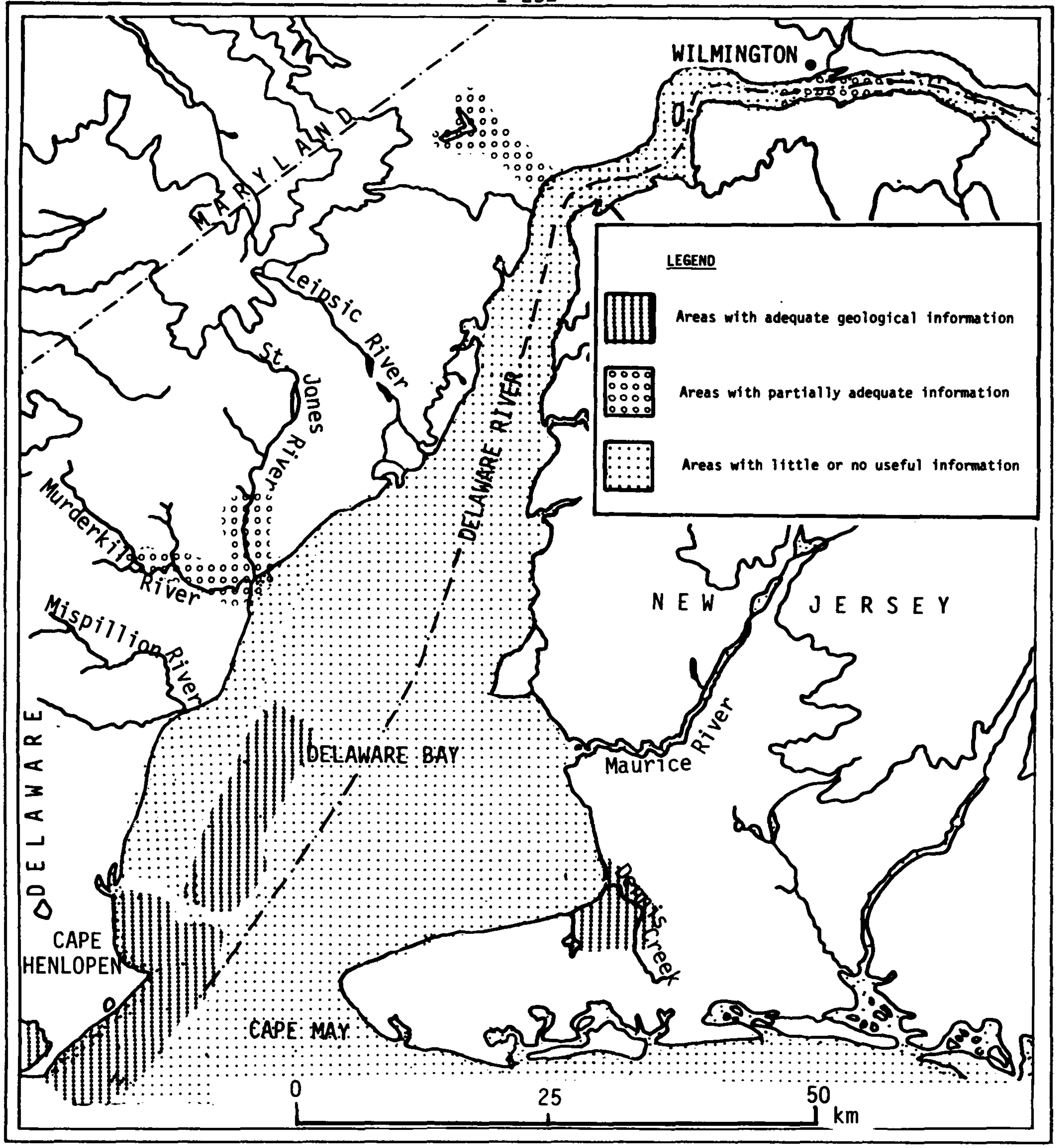


Fig. I-116

Contribution of literature for making an assessment of the amount of pre-transgressive subaerial surface preserved in Delaware Bay. Sources used to compile this figure are Belknap and Kraft (1977); Kraft (1971, 1977); Kraft and others (1974, 1978); Meyerson (1972); Sheridan and others (1974).

fication is based on published information and one to denote that it has been assigned on indirect or purely hypothetical grounds. As the figures illustrate, very few investigations of the CS have been published in sufficient detail to allow a determination of whether or not the subaerial surface is intact. Consequently, most of the CS has been mapped on the basis of indirect evidence, using hypothetical preservation models and shelf bathymetry as aids.

Applying preservation classification to the CS is not without its share of problems. For example, the Inner Shelf along the Delaware coast south of Cape Henlopen has been investigated in detail (Kraft 1971, 1977; Kraft and Maurmeyer 1978; Sheridan and others 1974), and these investigations have made possible the identification of several incised stream valleys (see Fig. I-43 in this report) and several truncated interfluvial headlands. Consequently, using such detailed data, it was possible to determine preservation over intervals measured in fractions of a kilometer. If no information had been available, this region of the Delaware coast would have been assigned to the partial preservation class. Instead, it has been mapped in greater detail allowing the area to be separated into two basic components: considerable areal preservation and negligible areal preservation of the subaerial surface. As this example illustrates partial preservation areas may be reclassified into subregions containing considerable preservation and negligible preservation. Along the flanks of major rivers, preservation will follow linear patterns conforming closely to the drainage pattern. On the CS, most drainage patterns were probably dendritic, trellis, or a combination since they would have formed on unconsolidated sands and silts during the Late Pleistocene regression between 35,000 and 18,000 B.P.

In addition to the figures showing the preservation class assigned to sections of the CS, figures have been included to show the distribution of information useful in determining subaerial surface preservation on the Shelf. These index maps may be used to evaluate the assigned preservation class when new data become available in the future.

The observed or predicted amount of preservation of the pre-transgressive subaerial surface on the northern North Carolina-southeastern Virginia Shelf is shown in Fig. I-104. As this figure illustrates, erosional shoreface retreat has probably removed much of the pre-transgressive subaerial surface outside of the major river valleys. But within these valleys marsh and estuarine silts have buried the flood plain, protecting the subaerial surface from erosion. In Pamlico Sound considerable preservation of the subaerial surface probably exists beneath lagoon and marsh sediments. Seaward of the barriers forming Pamlico Sound erosional shoreface retreat has destroyed most of the surface except in the major river valleys.

Fig. I-106 shows the predicted or observed class of preservation of the subaerial surface along the Delmarva Shelf. As this figure indicates, the subaerial surface is considered to be preserved more frequently along this portion of the shelf than along the shelf south of it.

Besides considerable preservation along the two major rivers at either end of this shelf compartment, it is hypothesized on the basis of information provided by Field and Duane (1976) that considerable preservation of the subaerial surface also exists along the Inner Shelf.

Preservation of the subaerial surface along the New Jersey Shelf has been mapped in Fig. I-108. Based on the investigations of Stahl and others (1974), Stubblefield and others (1975), and Stubblefield and Swift (1976), considerable areal preservation of the pre-transgressive subaerial surface has been predicted for the Inner and Middle Shelf regions. The upland area along the Middle Shelf and flanking the south side of the Hudson Shelf Valley has been assigned to the partial and negligible preservation classes. The area has been hypothetically placed in these classes because it is elevated above the adjacent shelf regions. Some topographical construction is the result of littoral drift sediment deposition, creating a shelf high referred to as a shoal-retreat massif (Swift 1973). Shoal-retreat massifs form in the nearshore region and are eroded and reshaped as sea level rises and the shoreline moves further inland (Swift 1973; Swift 1976b; Swift and others 1972; Swift and Sears 1974). Before a massif may begin to form in an area, erosional shoreface retreat has already occurred, but usually to a lesser extent than along non-massif regions. In the southern Hudson highland area, erosional shoreface retreat would have migrated across a relatively flat plateau-like plain covering over 1,200 sq km. Because the area was relatively flat and slightly above the surrounding shelf areas, significant marsh and lagoon deposits did not form and in general the pre-transgressive subaerial surface was not protected by a thick covering against erosional shoreface retreat. The area gradually evolved into islands and eventually into shoals as transgression progressed.

The predicted amount of subaerial surface preserved on the Long Island Shelf is shown in Fig. I-110. As this figure illustrates, considerable preservation is predicted for the major shelf valleys (the Block Valley, Hudson Valley, and Long Island Valley). The remainder of this Shelf can probably be classified under partial preservation, except for several cuesta-like features next to the Hudson and Long Island Valleys.

The predicted amount of subaerial surface preserved in the southeastern New England Shelf is shown in Fig. I-113. Minimal preservation of the pre-transgressive subaerial surface is predicted for Nantucket Shoals. It is believed that considerable erosion and transport of pre-transgressive deposits have taken place. The distribution of sand ridges and waves in this area indicates that erosional processes have been at work. Field work in this area, focusing on the amount of subaerial surface preserved, would be extremely useful, however, and should be done to check the predicted values.

The remaining portion of the southeastern New England Shelf is predicted to have partial preservation of the subaerial surface except along several probable drainage paths from Narragansett and Buzzards Bays and along the Outer Shelf. No previously performed surveys of the Holocene sediments were of use for determining the amount of subaerial surface preserved along this section of the CS. Consequently, there is no index figure for this area.

Index figures showing investigated areas used to determine the preservation of the subaerial surface along Georges Bank and the Gulf of Maine also have been omitted. In these subregions, the available areal investigations do not provide specific information on the distribution of Late Pleistocene-Holocene subaerial surface (see for example Garrison and McMaster 1966; Hoskins and Knott 1961; Knott and Hoskins 1968; McMaster and Ashraf 1973a, 1973b, 1973c; McMaster and others 1968; Oldale and others 1973; Schlee 1973).

The amount of pre-transgressive subaerial surface preserved on Georges Bank is difficult to assess given the absence of studies dealing with deposits immediately beneath the "surficial sand sheet" and sand ridges. At this time, it is predicted that the central and western flanks of Georges Bank display negligible preservation of the subaerial surface. Post-transgressive wave, tidal, and storm-current erosion have probably had considerable net effect on this portion of the Bank. The east side of Georges Bank, on the other hand, has probably received some sediment transported from the west and this material may have helped to bury surfaces and protect them from destruction.

The amount of areal preservation of the pre-transgressive subaerial surface in the Gulf of Maine is shown in Figs. I-115 and I-116. As these figures illustrate, preservation is restricted to a narrow strip along the present coastline. Further seaward in the Gulf of Maine, Holocene-Late Pleistocene subaerial surfaces were never formed because marine transgression was coincident with deglaciation. But above the 60-m bathymetric contour, the Shelf was exposed to subaerial forces during the Holocene. Preservation of the subaerial surface in this narrow strip is mainly confined to Late Pleistocene river and stream valleys where flood plain and marsh deposits accumulated.

Besides the open areas of the CS, five large bays or two sounds exist between Cape Hatteras, North Carolina and Canada. Preservation of the subaerial surface within each of these is assessed as considerable except for some central sections of Long Island Sound. Along the center of Long Island Sound, some regions may never have been subaerial because of enclosed depressions and poor drainage. Delaware Bay, Chesapeake Bay, Albemarle Sound, and Pamlico Sound all hold a rather high likelihood for intact pre-transgressive subaerial deposits. The center of each basin has acted as a sediment sink throughout the Holocene, burying older deposits under dozens of meters of material. Consequently, each area would be classified as displaying considerable preservation of the pre-transgressive subaerial surface.

In summary, Figs. I-104 to I-116 have shown the observed or predicted amount of subaerial surface preserved on the CS between Cape Hatteras, North Carolina and Canada. Information on sea level and glaciation has allowed us to designate a significant portion of the Outer Shelf and Continental slope as having low archaeological potential. For the remaining Shelf, high preservation values for the pre-transgressive subaerial surface generally correlate with major valleys and sometimes with barrier-protected shorelines. Transgressed headlands and upland regions seem to provide the least protection for the subaerial surface.

This assessment of the CS's potential for containing intact subaerial surfaces is based on indirect evidence and models of transgression in many areas of the Shelf. It would greatly aid cultural resource management if the studies were designed specifically to test and refine some of the data and concepts presented here.

16.0 PROBLEMS, DATA GAPS, AND RECOMMENDATIONS

This study of the geology of the Continental Shelf encountere two major obstacles which partially impeded our reconstruction of the Late Pleistocene-Holocene subaerial CS. The first was a lack of precise local sea-level data, especially for the period before about 8000 B.P. The second was the gap in our knowledge of the fate of the subaerial surface during marine transgression. In the future, it would be of immense importance if buried deposits in those areas considered to contain preserved subaerial surfaces were investigated in order to shed more light on past sea-level positions and on the integrity of the buried subaerial surface.

At present, little is known about the portions of shorelines during most all of the Holocene. Rates of erosional shoreface retreat and sea-level positions need to be reconstructed before one can accurately locate the position of a former shoreline in a given period. Knowing the level of the ocean at a specific period in the past and its equivalent bathymetric contour would not produce an accurate reconstruction of the shoreline because it neglects the net effect of erosional shoreface retreat. These problems could be overcome through a rigorous program of coring and sampling carefully selected in situ material for radiocarbon dating. At present, much of the controversy over shoreline positions and sea level arises from the use of potentially mobile or inadequate samples to construct sea-level curves. If shell material is used for radiocarbon dating, it is strongly advisable to use the interior of the shell and not the total shell unless the specimen is absolutely unweathered.

The use of a systematic coring program along the former retreat paths of large estuaries on the Middle and Outer Shelves should provide the greatly needed data for sea level positions before 8000 B.P. The required material for radiocarbon dating should be obtainable from peat and shell buried in the estuary and in the marsh sediments found along buried valleys.

The collection of data for reconstructing Middle and Early Holocene sea levels should help to refine the shorelines shown on Chart I-1a. These data should also indicate that stillstand periods do not mean the shoreline remained absolutely fixed, but rather that the level of the ocean remained the same. In other words, during stillstands, erosional shoreface retreat continued to move the shoreline landward along certain portions of the coast,

An understanding of the fate of the pre-transgressive subaerial surface during erosional shoreface retreat and sea-level rise would be greatly enhanced if several systematic areal investigations were done along the

Inner, Middle, and Outer Shelves. To date, few geologists have tried to identify the buried subaerial surface in Shelf cores which have penetrated the "surficial sand sheet" and estuary or lagoon sediments. In the future, it may prove to be frequently impossible to identify the buried subaerial surface in cores, even if it does exist intact. If this becomes a problem it is important to recognize it as soon as possible. The work done by Kraft (1971, 1977) offers some insight into problems associated with the identification of truncated pre-transgressive deposits. But in order to understand the archaeological potential of these buried deposits, we need additional information regarding the amount of disruption and erosion which generally precedes the burial of a subaerial surface during transgression.

Before any new investigations are initiated on the CS, it would be advisable to look at some of the data already in existence, such as the cores and seismic profiles for some regions of the CS (see for example Cousins and others 1977). It was not possible during this project to review raw data and conduct analyses on existing cores. In the future, it may prove to be more cost-efficient to use some of the data collected by major marine science institutions than to collect new data. A review of the data at Woods Hole Oceanographic Institute, the Virginia Institute of Marine Sciences, and similar agencies should be included in future plans. After the raw data from these agencies have been investigated, a more accurate assessment of data gaps would be forthcoming. Then, instead of a data gap, it might be more realistic to talk about a gap in data assimilation.

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GLOSSARY

ACCOUSTICAL RETURN. The return of a seismic signal back to its original location after being "bounced" off a layer of sediment or rock.

ARCUATE. Curved or bowed.

BARRIER. A low sandy landform either attached (bay mouth barrier or barrier spit) or unattached (barrier island) to the coast.

BARRIER ISLAND. A low, sandy island separated from the mainland coast by either a bay or lagoon.

BATHYMETRY. A measurement of the depth of water in oceans, seas and lakes.

BAY MOUTH BARRIER. A barrier extending across the mouth of a bay.

BEACH ACCRETION PLAIN. A region where beaches have prograding as a result of nearshore processes and sediment deposition to produce a low sandy plain marked by slight ridges.

BED LOAD. Material such as sand and gravel being moved by river or marine currents.

BERM. Flat above-water feature that forms the beach at the upper end of the shoreface arc.

BIGHT. An inward or concave bend in a coastal configuration.

BEFORE PRESENT (B.P.). Used to indicate number of years before the "present," with A.D. 1950 used as a beginning date. For example, 2000 B.P. is equivalent to 50 B.C.

CATCHMENT BASIN. Drainage basin, i.e., an area occupied by a drainage system.

COASTWISE SPIT PROGRADATION. Lateral growth and development of a spit nourished by littoral drift.

CONCRETIONARY. Pertaining to properties of a concretion, an accumulation of mineral matter that forms around a center or axis of deposition. It is commonly spherical or disk-shaped and composed of cementing materials such as calcite, dolomite, iron oxide, or silica.

CONTINENTAL SLOPE. Portion of the ocean floor extending from the seaward edge of the continental shelf to the ocean deeps. The upper boundary of the continental shlope is defined by the shelf break, which varies from 50 to 150 m in depth between Cape Hatteras, North Carolina and Canada.

CS. Abbreviation used in this text for continental shelf.

CUSPATE FORELAND. Sector of coastline projecting seaward to form a pointed headland.

DEGLACIATION. Melting and retreat of glacial ice accompanying a major climatic warming trend.

DELTAIC-FLUVIAL SEDIMENTS. Sediments deposited by a river along a river valley or where the river flows into a body of standing water.

DISTRIBUTARY CHANNEL. A river branch that flows away from a main channel and does not rejoin it (characteristic of deltas).

DRAINAGE BASIN. The area from which a given stream and its tributaries receive their water.

EMBAYMENTS. Waters partially sheltered from the open ocean by coastal landforms.

END MORAINE. A ridge or belt of till which marks the farthest advance of a glacier.

ENTRENCHED. Describing a stream or river channel which has cut into the adjacent plain as a result of lowering sea-level.

ENVELOPE OF EROSION. An arc encompassing the section of the shoreface undergoing active erosion and redistribution.

EROSIONAL SHOREFACE RETREAT. The erosion and redistribution of coastal, nearshore, and marine deposits by submarine and nearshore hydraulic processes.

ESTUARINE. Pertaining to an estuary.

ESTUARY. A body of water freely connected to the ocean and in which fresh water is measurably diluted by salt water.

EURYHALINE. Tolerant of a wide range of salinity values in water.

EUSTATIC. Pertaining to the world-wide level of the oceans.

EUSTATIC CURVES. Curves or graphs depicting successive sea-levels over time.

FACIES. An accumulation of sediment that exhibits specific characteristics which differ from the characteristics of an adjacent deposite formed contemporaneously and associated laterally.

FALL LINE. A zone characterized by numerous waterfalls and rapids.

FINE(S). Fine sediment. The term is generally used to refer to silt- and clay-sized particles.

FLOOD PLAIN. An area of low relief bordering a stream or river over which water spreads during flooding.

FLUVIAL. Pertaining to a river or stream.

FORAMINIFERA. Very small marine organisms whose concentrated shells form chalf and varieties of limestone.

FOREBULGE. Region uplifted ahead of a glacial front due to dynamics associated with crustal movement and isostacy.

GEOID. An imaginary surface coinciding with the mean sea-level and extending through the continents.

GLACIATION. Pertaining to a period when colder climates prevailed and glaciers covered a large portion of the temperate zones.

GLACIO-LACUSTRINE DEPOSITS. Deposits formed by glaciers or lakes.

HOLOCENE. A periods characterized by world-wide warming starting after the last glaciation and continuing to the present (from about 15,000 B.P. to the present).

HUMIC FRACTION. A mixture of dark-colored organic substances found in a soil horizon and formed by decomposition of organic matter.

HUMMOCKY. Terrain composed of numerous rounded knolls or small hills.

HYPSITHERMAL. A period during the Holocene characterized by world-wide climates warmer than the present.

INFLECTION ZONE. A bend or change in direction with respect to a straight line.

INNER CONTINENTAL SHELF. The portion of the continental shelf lying at a depth of less than one third the depth of the adjacent shelf break.

INTERFLUVE. The area between adjacent rivers or streams flowing in the same general direction.

INTERSTADIAL. A period of warming and glacial stagnation during a major period of glaciation.

INTERTIDAL. Pertaining to the littoral zone above the low tide mark.

ISOSTACY. The ideal condition of balance attained between earth materials of differing densities if gravity were the only force governing their heights relative to each other.

ISOSTATIC. Pertaining to isostacy.

LAG GRAVEL. Accumulation of coarser particles left behind after finer particles have been carried away by wind or water currents.

LAGOONAL MUDS. Silt and mud deposited in a lagoon.

LAGOONAL SEQUENCE. Fine-grained material such as silts and fine sands deposited in a lagoon.

LITHIFY. To convert unconsolidated sediments into their consolidated or rock counterparts, such as the formation of sandstone from sand cemented with calcium carbonate under pressure.

LITTORAL DRIFT. Movement of gravel, sand and other material along a coast induced by waves and currents.

LOBATE. Pertaining to a curved or somewhat rounded projection.

LOW-STAND. Position of the coastline during a glacial maximum at which greatest emergence of the continental shelf takes place.

MARINE TRANSGRESSION. The advance of the sea over coastal land areas due to either sea-level rise or erosional shoreface retreat.

MASSIF. A large submarine feature consisting of positive relief and formed from sediment deposited by currents and littoral drift converging on a depositional center.

MEANDER PLAINS. A low plain along a valley floor where river meanders occur frequently.

MID-CONTINENTAL SHELF. The section of the continental shelf lying between the inner and outer continental shelves.

MORAINE. A general term applied to certain glacial landforms composed of till.

NON-FOSSILIFEROUS. Containing no evidence or remains of past forms of life.

OOLITE. Spherical grains of sand-sized calcium carbonate (usually) considered to have formed by inorganic precipitation.

OOLITIC ROCK. A rock comprised mainly of oolites.

OUTER CONTINENTAL SHELF. That portion of the continental shelf lying in between the shelf break and a line corresponding to two thirds of the depth found at the adjacent shelf break. In this study, the outer continental shelf also includes a portion of the upper section of the continental slope.

OUTWASH. Material carried from a glacier by meltwater and laid down in stratified deposits.

OUTWASH PLAIN. Flat or gently sloping surface underlain by outwash deposits.

PEAT. The residual product of partially decomposed plants accumulating in wet environments, such as a marsh, and partially compressed.

PIEDMONT. Plains spreading adjacent to mountains.

PLEISTOCENE. An epoch in geological time forming the earliest section of the Quarternary and spanning from about two million years ago up to about 15,000 B.P. It is a time characterized by several periods of world-wide glaciation.

PRE-TRANSGRESSIVE. Pertaining to a period before marine transgression.

PROGRADING SHORELINE. A shoreline which is advancing seaward due to the amount of sediment being deposited from rivers or long shore currents.

PROVENIENCE. Location in 3-dimensional space of an item or artifact.

QUATERNARY. A portion of geologic time consisting of both the Pleistocene and Holocene.

RADIOCARBON. Radioactive isotope of carbon with a half life of about 5,720 years and used to date events back to about 50,000 years ago.

RECURVED SPIT TIPS. Spit tips which curve strongly inward.

REFLECTOR. A surface or horizon which reflects seismic waves.

REGRESSIVE DEPOSITS. Deposits laid down during marine regression.

RELICT. Inactive and consequently pertaining to an earlier period of formation.

RELIEF. Elevations of a land surface.

SCARP. Steep slope or cliff like landform.

SCOUR. Erosion continuing below a former surface or horizon and forming a depression.

SEDIMENTOLOGICAL. Pertaining to the study of sediments.

SHALLOW STRUCTURE. Geologic structure found close to the surface of the earth.

SHOALING. Becoming shallower or less deep.

SHORELINE MIGRATION. Movement of the shoreline seaward or landward of its former position.

SLOPEWASH. Soil and rock material moved down a slope by the force of gravity and running water.

SPIT. A sandy bar formed by currents and extending into a bay from a promontory or headland.

STILLSTAND. A period during which the level of the ocean remains stable.

SUBAERIAL. Pertaining to land surfaces covered by air as opposed to water.

SUBMARINE HYDRAULIC PROCESSES. Processes consisting of the natural movement of water in the ocean.

SUBSIDENCE. Sinking or lowering of a surface relative to a fixed plane of reference.

SURF ZONE. The area between the seaward limit of breaking waves and the upper beach face.

TECTONICALLY. Pertaining to movement of land surfaces.

TERMINAL MORAINE. A ridge or belt of till marking the farthest advance of a glacier.

TERTIARY. The earliest portion of the Cenozoic period starting about 63,000,000 years ago and ending with the start of the Pleistocene epoch about 2,000,000 years ago.

THALWEG. A line defining the lowest point along the longitudinal axis of a valley.

TIDAL RANGE. Distance in elevation between mean low and mean high tide.

TILL. Unstratified and unsorted glacial drift deposited directly by glacial ice.

TRELLIS DRAINAGE. A drainage pattern composed of roughly rectilinear arrangements of stream courses.

TRUNCATED. Pertaining to a bevelled, cropped, or cut-off apex or upper section of a feature.

UNCONFORMITY. A buried erosion surface separating two units of sediment.

VARVED CLAY. A pair of thin sedimentary beds containing clay and coarser sediments (silts) usually grading from coarse to fine and interpreted to represent a cycle of one year.

WAVE FETCH. The distance over water in which wind is able to build up waves and drive them landward.

WELL LOGS. Subsurface geological information collected by drilling.

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