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Drainage Displacement by Sea-Level Fluctuation at the Outer Margin of the Chesapeake Seaway

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ABSTRACT

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The trace of the ancestral Susquehanna River across the modern coastline has been displaced over 40 km southward since the middle Pleistocene. High-resolution seismic reflection data from the inner shelf suggest that submerged-channel fill is responsible for at least three "time-lagged" channel shifting events during regressive parts of Pleistocene eustatic cycles.

Traditional models for tidal-inlet shift are dependent on migrating spit platforms that progressively fill a portion of the updrift side of a channel and directly force excavation of an equivalent portion of material from the downdrift side. Channel migration results in a continuous broad channel-scar representing the integrated positions of channels at time scales of 10^0 to 10^2 years.

Seismic data adjacent to the southern Delmarva Peninsula illustrate four separate Pleistocene lowstand pathways for the ancestral Susquehanna River. Although regional spits have migrated between these pathways, there is no evidence of continuous channel migration between the channel traces. The mechanism for channel shifting was glacio-eustatically controlled, and occurred at time intervals of 10^5 years. Channel shifting was operative during early regression when fluvial channels were not confined to their previous antecedent lowstand channels and jumped laterally to new locations.

Time-lagged channel shifts are indirectly forced by spit growth during major highstand events and widespread regional flooding. During these events, the ancestral Susquehanna River was no longer confined to its fluvial valley but spread into broad ancestral Chesapeake Bays formed in the drowned Chesapeake Basin. Estuarine sediments, migrating spit platforms, and bay-entrance shoals filled the drowned fluvial channels in the seaward parts of the ancestral bays. During subsequent regressions, the bays shrunk and the flow of the ancestral Susquehanna River jumped to adjacent tidal-flushed pathways and subsequently drained to the shelf edge.

During late Pliocene and early Pleistocene lowstands, at least six major rivers drained across the Chesapeake Basin. Time-lagged channel shifts and stream captures have progressively reduced the number of drainways leaving the basin. During the most recent Stage 2 lowstand, only two or possibly three drainways may have crossed the entrance of the Chesapeake Basin.

ADDITIONAL INDEX WORDS: *Paleochannels, Quaternary stratigraphy, seismic stratigraphy, Chesapeake Bay, Susquehanna River, channel migration.*

INTRODUCTION

The Chesapeake Basin

The Chesapeake Basin is an informal geographic name given to a broad depression associated with the modern Chesapeake Bay and the lower reaches of the watersheds that drain into the bay (Figure 1). The modern Susquehanna River enters the basin at the northern end, has the largest watershed of the six rivers emptying into the basin (≈ 70,000 km^2; MIXON, 1985), and supplies the greatest volume of river runoff to the system. Major rivers draining into the western side of the basin include the Patuxent, Potomac, Rappahannock, York and James. The eastern side of the basin is fed by smaller watersheds on the Delmarva Peninsula. The drainage divide on the Delmarva Peninsula is offset to the northeast side of the peninsula. Consequently, over 70% of the surface drainage is southwestward toward the axis of the Chesapeake Basin (see OERTEL and KRAFT, 1994). The major rivers draining southwestward from the Delmarva Peninsula are the Chester, Choptank, Nanticoke, Wicomico and the Pocomoke.

During the Pliocene, an early Chesapeake Basin was located in the Salisbury Embayment (WARD, 1985) of the subsiding Baltimore Canyon Trough. The northeastern boundary of the basin was formed by deltas of the ancestral Susquehanna and Delaware Rivers (Figure 1; RAMSEY, 1992). Deltaic lobes of these rivers spread across the

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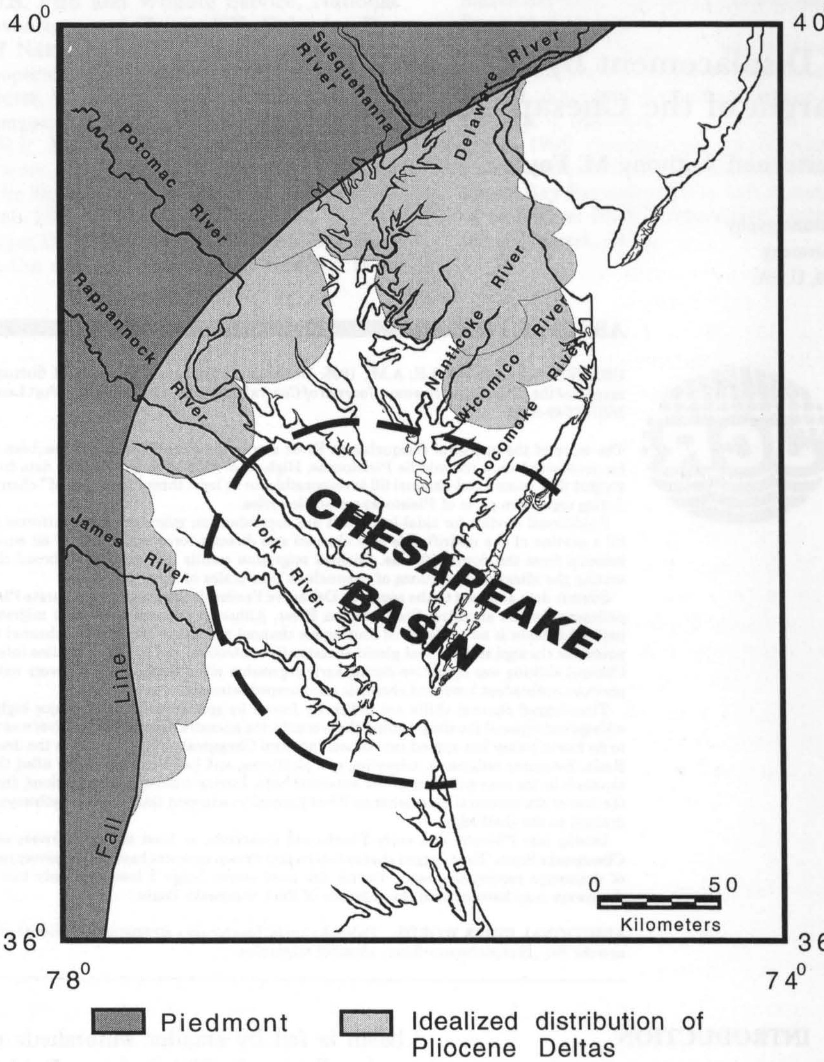


Figure 1. Map of the middle Atlantic region illustrating the location of the late Pliocene to early Pleistocene Chesapeake Basin. The Chesapeake Basin is located within the broader confines of the Cenozoic/Mesozoic Salisbury Embayment. The boundaries of the Chesapeake Basin are defined by the Fall Line to the west, and by Pliocene deltaic depocenters (idealized in figure) to the northeast and southwest.

northern part of the Salisbury Embayment and extended into Delaware and southern Maryland forming the core of the northern Delmarva Peninsula (RAMSEY, 1992). MIXON (1985) suggested that the Tunnels Mill Member of the Yorktown Formation may have been the pro-delta and delta-front deposits of these highstand facies. The southern margin of the basin was offset further to the west and formed by Pliocene fluvial deltaic systems southwest of the James River. During several Plio-Pleistocene highstand events, the Chesapeake Basin was repeatedly flooded forming broad embayments that had relatively wide openings to Pliocene and Pleistocene seas.

Although highstand deltaic deposits and regional uplift helped define the margins of the early Chesapeake Basin, the present basin is mainly an erosional feature (MIXON, 1985). During the Pleistocene, transgressions deposited coastal and shallow-marine deposits that were subaerially altered during intervening regressive events. Lowstand exposures of the Chesapeake Basin during the Pliocene and Pleistocene caused reduction in the elevation of the basin surface through the maturation and development of the numerous watersheds.

The Pleistocene evolution of the southern Delmarva Peninsula illustrates how alternating glacio-eustatically controlled transgressions and regressions affected stratigraphy and drainage evolution within the basin. The southern Delmarva Peninsula is the narrow southern part of the peninsula primarily located south of the Maryland-Virginia state line. MIXON (1985) described how highstand erosion of deltaic/marine deposits (possibly Beaverdam Formation?; GROOT *et al.*, 1990) at the northern part of the peninsula resulted in southward spit growth that built the southern part of the peninsula. Mixon's findings were based on an extensive set of drill logs from the Delmarva Peninsula. High-resolution seismic data from the Chesapeake Bay allowed COLMAN and HOBBS (1987), COLMAN *et al.* (1988), COLMAN and HALKA (1989), and COLMAN *et al.* (1990) to reconstruct Pleistocene fluvial drainage patterns beneath the Chesapeake Bay floor. MIXON (1985) and COLMAN and MIXON (1988) indicated that the highstand spits which formed the core of the southern Delmarva Peninsula may have altered the courses of the Susquehanna River during subsequent lowstands.

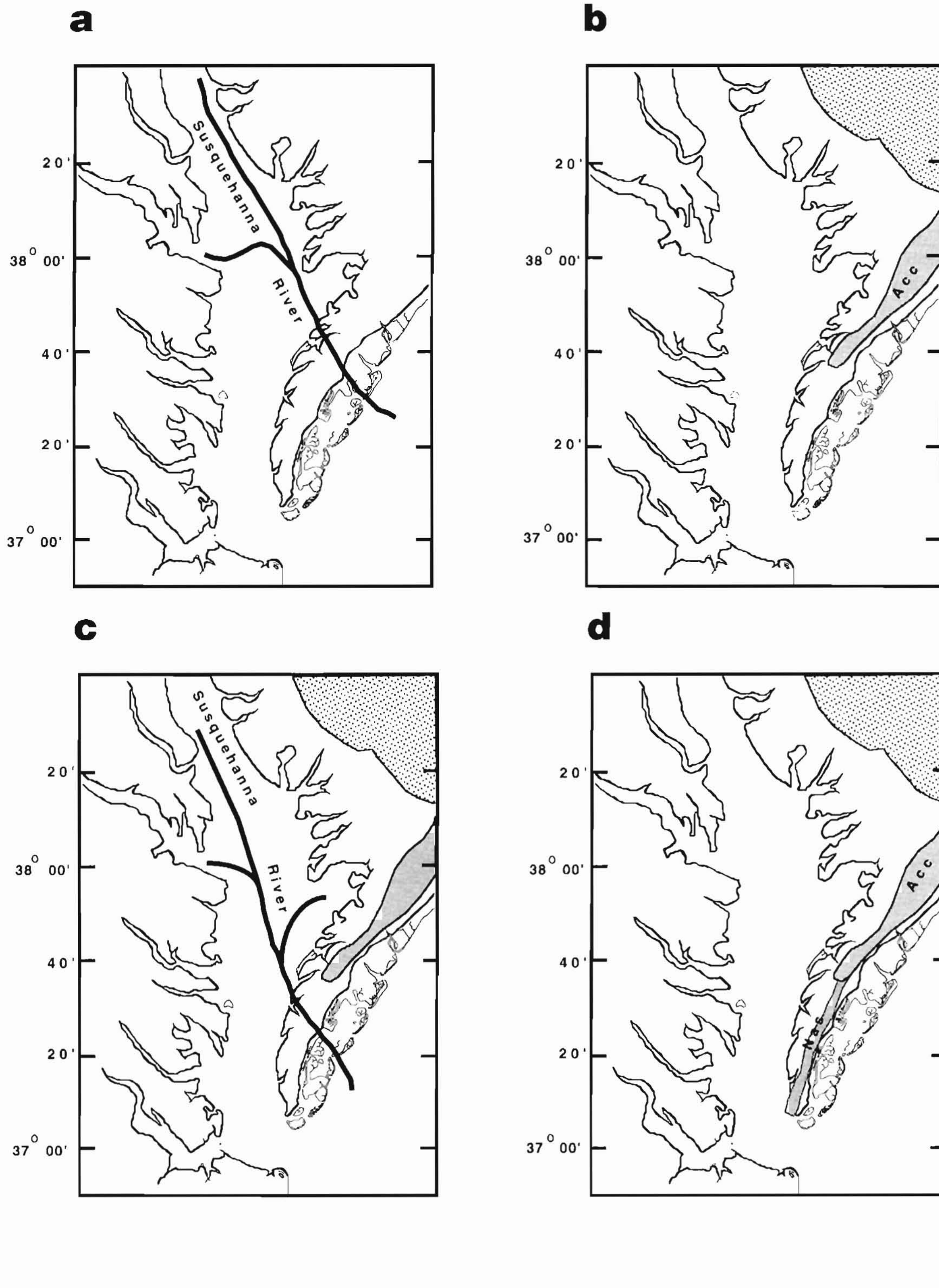
Susquehanna Drainage History

COLMAN and MIXON (1988) provided the ground work for our investigation by tracking buried fluvial channels beneath the axis of the Chesapeake Bay and correlating them with buried channels interpreted from drill logs of the southern Delmarva Peninsula. They suggested that traces of three buried channel systems correlated with two buried channels under the southern Delmarva Peninsula and with one partially filled channel beneath the northern side of the Chesapeake Bay mouth (Figure 2). The channels were all believed to be ancestral channels of the Susquehanna River that had been displaced southward during successive Pleistocene glacio-eustatic cycles. In the COLMAN and MIXON (1988) model, the Exmore paleochannel was the northernmost channel incised during Stage 8 or 12 (Figure 2a). During highstand Stage 7 or 11, the channel was subsequently capped by the "Accomack spit" (Figure 2b). This left the Stage 6 lowstand drainage of the Susquehanna to be diverted southward around the distal end of the Accomack spit forming the Eastville paleochannel (Figure 2c). The following Stage 5e highstand produced the Nassawadox spit (MIXON, 1985) which forced subsequent lowstand drainage (Stage 2) of the Susquehanna River to flow through the Cape Charles paleochannel at the northern end of the modern Chesapeake Bay mouth (Figure 2d and 2e). During the late Holocene, spit growth at the southern tip of the southern Delmarva Peninsula (Figure 2f) is believed to have displaced the modern tidal-flushed Chesapeake Channel from its original location above the axis of the Cape Charles paleochannel to a location 12 km southward (COLMAN and MIXON, 1988; COLMAN *et al.*, 1988; COLMAN *et al.*, 1992). This latter mechanism of tidal-channel displacement within a flooding paleovalley is similar to traditional mechanisms of continuous spit forcing formerly described by GILBERT (1885) and SHEPARD (1960).

Mechanics of Channel Shift by Spit Forcing

The COLMAN and MIXON (1988) and COLMAN *et al.* (1988) model for the southward shift of the Susquehanna River paleovalleys during Pleistocene highstand events is dependent on the southward progradation of the Accomack and Nassawadox spits. They believed that the southward progradation of the barrier-spit system was the result of nearshore sand transport in a strong southward littoral system that flowed along the east coast of the Delmarva Peninsula (COLMAN *et al.*, 1988). COLMAN and MIXON (1988) and COLMAN *et al.* (1988) also suggested that during the present highstand, a strong littoral drift system caused the upper tidal part of the Cape Charles paleochannel to be displaced 12 km south to the modern Chesapeake Channel. During the two previous highstands, channel shifting by spit forcing was not an important mechanism.

The "traditional" mechanism of inlet-channel shift by prograding spits (at time scales of 10^0–10^2 years) has been known for over a century (GILBERT, 1885). SHEPARD (1960) illustrated how longshore sediment drift caused the accumulation of sediment on the updrift side of an inlet which produced spits that prograded into the inlet. Large-scale spit progradation is dependent on high rates of longshore sediment transport that are rapidly reduced at zones of decreased transport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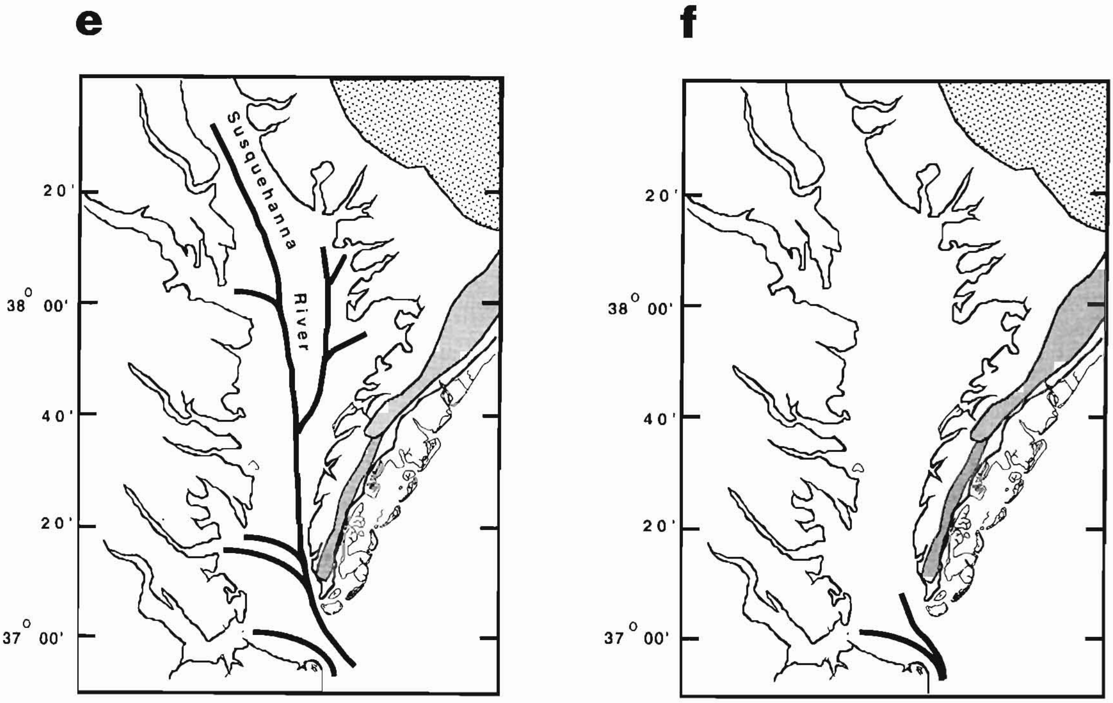

Figure 2. COLMAN-MIXON MODEL for the geologic evolution of the southern Delmarva Peninsula and migration of the Susquehanna River (modified from MIXON, 1985; COLMAN and MIXON, 1988; COLMAN *et al.*, 1990). (a) Stage 12 lowstand drainage path of the Susquehanna River along the northern margin of the Chesapeake Bay and through the Exmore paleochannel in the southern Delmarva area. (b) Stage 11 highstand development of the Accomack spit across the northern part of the entrance to the Chesapeake seaway. (c) Stage 6 lowstand drainage path of the Susquehanna River along the central axis of the Chesapeake Basin and through the Eastville paleochannel in the southern Delmarva area. (d) Stage 5e highstand development of the Nassawadox spit across the central entrance area of the Chesapeake seaway. (e) Stage 4-2 lowstand drainage of the Susquehanna River through the Cape Charles paleochannel in the south central entrance area of the Chesapeake Basin. (f) Modern tidal drainage of the Chesapeake Bay.

competence, at estuarine embayments and coastal inlets. As spit platforms fill the updrift side of an inlet channel, the cross-sectional area of the inlet channel is initially reduced, but ultimately an equilibrium cross-sectional area is maintained as the inlet throat migrates downdrift (Figure 3). The cross-sectional trace of the filled channel is always wider than the actual width of the active inlet channel (HOYT and HENRY, 1967; MOSLOW and TYE, 1985; OERTEL *et al.*, 1991). The spit-forcing mechanism operates at short time scales (10^0–10^2 years) when fluctuations in sea level are relatively minor. However, it requires a relatively continuous and large supply of littoral sediment, often from major headlands.

MIXON (1985) and COLMAN and MIXON (1988) illustrated that the Accomack and Nassawadox spits overlie the Exmore and the Eastville paleo-

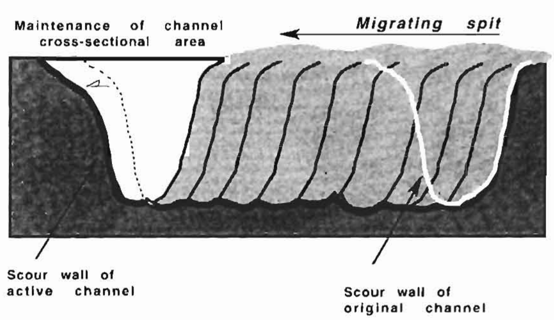

Figure 3. Spit-fill model for channel migration modified after GILBERT (1885) and SHEPARD (1960). Spit fill on the updrift side of a tidal or fluvial channel directly forces downdrift channel migration to maintain equilibrium cross-sectional area.

channels, respectively, beneath two separate areas of the southern Delmarva Peninsula (Figure 2). MIXON (1985) illustrated that the channels were primarily filled with fluvial and estuarine deposits and that paleospit sands were deposited above the filled paleochannels. Therefore, the COLMAN and MIXON (1988) stratigraphic model requires a mechanism for channel migration which produces fluvial-estuarine-spit channel fill prior to capping by spit facies.

Mechanics of Channel Shift by Coastal Inundation

The time-lagged mechanism for channel shifting operates at much larger time scales (10^5 years, during the Plio-Pleistocene), than the spit-forcing mechanism (10^0 to 10^2 years). The mechanics of shift are controlled by glacially induced sea-level fluctuation, not sediment supply *per se,* as in the traditional spit-forcing model.

Glacio-eustatic changes serve to drastically modify sediment supply and provenance and the timing of erosional and depositional processes. During lowstand, fluvial-channel entrenchment occurs in response to base-level lowering and channel infill is minimal to absent. During transgression, fluvial and estuarine sediments initially accumulate in the channel. Small-scale spits at estuary-entrance margins move sediment from interfluves into shoaling estuary entrances where tidal channels migrate or fill. When interfluve areas are flooded, tidal flow through estuary entrances is no longer confined to the antecedent fluvial channels and forcing by marginal spits ceases. If sediment from a major headland source (*i.e.*, an eroding deltaic headland) is available during highstand, then a regional spit may cap the fluvial-estuarine channel fill facies and force a much shallower tidal channel to migrate over the interfluve. MIXON (1985) has illustrated that this has occurred at the Eastville and Exmore paleochannels beneath the southern Delmarva Peninsula. COLMAN and MIXON (1988) used the spit clinoforms in the upper part of the Cape Charles paleovalley to illustrate recent tidal channel forcing at the Chesapeake Bay mouth. During subsequent regression, renewed fluvial entrenchment of the tidal channel provides a pathway for captured drainage.

The purpose of this research was to investigate the seaward side of the southern Delmarva Peninsula and determine how the seismic record of channel distribution and fill could be used to interpret mechanics of channel shift related to sedimentary and morphodynamic processes.

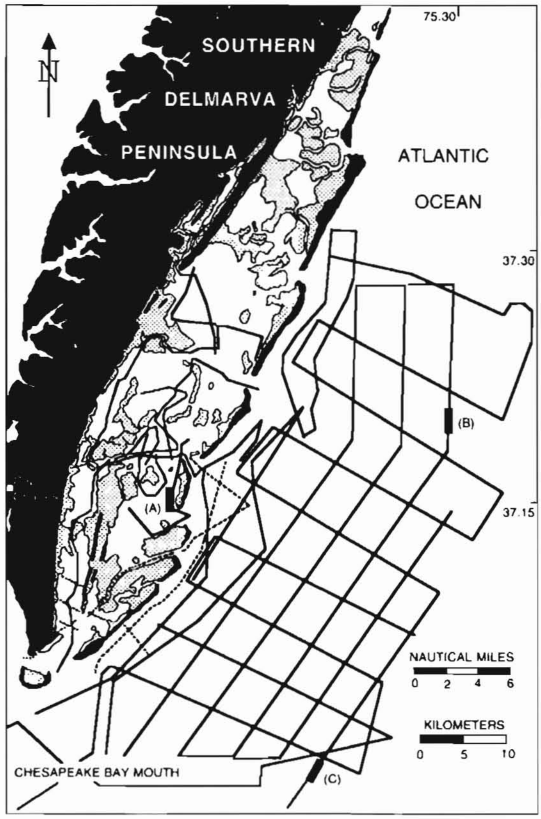

Figure 4. Location of geophysical survey lines on the shoreface and inner shelf seaward of the southern Delmarva Peninsula. Data collected by SHIDELER *et al.* (1984) are indicated by dashed lines. Lagoonal marshes indicated by stippled pattern. Barrier islands on the seaward side of the lagoons indicated by solid black pattern. Bars labelled A, B, and C refer to seismic reflection profiles shown in Figures 6, 7, and 8, respectively.

METHODS

The Quaternary record east of the Delmarva Peninsula was primarily developed from high-resolution, seismic-reflection profiles (Figure 4). The seismic data were collected using a Geopulse® boomer system generally run at 175 J, fired at 0.25 s intervals and band-pass filtered between 750 and 2000 Hz. Loran C with first-order corrections was used for navigation during the surveys. Approximately 1,000 kilometers of track were surveyed adjacent to the southern Delmarva Peninsula. In addition, approximately 90 km of sin-

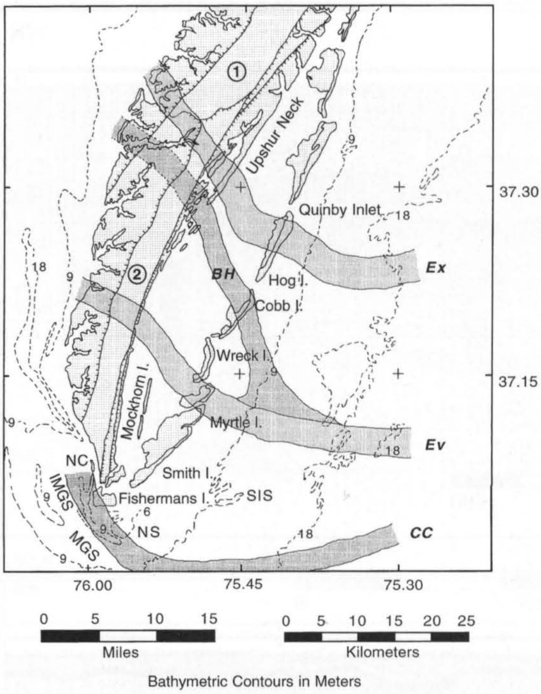

Figure 5. Map showing the locations of the Exmore (Ex), Belle Haven (BH), Eastville (Ev) and Cape Charles (CC) paleochannel tracts as identified in this study. Exmore and Easville paleochannel locations beneath the axis of the southern Delmarva Peninsula based on MIXON (1985) and COLMAN *et al.* (1990). Geographic features referred to in text are: Inner Middle Ground Shoal (IMGS), Middle Ground Shoal (MGS), North Channel (NC), Nautilus Shoal (NS). Accomack and Nassawadox highstand spits indicated by scarp-bounded units (1) and (2), respectively. Lagoonal marshes are not shown.

gle-channel seismic-reflection data collected by SHIDELER *et al.* (1984) were reinterpreted in this study. In the shore-normal direction, the study area extended from the landward margin of the coastal barrier lagoon (approximately 12 km wide), to approximately 26 km seaward of the outer coastline formed by the barrier islands. In the shore-parallel direction, the study area was from Middle Ground Shoal in the northern Chesapeake Bay entrance to Quinby Inlet at the north end of Hog Island, Virginia. Water depth above the inner continental shelf along the outer part of the study area reached approximately 20 m. In the laboratory, major reflectors were mapped over the study area and relative ages were determined by the relationships of truncated beds at unconformities.

RESULTS

The Quaternary stratigraphy beneath the floor of the coastal barrier lagoons and inner continental shelf illustrates two major filled-channel systems and associated buried interfluve surfaces. MIXON (1985) described a buried channel that crossed the southern Delmarva Peninsula near Eastville, Virginia, as the "Eastville paleochannel". Our seismic profiles illustrate a large buried paleochannel (P_{el}; Pleistocene Eastville lowstand unconformity) located seaward of the Eastville paleochannel which we believe to be the seaward extension of the Eastville drainage system. This portion of the Eastville paleochannel appears beneath the floor of the coastal barrier lagoon just north of the town of Oyster and can be traced in an east-southeast direction crossing the outer barrier coastline beneath Wreck and Myrtle Islands, and then out beneath the continental shelf (Figure 5; FOYLE and OERTEL, 1992). Under the coastal lagoon and inner continental shelf, the paleochannel trace becomes sinuous with four gentle meanders. Seaward of Myrtle Island, the paleochannel trace abruptly changes trend and heads eastward. The paleochannel has an average width of approximately 3 km and has an axial depth which ranges from 55 to 62 m (below present msl-datum). The relief of the paleochannel (between its relict interfluve surface and thalweg) averages 30 m. Reflectors within the paleochannel cross-section occasionally indicate a seismically noisy (high-amplitude and discontinuous reflections) lower fluvial unit (Seismic Facies I). Generally, however, an estuarine facies (Seismic Facies II; Figure 6) is present, overlain by an upper unit containing inclined reflectors indicating a prograding-fill sequence from south to north (Seismic Facies III; Figure 6). A major ravinement (P_{tr}; Pleistocene transgressive ravinement) lies directly above the fill sequence of the Eastville paleovalley and has been traced laterally throughout the study area.

MIXON (1985) also described a buried channel that crossed the southern Delmarva Peninsula further north near Exmore, Virginia, as the "Exmore paleochannel". Seismic profile data from our study indicate that the Exmore paleochannel appears below the floor of the coastal barrier lagoon near the northern part of Upshur Neck and extends in a southeast direction, crossing the outer barrier coastline beneath northern Hog Island (Figure 5). Under the seaward part of the coastal lagoon and inner continental shelf, the paleo-

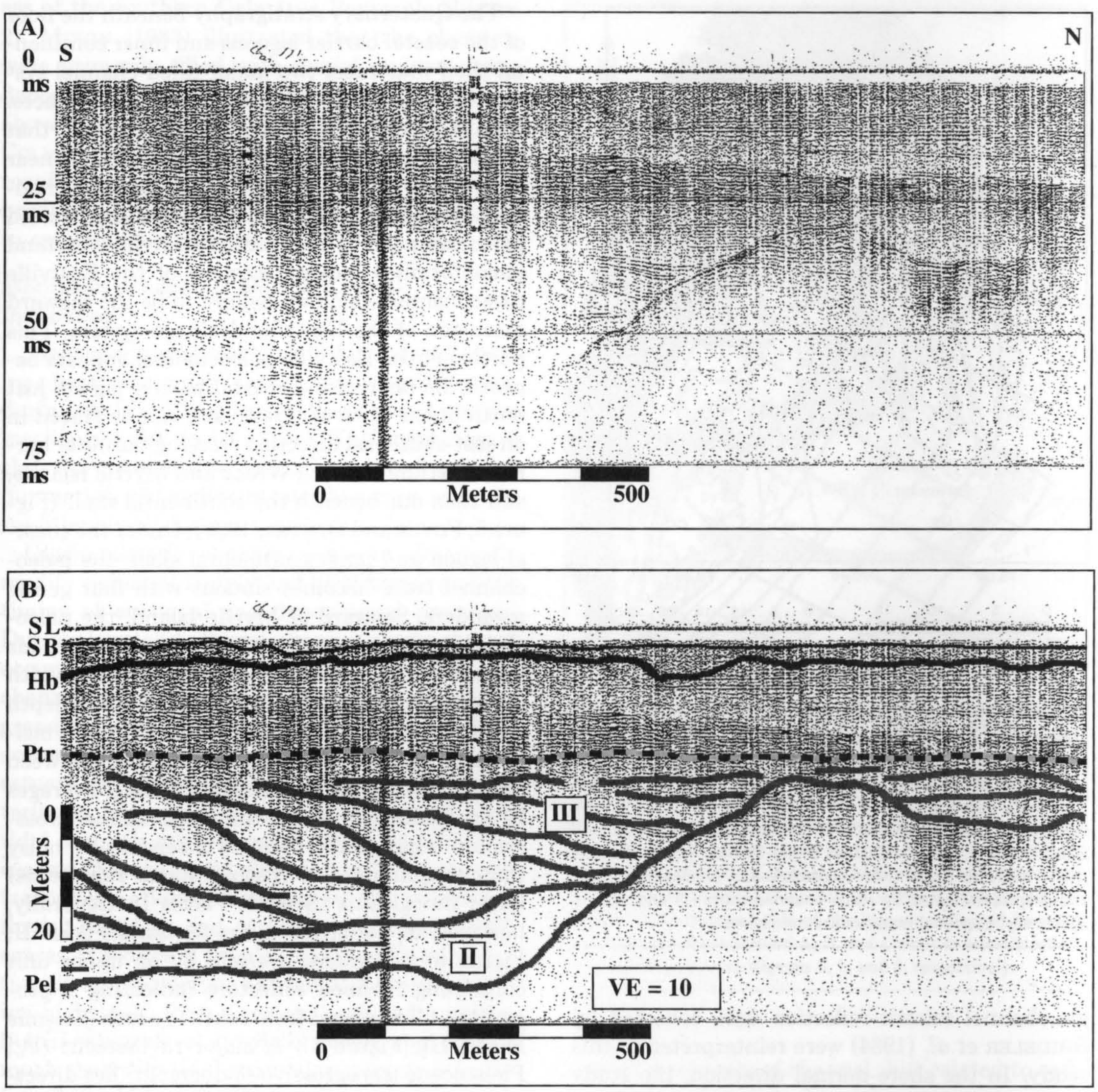

Figure 6. Uninterpreted (A) and interpreted (B) versions of seismic-reflection profile A, from South Bay, landward of Wreck Island. Figures show the Eastville paleochannel, containing Seismic Facies II (estuarine deposits) and overlying Seismic Facies III (estuary-entrance spit fill). Pel = Pleistocene Eastville lowstand erosional surface; Ptr = Pleistocene transgressive ravinement surface; Hb = late Wisconsinan lowstand erosional surface (basal Holocene reflector). Seismic Facies I (fluvial fill) was not preserved. Location of profile shown in Figure 4.

channel trace is sinuous with three broad meanders. The seawardmost part of the paleochannel trace heads off in an east-northeast direction. The paleochannel has an average width of 4.5 km and the thalweg depth (below present msl-datum) ranges from 50 to 70 m and relief averages 25 m. Inclined reflectors within the valley fill show a dominantly south-to-north fill pattern that overlies a subhorizontal-parallel reflector pattern (Seismic Facies III and II, respectively; Figure 7). The seismic reflectors representing the Exmore surface and paleochannel (P_{xl}; Pleistocene Ex-

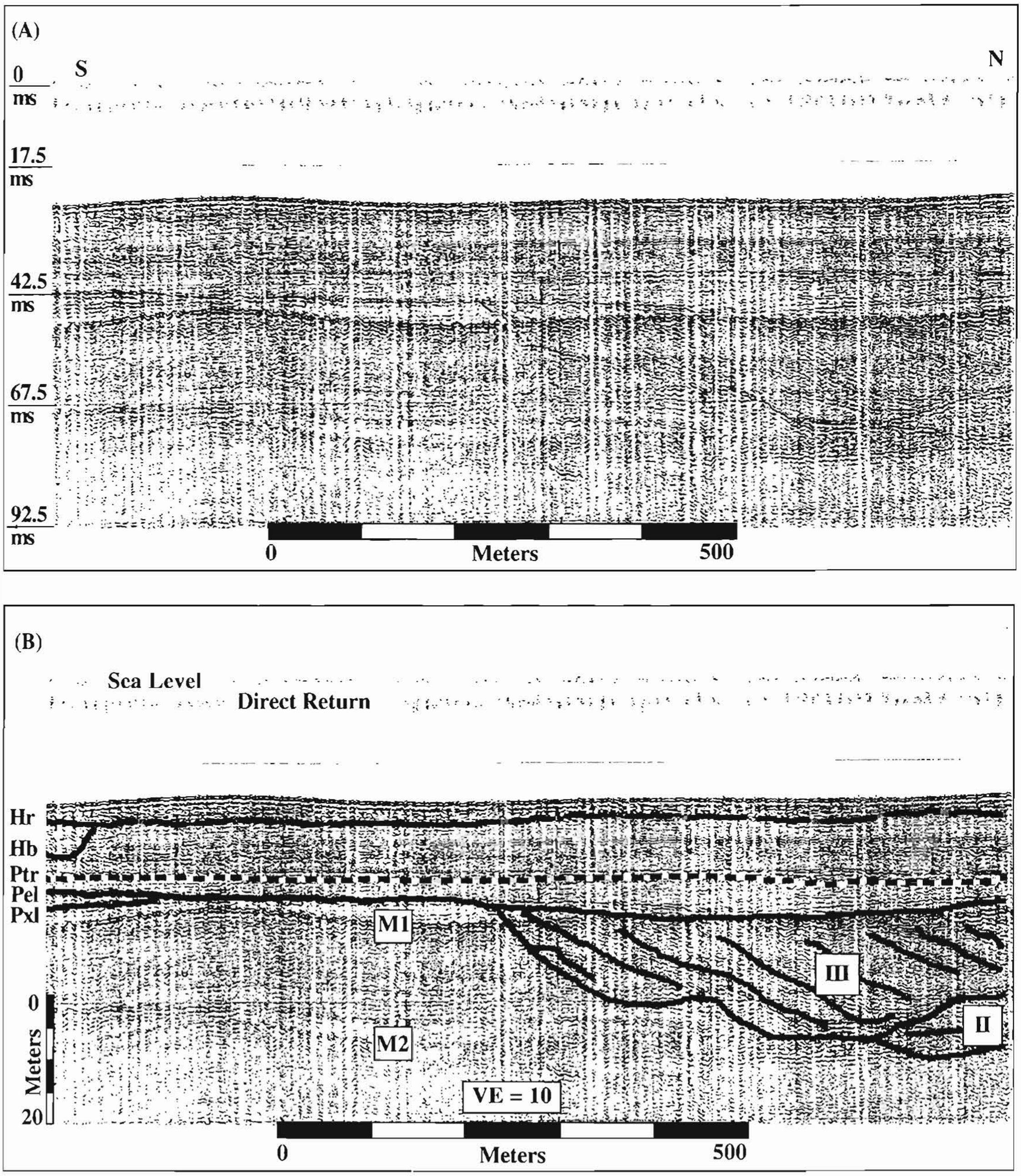

Figure 7. Uninterpreted (A) and interpreted (B) versions of seismic reflection profile B, from the inner shelf seaward of Cobb Island. Figures show the Exmore paleochannel, containing Seismic Facies II (estuarine deposits) and overlying Seismic Facies III (estuary-entrance spit fill). Pxl = Pleistocene Exmore lowstand erosional surface; Pel = Pleistocene Eastville lowstand erosional surface; Ptr = Pleistocene transgressive ravinement surface; Hb = late Wisconsinan lowstand erosional surface; Hr = Holocene ravinement surface; M1 = seabed first-order multiple. M2 = seabed second-order multiple. Seismic Facies I (fluvial fill) was not preserved. Location of profile shown in Figure 4.

more lowstand), and channel-fill succession, have been extensively truncated by the overlying Eastville unconformity surface (P_{el}).

Our seismic profiles have also revealed a paleochannel (here named the Belle Haven paleochannel) that has not been described previously. Geographically, it is located between the Exmore and Eastville paleochannels. This paleochannel runs diagonally from the southern part of Upshur Neck, south-southeast across the lagoon and crosses the barrier coastline at the north end of Cobb Island (Figure 5). Below the inner shelf, the paleochannel continues southeastward and is truncated by (and enters) the Eastville paleovalley approximately 10 km seaward of Wreck Island. The paleochannel has an average width of 4 km and the thalweg depth (below present msl-datum) ranges from 36 to 55 m. The paleochannel relief averages 20 m. Beneath Hog Island Bay, inclined reflectors within the channel fill show a south-to-north fill pattern (Seismic Facies III). We feel that the Belle Haven paleochannel provides a key piece of information needed to understand the mechanics of stream transfer between the Exmore and Eastville drainage systems.

Seismic profiles also reveal a fourth paleochannel, the Cape Charles paleochannel (COLMAN and MIXON, 1988; COLMAN *et al.* 1990), at the Chesapeake Bay entrance. The trace of the paleochannel runs southeastward along the west side of the southern Delmarva Peninsula in a large arc that bends gently eastward and then northeastward below the seabed to the south and east of Fishermans Island (Figure 5). At the seawardmost side of the study area, the channel trends north-northeast at a bearing that would produce an intersection with the Eastville paleochannel about 40 kilometers seaward of Wreck Island. The basal Holocene reflector (H_b), representing the floor of the filled channel and its associated interfluve surface, truncates both the regional Pleistocene ravinement (P_{tr}) and the Eastville lowstand surface (P_{el}) (Figure 8). Within the baymouth, the southern flank of the Cape Charles paleochannel slopes upward to an interfluve about 13 m above the channel floor. This interfluve was probably originally at a higher elevation prior to tidal scouring during Holocene marine flooding. The channel has a width in excess of 6 km, and the thalweg depth (below present msl-datum) ranges from 38 to 50 m. Paleochannel relief (compared with northern interfluve elevation) averages 25 m. Seismic facies in the buried channel consist of poorly preserved lower fluvial facies (Seismic Facies I), and an estuarine succession (Seismic Facies II), which have filled the channel to within 6 m of the interfluvial surface. Seismic Facies II is disconformably overlain by clinoforms of estuary-entrance spits (Seismic Facies III), whose reflectors dip southward and bayward. This unit has filled the remaining upper 13.5 m of the channel and has migrated across the interfluvial surface to the south (also see MEISBURGER, 1972; COLMAN *et al.*, 1988). An uppermost unit (Seismic Facies IV; not seen in the Exmore, Belle Haven, and Eastville paleochannels), varies in thickness from 0 to 11 m and is believed to represent deposition at modern estuary-entrance shoals (*e.g.*, Nautilus Shoals and Middle Ground Shoals). Seismic Facies IV appears best developed downdrift of Nautilus and Smith Island Shoals, and is more discontinuous offshore due to shoreface scour. To the west, in the Chesapeake Bay mouth just seaward of North Channel, Seismic Facies IV is truncated locally by tidal scour, and Seismic Facies III crops out on the Chesapeake Bay floor.

A fifth paleochannel (tentatively named the Middle Ground Shoal paleochannel) is a minor fluvial channel located between the Cape Charles paleochannel and the modern Chesapeake Channel. This paleochannel is approximately 1 km wide and correlates with "Channel B" illustrated in MEISBURGER (1972), and has also been identified by COLMAN and HOBBS (1987). The Middle Ground Shoal (MGS) paleochannel has fill facies similar to those described above for the Cape Charles paleochannel. Seismic Facies IV is represented by a discontinuous (locally scoured by tidal currents) 3 to 6 m thick sequence of the modern Middle Ground Shoal complex that caps the Middle Ground Shoal paleochannel and extends southward to the margin of the modern Chesapeake Channel.

INTERPRETATIONS

The development of the Accomack and Nassawadox spits has had a major influence on the pathways of fluvial systems draining from the Chesapeake Basin. However, sea-level elevations during the respective highstand events may have had an equally important impact on the mechanics of channel fill and migration.

Interpretations of our seismic profiles from the seaward side of the southern Delmarva Peninsula concur with those of COLMAN and MIXON (1988), in that the Exmore, Eastville and Cape Charles

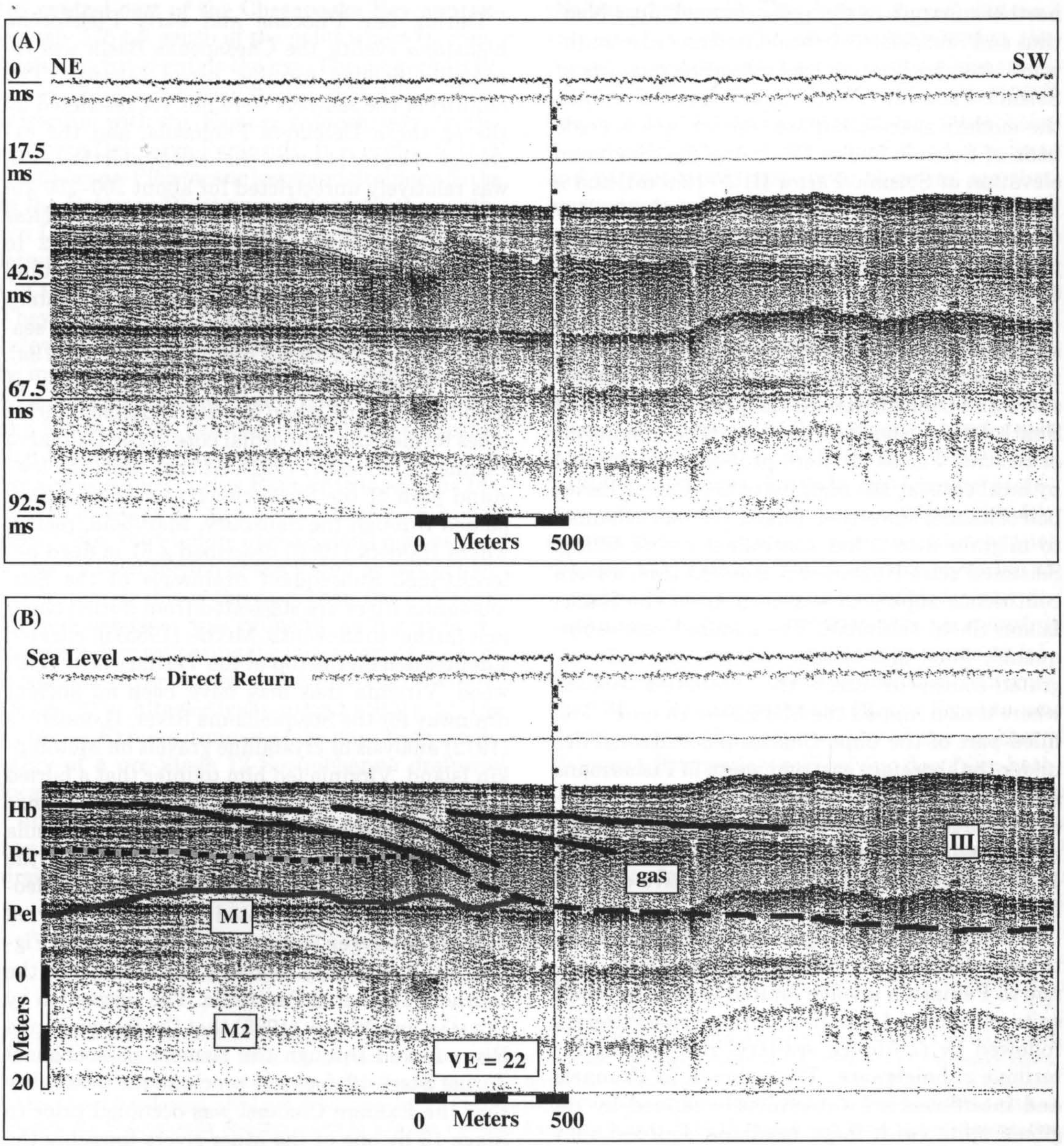

Figure 8. Uninterpreted (A) and interpreted (B) versions of seismic reflection profile C, from the inner shelf seaward of the Chesapeake Bay mouth. Figures show the Cape Charles paleochannel, containing Seismic Facies III (estuary-entrance spit fill). Pel = Pleistocene Eastville lowstand erosional surface; Ptr = Pleistocene transgressive ravinement surface; Hb = late Wisconsinan lowstand erosional surface; M1 = seabed first-order multiple. M2 = seabed second-order multiple. Seismic Facies I (fluvial fill) and II (estuarine deposits) were obscured by gas. Location of profile shown in Figure 4.

paleochannels were discrete fluvial valleys. However, east of the axis of the southern Delmarva Peninsula, the direction of tidal-channel migration within the Eastville, Belle Haven, and Exmore paleochannels (based on apparent dips in Seismic Facies III) was from south to north. The only evidence of southward channel migration was in the upper "tidal" section of the Cape Charles paleochannel beneath the modern Chesapeake Bay entrance. Inclined reflectors of Seismic Facies III in the Cape Charles paleochannel originate on the northern margin of the paleochannel, near Nautilus and Smith Island Shoals, and record a southward lateral migration over estuarine deposits of Seismic Facies II. These clinoforms do not reach the modern seabed, but are capped by low-angle beds of Seismic Facies IV. Using the maximum elevation of Seismic Facies III (−12.6 m), and a Holocene sea-level curve for the Chesapeake Bay (COLMAN *et al.*, 1992), we calculated that the spit prograded over the bay-mouth interfluve about 6,500 BP when sea level was at about −10 m (relative to modern sea level). Thus, during initial stages of paleochannel fill by Seismic Facies III, the paleocoastline was approximately 15 km offshore and coincident with the present location of Smith Island Shoals (Figure 5). The clinoforms of Seismic Facies III were probably not formed by local spits at the margins of the Cape Charles paleochannel since they forced the paleochannel to migrate 2 to 3 km southward before filling. Seismic Facies III probably resulted from a large continuous supply of sediment from the Smith Island Shoal shoreline. The channel was subsequently filled and abandoned, and the spit migrated southward across the interfluvial surface where it also capped the MGS paleochannel. The filled part of the Cape Charles paleochannel lies under the shoreface east and south of Fishermans Island. Bayward of Fishermans Island, the channel is still open, and is designated as North Channel on modern navigation charts (*e.g.*, NOAA chart 12221). The North Channel is an active conduit for inlet tidal currents.

Our ideas for the origin and evolution of the three paleochannels beneath the southern Delmarva Peninsula involve limited forcing by small spits on channel margins during transgression, followed by continued sea-level rise causing interfluve submergence. The submerged channels and interfluves are subsequently capped by regional spits (with large headland sources) that migrate over both surfaces during highstand. A subsequent regression causes fluvial drainage to follow the topographically low areas downdrift of the regional spit. The mechanism requires relatively large fluctuations in sea level which are associated with glacio-eustatic events, hence the term "time-lagged" channel-jumps. The model's application to the southern Delmarva Peninsula is dependent on the involvement and concurrent maturation of all of the major fluvial-channel systems in the Chesapeake Basin since the Pliocene.

The Susquehanna System

During late Pliocene and early Pleistocene highstand events, the Chesapeake Basin was repeatedly flooded forming broad seaways. The mouths of the seaways were not yet restricted by the southern Delmarva Peninsula, and the exchange of water between the seaway and the ocean was relatively unrestricted for about 200–250 km between headlands formed by the Pliocene deltas to the north and southwest (see Figure 2 in KRANTZ, 1991; RAMSEY, 1992). During lowstand events, watersheds on the eastern and western sides of the basin matured as they drained seaward through the broad basin mouth (Figure 9a).

The Susquehanna River has existed since at least the Pliocene, when highstand deltaic-marine lobes formed the primordial core of the Delmarva Peninsula (RAMSEY, 1992). At that time, the lowstand path of the Susquehanna River may have passed through the Salisbury, Maryland, region, where HANSEN (1966) described a 61 m deep paleochannel. Subsequent drainways of the Susquehanna River are suggested from buried channels farther to the south. MIXON (1985) illustrated a depression on the top of the Tertiary near Hallwood, Virginia that may have been an ancient drainway for the Susquehanna River. HARRISON's (1972) analysis of crystalline gravels on Metompkin Island, Virginia led him to infer that a buried paleochannel (from the Susquehanna or Potomac Rivers) passed beneath the Delmarva Peninsula in that area. Thus, prior to the time when the Susquehanna River occupied the Exmore paleochannel (generally accepted as having occurred during the oxygen isotope Stage 12 lowstand; Figure 2a), it probably followed paths north of the southern Delmarva Peninsula. The suggestion by COLMAN and MIXON (1988) that the Susquehanna River flowed through the Exmore paleochannel during Stage 12 does not preclude the possibility that the Exmore Channel was occupied prior to Stage 12 by one of the other rivers (possibly the Potomac River) that drained across the basin during earlier lowstand events. We believe that the Exmore paleochannel was probably already active prior to Stage 12 and was providing drainage to the Potomac watershed during lowstand events (Figure 9a).

The Potomac-Exmore System

The modern Potomac River has the second largest watershed (≈ 30,000 km^2; MIXON, 1985) of the numerous Piedmont rivers that flow through the Chesapeake Basin. This system drains into the central part of the Chesapeake Bay approximately 175 km south of the point where the Susquehanna River enters the bay. The modern trend of the Potomac estuary is on a direct line-of-intersection with the Exmore paleochannel on the southern Delmarva Peninsula. It is probable that the Pliocene-Pleistocene lowstand drainage paths of the Potomac River passed beneath the southern Delmarva area, prior to the formation of the latter (Figure 5). Seismic evidence shown by COLMAN and MIXON (1988) may support this idea. They illustrated a line of buried paleochannels below the floor of the Chesapeake Bay that could be interpreted to link the Potomac River and the Exmore paleochannel (see Figure 7 in COLMAN and MIXON 1988). COLMAN *et al.* (1990) suggested that the mouth of the present Potomac River has been re-occupied during Exmore (Stage 12 or 14), Eastville (Stage 6) and Cape Charles (Stage 2) time. Thus, the Exmore paleochannel may be considered a local section of the Potomac paleochannel system. If this is the case, then the Potomac-Exmore drainway was probably an independent system before capturing the lower part of the Susquehanna River during a more recent lowstand (Stage 12?). Alternatively, prior to Stage 12, the Potomac-Exmore drainway may have been a tributary of a pre-Stage 12 Susquehanna drainway located to the north.

It is difficult to assign a date to the Susquehanna capture event (by the Potomac-Exmore drainway), since it must be based on the age and superposition of spit deposits that prograded over the buried channel. The dating of overlying marine spits can only be used to suggest that the fluvial channel was present prior to the highstand spit fill. Although MIXON (1985) and SZABO (1985) suggest that the Accomack spit may have been deposited during a Stage 7 highstand, COLMAN and MIXON (1988) appear to favor an age that corresponds with oxygen-isotope stage 11. This would make the Potomac-Exmore drainway Stage 12 or older. WEHMILLER *et al.* (1988) suggested that the Shirley Formation on the western side of the Chesapeake Bay (which is believed to be correlative with the Accomack barrier spit of the Omar Formation) was a Stage 13 marine deposit. However, DEMAREST and LEATHERMAN (1985) suggest that deposits of the Accomack barrier spit may be older than Stage 15 (> 600,000 BP). This would require that the Potomac-Exmore drainway was open during or prior to Stage 14. We suggest that the Potomac-Exmore paleochannel drained the Potomac River exclusively prior to Stage 13. Then the Susquehanna River was diverted into the Potomac-Exmore drainway during oxygen-isotope Stage 12, following a Stage 13 initial progradation of the Accomack spit (Figure 9b). Thus, by isotope Stage 12, the Exmore paleochannel on the eastern side of the southern Delmarva Peninsula was receiving flow from both the Susquehanna and Potomac watersheds (Figure 9b). A second stage of Accomack spit development during Stage 11 resulted in the final abandonment of the Exmore paleochannel beneath and

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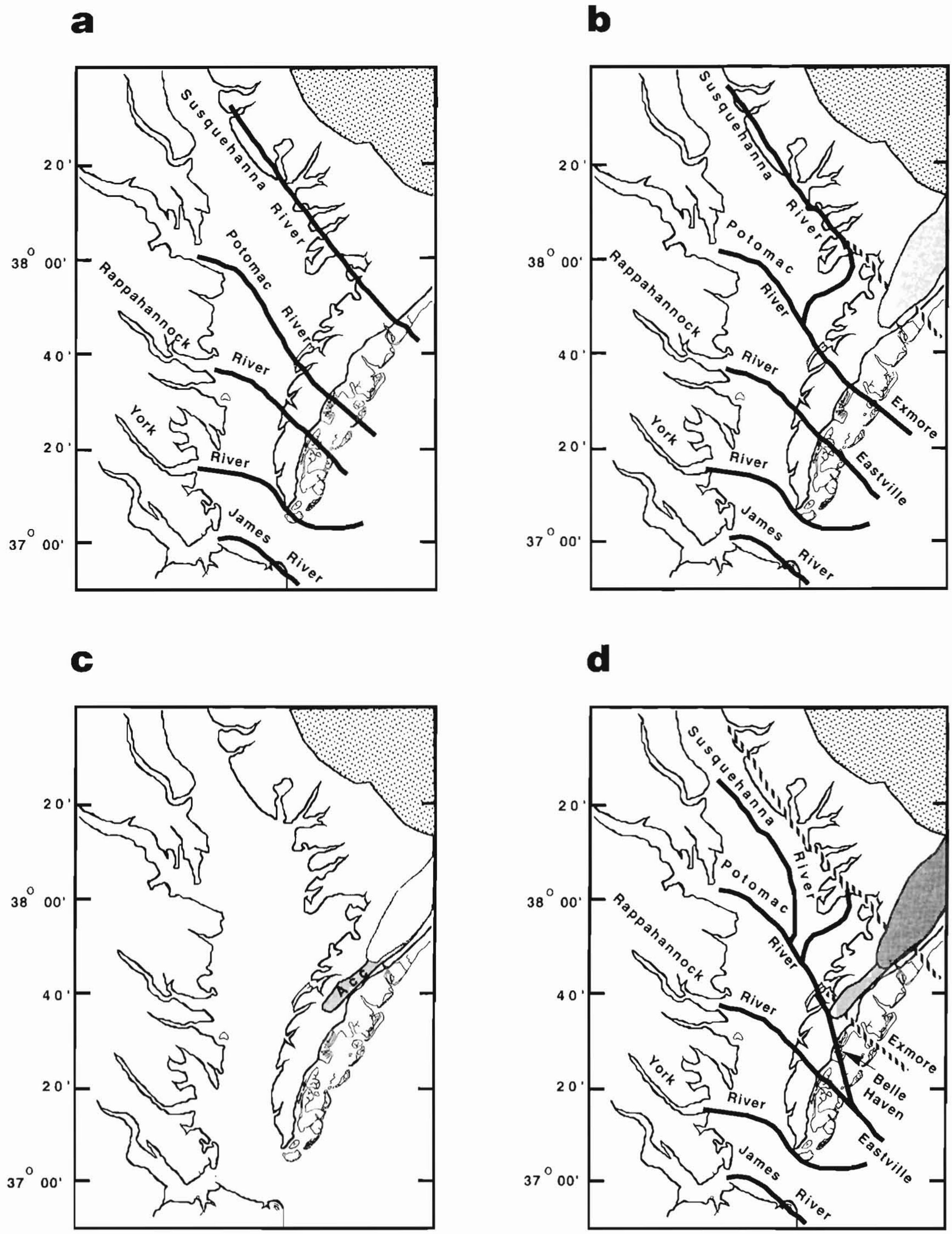
Figure 9. Revised model for the geologic evolution of the southern Delmarva Peninsula and time-lagged channel diversion and shifting of the Susquehanna River. (a) Depiction of late Pliocene lowstand drainage pattern of the primordial Chesapeake Basin showing five northeast-to-southwest drainways flowing through a broad basin mouth. (b) Depiction of an early Pleistocene lowstand (Stage 12 or 14) following initial spit development from Pliocene deltaic deposits of the northern Delmarva Peninsula. The spit has diverted the Susquehanna drainage south to join the Potomac-Exmore drainway. The Rappahannock River flowed through the Eastville paleochannel. The York River may have joined the Rappahannock-Eastville system offshore. The James River exited the basin in the southern entrance area. (c) Stage 11 highstand development of the Accomack spit across the northern part of the entrance to the Chesapeake seaway. (d) Stage 10 lowstand drainage of the Chesapeake Basin illustrating southward shift of the Susquehanna River, and flow of the Susquehanna and Potomac systems through the Belle Haven paleochannel and then into the Rappahannock-Eastville paleochannel. The Exmore paleochannel was filled with sediment and was no longer active. The York River joined the Rappahannock-Eastville system offshore. The James River exited the basin in the southern entrance area. (e) Stage 7/9? highstand development of the initial Nassawadox spit (Nas 1) across the central part of the Chesapeake seaway. (f) Stage 6 lowstand development of the Chesapeake Basin illustrating the filled Belle Haven paleochannel and diversion of flow from the Susquehanna and Potomac along the west side of Nas 1 to the Rappahannock-Eastville drainway. (g) Stage 5e highstand development of the Nassawadox spit (Nas 2) across the south-central entrance of the Chesapeake seaway. (h) Stage 2 lowstand drainage of the Chesapeake Basin illustrating closure of the Eastville section of the Rappahannock system forcing all drainage north of the York River to be diverted into the Cape Charles paleochannel before returning to the Eastville trend offshore. The James River exited the basin in the southern entrance area.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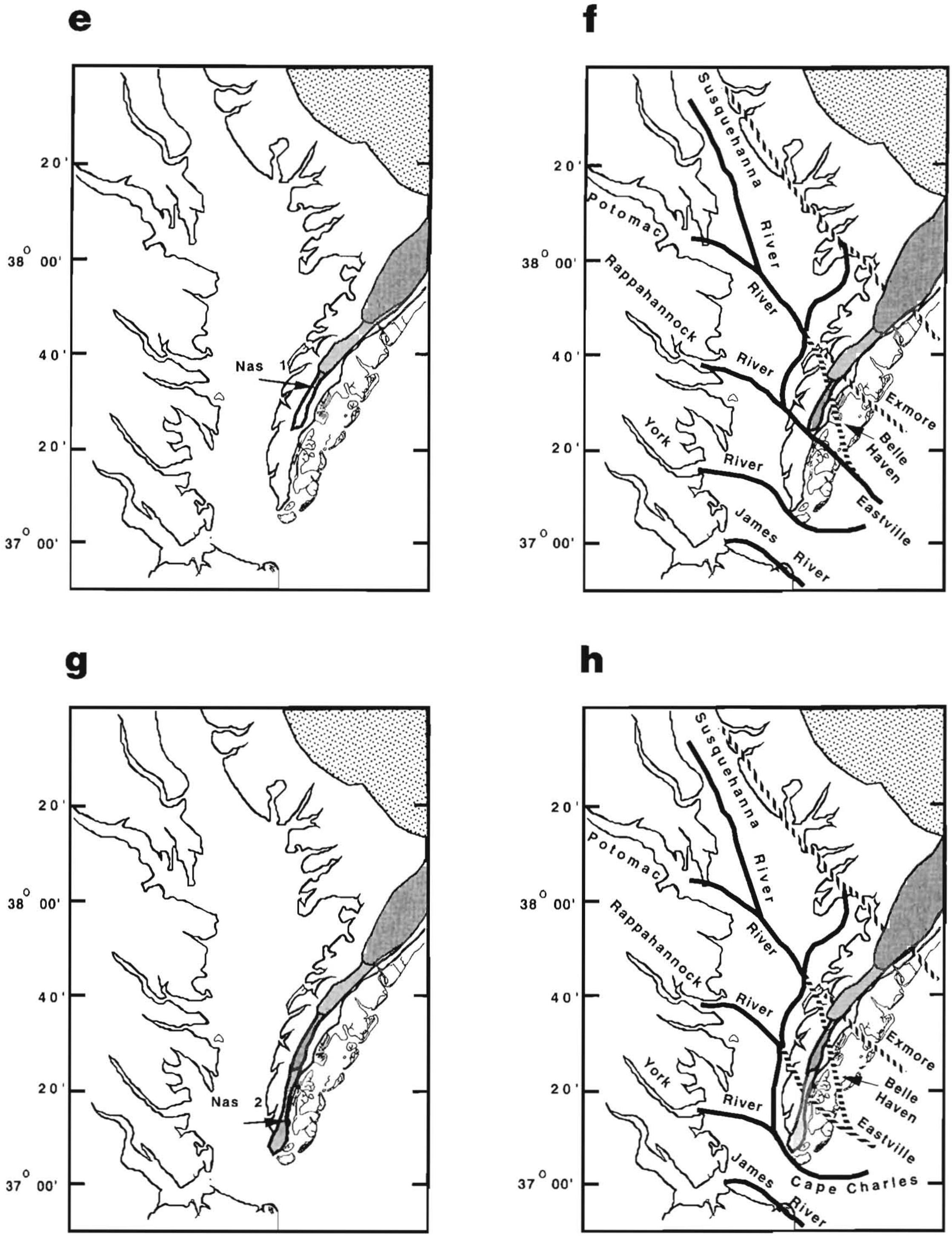
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seaward of the southern Delmarva Peninsula (Figure 9c and 9d).

While a Stage 11 highstand progradation of the Accomack spit contributed to filling of the Exmore paleochannel (based on peninsula well-log data; MIXON, 1985; COLMAN and MIXON, 1988), the large-scale spit which capped the paleochannels is not apparent farther seaward, beneath the lagoon and inner shelf (Figure 7). In these areas, much of the channel fill shows south-to-north dipping beds produced by estuary-entrance shoals and estuary-entrance spits that migrated in from the margins of the paleochannel.

The Belle Haven Paleochannel

The Belle Haven paleochannel, located between the Exmore and Eastville paleochannels runs northwest-southeast under the coastal lagoon. It extends southeast from the Exmore paleochannel, near Belle Haven, Virginia and terminates beneath the shoreface where it joins the Eastville paleochannel seaward of Wreck Island (Figure 5). Channel fill and capping of the Exmore paleochannel by the Accomack spit during Stage 11 made the Belle Haven paleochannel a more efficient pathway for the subsequent lowstand drainage of the Potomac and Susquehanna Rivers (Figure 9d). The Belle Haven paleochannel was closed by an early phase of development of the Nassawadox spit (prior to Stage 5e) which prograded southward to Eastville, Virginia sometime between Stage 10 and Stage 6 (Figure 9e). This interpretation suggests that the Nassawadox spit, generally accepted as being of oxygen isotope Stage 5e age, may have an older northern section, and therefore be multicyclic. The early Nasssawadox spit migrated over the filled Belle Haven paleochannel and across its southern interfluve (composed of Tertiary marine deposits) between Belle Haven and Eastville, Virginia (see Figure 9 in MIXON, 1985). Sea level was sufficiently high during the highstand event to totally submerge the interfluve on the south side of the Belle Haven paleochannel and the low-lying area of the Chesapeake Basin to the south. The inundation of this part of the Chesapeake Basin allowed for a free exchange of water in the Chesapeake seaway (an ancestral Chesapeake Bay) without putting hydraulic pressure on the downdrift side of the Belle Haven channel. Thus, only minor channel migration was necessary to maintain dynamic equilibrium with the upper tidal section of the active Belle Haven Channel.

The Rappahannock-Eastville System

We believe that the Eastville paleochannel originally may have been an extension of the Rappahannock River and was in existence prior to its generally accepted Stage 6 incision date. A regional overview of the Chesapeake Basin reveals that the orientation of the Rappahannock River is on a line-of-intersection with the Eastville paleovalley. Buried paleochannel orientations (COLMAN and MIXON, 1988; COLMAN *et al.*, 1990) allow a partial reconstruction of the channel trace between the Rappahannock estuary and the Eastville paleochannel. If so, the Eastville paleochannel may be considered a local section of the Pleistocene Rappahannock drainway and not just a Stage 6 channel opened by the diverted Susquehanna River as suggested by COLMAN and MIXON (1988). Our interpretation is that the Eastville section of the Rappahannock drainage may have been active since at least the Stage 12 lowstand (Potomac-Exmore time). This would indicate that the Eastville paleovalley was then reoccupied during Stage 6 and potentially during intervening lowstands also.

During the Stage 6 lowstand, the combined flow of the Susquehanna and Potomac Rivers likely drained along the western side of the southern Delmarva Peninsula and was captured by the Rappahannock-Eastville system just south of Eastville, Virginia (Figure 9f). The addition of discharge from the Potomac and Susquehanna drainage systems thus increased the rate of channel scour removing earlier (pre-Stage 6) records of the more diminutive Rappahannock drainage.

During the Stage 5e highstand, the Eastville paleochannel was filled by fluvial and estuarine sediments and capped by a second phase of development of the Nassawadox spit (see cross sections in MIXON, 1985; COLMAN *et al.*, 1990). The Stage 5e development of the Nasssawadox spit migrated over the filled channel, across the interfluve (composed of Tertiary marine deposits) and extended southward to the modern entrance to the Chesapeake Bay (Figure 9g). This highstand also re-submerged the main drainways and interfluves of the York and James Rivers on the south side of the Chesapeake Seaway.

During the subsequent lowstand (Stages 4-2), the drainage of the northern watersheds was diverted along the western side of the Nassawadox spit, and around the tip of the southern Delmarva Peninsula where it was captured by the York-

Cape Charles drainage system (Figure 9h). On the south side of the Bay entrance, the paleochannel of the James River headed southeast below Cape Henry, Virginia (HARRISON *et al.*, 1965; MEISBURGER, 1972; SWIFT *et al.*, 1972) and onto the southern Virginia inner shelf (Virginia Beach Shelf Valley; SWIFT *et al.*, 1977).

The York-Cape Charles System

A regional overview of the Chesapeake Basin shows that the trend of the York River is on a line-of-intersection with the modern tidal-flushed North Channel and the Stage 2 Cape Charles paleochannel. COLMAN and MIXON (1988) and COLMAN *et al.* (1990) illustrate that during oxygen isotope Stage 2 the York River extended eastward under the floor of the Chesapeake Bay and connected with the Cape Charles paleochannel south of the town of Cape Charles on the southern Delmarva Peninsula. Our data in the entrance to the Chesapeake Bay show that the Cape Charles paleochannel continues to run eastward and then east-northeastward beneath the inner shelf. Projection of the course would produce an intersection with the Eastville paleochannel approximately 50–60 kilometers east of the present York River mouth (Figure 5; FOYLE and OERTEL, 1992). This suggests that during previous Pleistocene lowstands (Stages 6 and 8, and possibly Stage 10) the York River was a tributary of the Eastville system (Figure 9). To support this interpretation, a tributary of the Eastville paleovalley has been observed to run along and beneath the northern flank of the York-Cape Charles paleochannel.

As described above, the Cape Charles system captured the drainage from the Eastville paleochannel (with its combined flows from the Rappahannock, Potomac, and Susquehanna Rivers) during the Stage 4-2 lowstand when water receded back into the channels on the floor of the Chesapeake Basin. The combined drainage of these systems joined the York River tributary near Cape Charles, Virginia and then continued seaward to probably reoccupy remnants of the Eastville paleochannel on the outer shelf.

The York-Chesapeake System

The Holocene transgression has produced the present sea-level highstand that has flooded much of the Chesapeake Basin producing the modern Chesapeake Bay (the most recent in the series of Chesapeake seaways). With each Quaternary highstand event, the Chesapeake seaway has become more restricted because of the progressive extension of the southern Delmarva Peninsula. The present highstand inundation has left three channels submerged on the floor of the Chesapeake Bay entrance. From north to south they are the North Channel, the Chesapeake Channel and the Thimble Shoals Channel. Thimble Shoals Channel is associated with the submerged Holocene valley of the James River (HARRISON *et al.*, 1965; COLMAN and HOBBS, 1987) and tidally exchanges James River water with the Atlantic Ocean. The Chesapeake Channel is the principal conduit exchanging the Chesapeake Bay water with the Atlantic Ocean (OERTEL and WADE, 1981). The "tidal front" forming the northern side of the bay-mouth plume is generally found between Middle Ground Shoal and the northern side of the Chesapeake Channel. Salinity gradients across this front are often 4 ppt per 50 m (OERTEL and WADE, 1981). The North Channel is a major conduit receiving well-mixed oceanic water. It connects with Beach Channel and other relict Susquehanna River channels on the Chesapeake Bay floor to form a continuous path to the modern Susquehanna River mouth at the head of the Chesapeake Bay (COLMAN *et al.*, 1992). Just east of the Chesapeake Bay entrance the North Channel ramps upward onto bay-entrance shoals and loses its bathymetric expression on the shoreface. The connection between the North Channel and the York River also has been filled during the recent highstand by shoal migration in the lower-bay-entrance area (COLMAN *et al.*, 1988).

Modern Processes

Open regions of broad estuary entrances lack the surf-wave dynamics needed to drive strong littoral-drift systems. As a result, spit progradation processes are inhibited away from the bay margins, and channel filling is related to entrance-shoal development. Sediment transport dynamics at estuary entrance shoals is much more complex than spit progradation. Aggradation of shoals involves mutually evasive tidal flow and sediment transport patterns (LUDWICK, 1972; 1974; 1975). Bathymetric shielding and shoal expansion is often controlled by strong bi-polar tidal currents that are oblique to the trend of the coastline.

The "chevron-shaped" shoal systems in the modern Chesapeake Bay entrance reflect the preferred flood path of sediment up the axis of the shoal and the preferred ebb path along the shoal margins (LUDWICK, 1972; 1974; 1975). Inner Middle Ground Shoal is a "chevron-shaped" shoal responsible for the blockage of the submerged valley of the York River on the bay floor inside of the Chesapeake Bay entrance. The landward migration of the shoal axis has filled a section of the York Channel that previously connected the York River with the North Channel (COLMAN *et al.*, 1988). The blockage has caused the modern tidal exchange of the York River to take a more southerly route, through the centrally located Chesapeake Channel in the Bay entrance.

The modern barrier island coastline of the southern Delmarva Peninsula terminates at Fishermans Island. This island was not present on early charts of this area but was described as a series of shoals known as the "The Issacs". The Issacs shoals were part of a spit platform which extended from the southernmost end of the Delmarva Peninsula. Emergence of the shoals produced Fishermans Island. This spit platform below Fishermans Island and the Isaacs shoals represents the most recent stage in the sequential progradation of the southern Delmarva Peninsula. The spit platform is relatively short, and landscape evidence for continuous spit progradation (*i.e.*, sequential sets of beach ridges) is only apparent as far north as the middle of Smith Island. North of Smith Island, most of the barrier islands are tide-dominated systems (OERTEL and KRAFT, 1994) and as such do not transmit a large volume of sediment across the deep tidal gorges. Thus, under the present hypsometric conditions, a large source of littoral sediments from a "regional" spit is not available to produce spit forcing and channel migration at the southern end of the Delmarva Peninsula. Modern spit progradation in the vicinity of Smith and Fishermans Islands is driven by a local estuary-margin spit. On the south side of the Chesapeake Bay entrance, Cape Henry was formed by the development of similar estuary-margin spits.

The Smith-Fisherman Island spit has filled the uppermost part of the Cape Charles paleochannel adjacent to the southern tip of the lower Delmarva Peninsula (Seismic Facies III). Our data show that the Cape Charles paleochannel was 25% filled with fluvial sediments (Seismic Facies I) and estuarine sediments (Seismic Facies II) prior to estuary-entrance spit filling (Seismic Facies III). Thus, the spit capped the fluvial paleochannel rather than forcing it laterally. Continued southerly spit migration across the bay mouth created thinning deposits to cap the Middle Grounds Shoal paleochannel. Spit clinoforms now define the northern margin of the modern Chesapeake Channel. The migration of the spit midway across the bay entrance is believed to have begun approximately 6500 BP, when sea level was about 10 m lower than present and a wave-dominated oceanic shoreline was located in the vicinity of the Smith Island Shoals. The present highstand has submerged the spit to depths of 12 m in the entrance of the Chesapeake Bay and on the northern half of the entrance area the spit clinoforms (Seismic Facies III) are capped by a sand-shoal facies (Seismic Facies IV) that is about 5 m thick. The two units are separated by a bay ravinement.

In summary, migration of the Cape Charles paleochannel by spit forcing in the Chesapeake Bay mouth area was probably only active during the early stages of spit progradation (about 6,500 yrs BP), associated with the Smith Island Shoal shoreline (15 km seaward of the modern coastline). When sea level rose above the southern margin of the Cape Charles paleochannel, the lower part of the paleochannel filled, and only the upper section continued to migrate southward (Figure 10). This mechanism may also explain why the buried Susquehanna River channel did not force the paleochannels laterally beneath the entire length of the southern Delmarva Peninsula.

The Next Regression

Assuming that the rhythmic cycle of transgressive and regressive events will continue, then on a geologic time scale, sea level is expected to fall in the future. As sea level falls during the next regression, water depths in the Chesapeake Bay will get progressively shallower, eventually causing river systems to drain through linear depressions of relict channels on the bay floor. The present bathymetry of the Chesapeake Bay still has the imprint of the previous, Stage 2, subaerial drainage (COLMAN and HOBBS, 1987; COLMAN and HALKA, 1989; COLMAN *et al.*, 1992). The main trunk-channel of the Stage 2 system may be traced along the Chesapeake Bay floor from the bay head near Havre de Grace, Maryland to the Cape Charles paleochannel near Cape Charles, Virginia as a series of scars on the bay floor.

Modern channel fill of the Cape Charles paleochannel in the Chesapeake Bay entrance will alter the future lowstand drainage (see the section on the York-Chesapeake above and COLMAN and MIXON, 1988) . The partial blockage of North Channel in the Chesapeake Bay entrance will cause

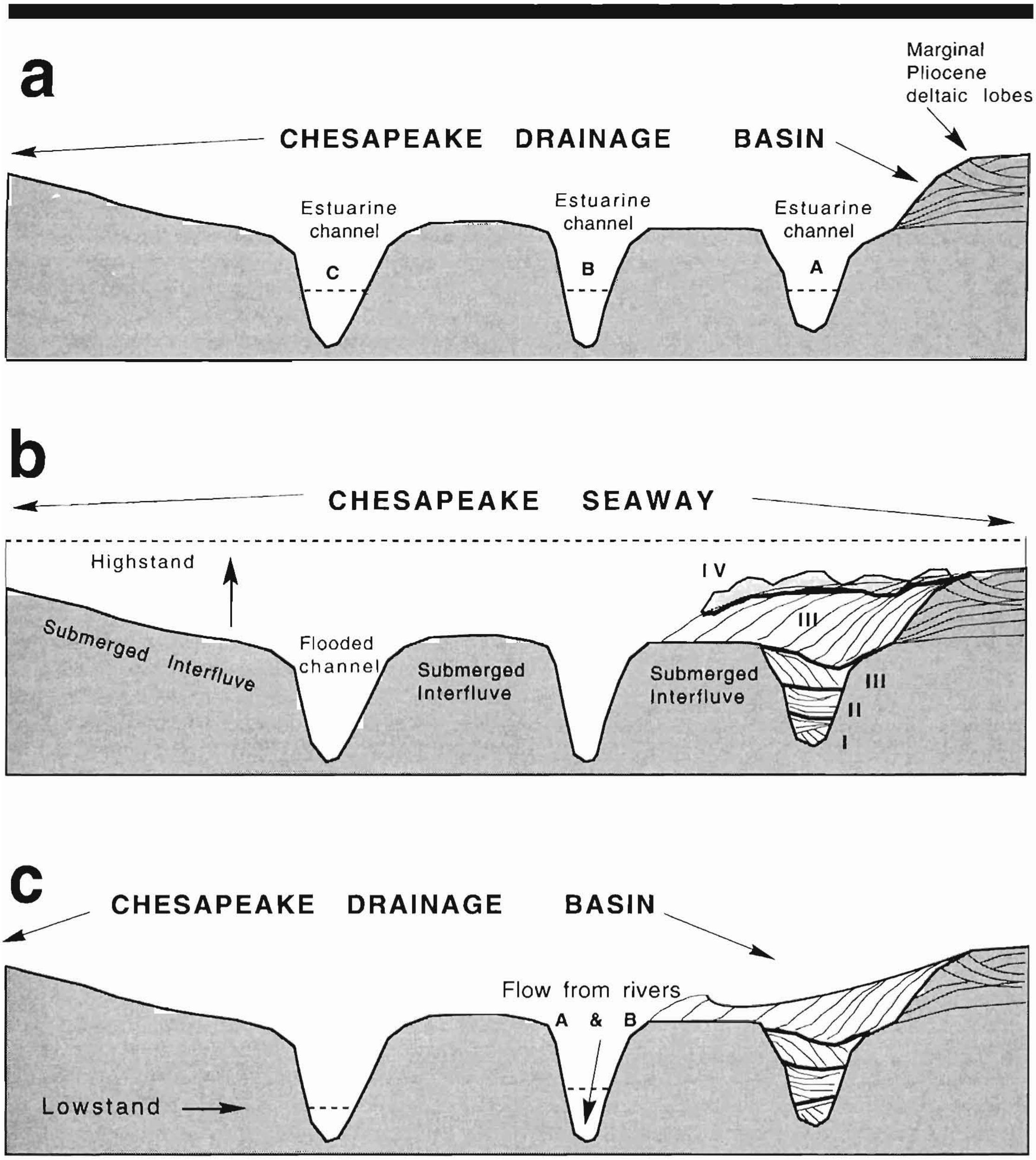

Figure 10. Idealized cross-section of the Chesapeake Bay entrance illustrating the effects of sea-level rise on spit migration, channel shifting, and channel filling. (a) Depiction of lowstand drainage of the Chesapeake Basin with multiple rivers crossing the entrance area. (b) Depiction of highstand submergence of channels and interfluves forming an ancestral Chesapeake Seaway. Estuarine paleochannel A filled upward with fluvial facies (Seismic Facies I), transgressive estuarine facies (Seismic Facies II), estuary entrance spits (Seismic Facies III), and then capped by a large regional spit (Seismic Facies III) and estuary-entrance shoals (Seismic Facies IV). Pliocene deltas provided an abundant source of sediment for the regional spit which prograded over the paleochannel A and onto the interfluve. (c) Diversion of channel A flow into channel B when sea level falls below interfluve depths.

the next lowstand drainage to seek an alternative path to the south. As sea level falls below the elevation of the estuary-entrance shoals, the main trunk drainage of the Chesapeake Basin will shift to the south side of the chevron-shaped Middle Ground Shoal and drain through the evolving Chesapeake Channel. Here it will capture the flow of the York River, thus reconnecting the York as a tributary of the Susquehanna system.

CONCLUSIONS

Spits prograding into channels with subaerially exposed margins cause the channels to shift laterally. This spit forcing mechanism is necessary to maintain a dynamic equilibrium between channel size and channel flow. The rate of channel migration is related to the hydraulic instability caused by channel constriction. When longshore sediment drift causes a spit to prograde into the updrift side of a channel, the accelerated flow caused by the channel constriction increases the rate of sediment removal from the downdrift side of the channel. In a stratigraphic framework, the lateral migration path of the channel can be determined by tracking the scar and the overlying channel-fill facies. The fill sequence is recognized by large inclined beds that dip normal to the channel axis.

Channel shift by the spit-forcing mechanism is primarily controlled by the wave-driven longshore drift of sediment. Measurable lateral shifts can be made in 10^0–10^2 years and, therefore, the vertical changes in sea level produced by transgressions or regressions are minor. The buried record of channel migration is confined to relatively thin vertical intervals which are approximately equal to the relief of the migrating channel.

In the time framework of transgressive-regressive cycles (10^4–10^5 years), sea-level rise often exceeds the relief of a channel and water floods the adjacent interfluve areas. When this occurs, continuous channel migration is no longer necessary to maintain dynamic equilibrium of channel cross-sectional area and channels may be preferentially filled and abandoned. Ultimately, the channels may be capped by a down-drift migrating spit which, if transgression continues, will be partly removed by erosive shoreface retreat.

During Pleistocene highstands, the southern Delmarva Peninsula developed as a series of spits that progressively migrated across the outer part of a broad Chesapeake Seaway. During lowstands prior to the extension of the Delmarva Peninsula, the basin was a pathway for the Susquehanna, Patuxent, Potomac, Rappahannock, York, and James Rivers. Transgressions associated with oxygen-isotope Stages 13, 11, 9, 7 and 5e drowned the channels of these river systems forming estuaries on the floor of the Chesapeake Basin. Fine-grained estuarine sediments fill most of the lower portions of the Eastville and Exmore paleochannels and were also recognized in the Cape Charles paleochannel (MIXON, 1985). Subsequent highstands eventually flooded the paleo-interfluves between all of the river systems in the Chesapeake Basin, thus forming broad ancestral Chesapeake Seaways. As the southern Delmarva Peninsula advanced southward during each highstand event, the ancestral seaways became progressively more embayed and protected. Many of the larger fluvial channels remained partially open on the floor of the ancestral Chesapeake Bays, because filling by fine-grained estuarine sediments was slower than filling by spits and shoals in entrance areas. Channel fill was most rapid in bay-entrance areas where coarse-grained sediments from spits, shoals and the inner shoreface spilled into the open depressions. During the initial stages of transgression in the area, channel margins were still subaerial or near sea level, and spits of the prograding Delmarva Peninsula may have forced limited southward channel migration. Continued sea-level rise submerged the channel margins, relieving some of the hydraulic pressure on the south sides of the channels. Continued sea-level rise caused flooding of ancestral Chesapeake Bays and drowning of fluvial channels on the floors of ancestral Chesapeake seaways. Once bays formed, channels were no longer forced laterally but began to fill with estuarine and spit-platform deposits. On the south side of the ancestral bays, submerged channels appear to have been more effectively flushed by inlet currents. During subsequent regressions, falling sea levels re-exposed interfluve areas on the Chesapeake Bay floor and rivers were forced to flow seaward through the unfilled, tidally flushed bay-floor depressions (time-lagged channel shift). The time-lagged channel-shift mechanism is comprised of transgressive paleochannel fill, interfluve submergence, tidal-channel displacement and lowstand river diversion. The mechanism requires complete transgressive-regressive cycles that occur at frequencies of 10^4 to 10^5 years.

An important implication of this model is that the seismic stratigraphic record beneath the

Chesapeake Bay is considerably more complex than has been proposed to date and the stratigraphic evolution of the southern Delmarva Peninsula involved more than two highstand spit progradational events. The early Chesapeake Basins had up to six major fluvial systems draining through a broad entrance area. Progressive stream capture following highstand events has constricted the outer part of the basin producing progressively fewer drainways.

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