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Birth of the Modern Chesapeake Bay Estuary Between 7.4 and 8.2 Ka and Implications for Global Sea-Level Rise

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Birth of the modern Chesapeake Bay estuary between 7.4 and 8.2 ka and implications for global sea-level rise

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Abstract Two major pulses of sea-level rise are thought to have taken place since the last glacial maximum – meltwater pulses (mwp) 1A (12 cal ka) and 1B (9.5 cal ka). Between mwp 1B and about 6 cal ka, many of the complex coastal ecosystems which ring the world's oceans began to form. Here we report data for rhenium, carbon isotopes, total organic carbon, and fossil oysters from Chesapeake Bay which span the transition from fresh to brackish water conditions in the bay in the mid-Holocene. These data constrain sea-level change and resulting environmental change in the bay. They indicate that the transition was rapid, and that it was produced by (1) a third pulse of rapid eustatic sea-level rise, or (2) a geometry of the prehistoric Chesapeake Bay basin which predisposed it to a nonlinear response to a steadily rising sea level. Similar nonlinear changes in vulnerable coastal environments are likely to take place in the future due to polar warming, regardless of the timing or rate of sea-level rise.

Introduction

Based on fossil corals (Lighty et al. 1982; Fairbanks 1989) and the initiation of formation of modern deltas (Stanley and Warne 1994) and other coastal features, such as salt marshes and barrier islands, the rapid rate of eustatic sea-level rise during the early to middle Holocene appears to have slowed considerably by 6 cal ka, to about 1.5 mm year⁻¹. Details of sea level during the

transition interval from rapid rise to slow rise, however, are not as well constrained as other parts of the Holocene record. For example, most peat underlying modern salt marshes which are used as proxies for sea level postdate this time; older, submerged or buried peat is commonly difficult to find, hard to sample, and hard to distinguish from freshwater peat. Constraining the details of the prehistoric transition in the rate of sea-level rise has some urgency, however, because the near-stillstand since about 6 cal ka may not be the ultimate highstand of the current interglacial; future, abrupt rises appear entirely possible (Oppenheimer 1998).

In this study we examine long sediment cores collected from shallow and deep sites in the Chesapeake Bay estuary which are expected to record the transition interval in sea-level rise. We combine well-dated profiles of sediment geochemistry from these sites with bathymetric, sedimentological, and faunal data. With these data sets we reconstruct a continuous history which spans the period of 9 to 6 cal ka and clearly shows the transition from fluvial fresh to consistently brackish conditions – i.e., the birth of the modern Chesapeake Bay estuary.

Background

Chesapeake Bay is the largest estuary in the United States. It has a surface area of 11,100 km² and is over 300 km long, with an average width of 20 km. Like other coastal-plain estuaries, Chesapeake Bay formed during the later stages of retreat of the continental ice sheets of the last glaciation. The ocean then flooded the paleo-river valley of the Susquehanna River and its tributaries which had been exposed during the glacial lowstand (Colman and Mixon 1988; Colman et al. 1990). The average depth of Chesapeake Bay is 8 m but an axial channel, a remnant of the old river valley, runs down the center of the bay, reaching depths of 54 m (Fig. 1). The work described here is part of a larger sediment coring effort begun in 1996 to complement

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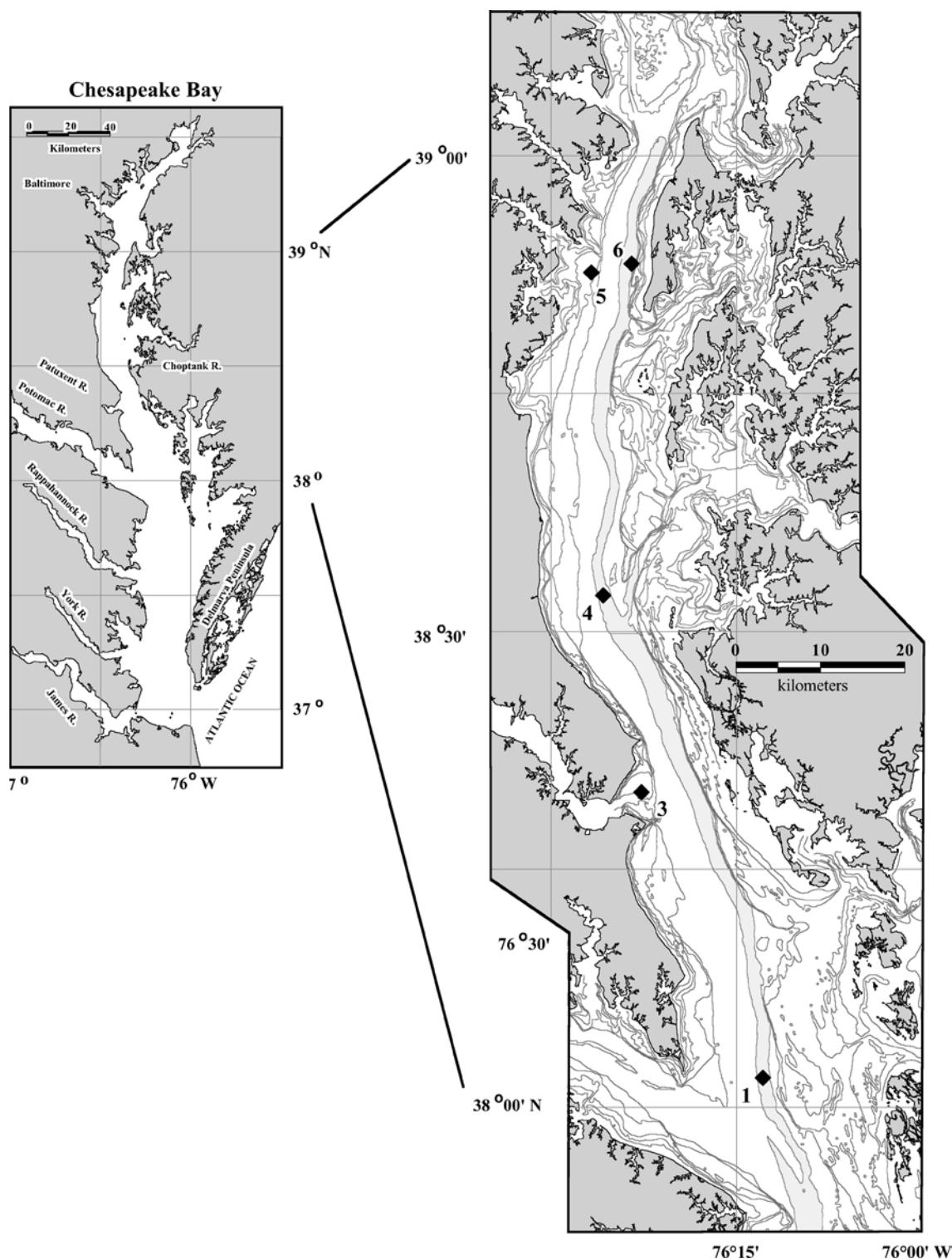
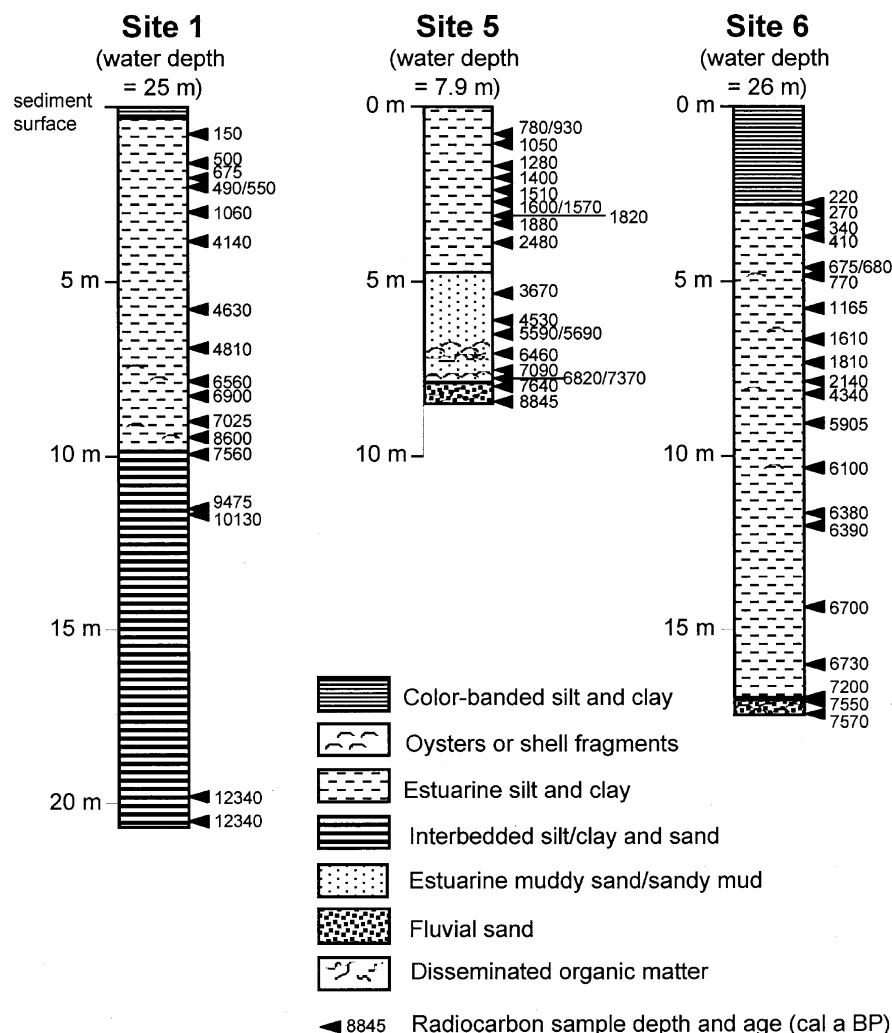


Fig. 1 Map showing Chesapeake Bay bathymetry and locations of sediment coring sites. Bathymetric contours are in feet (6, 12, 18, 30, 36, and 60; 1.8, 3.7, 5.5, 9.1, 11.0, and 18.3 m, respectively); the 60-foot contour (>18.3 m) corresponds to the approximate boundary of the freshwater Chesapeake prior to 8.2 cal ka. Bathymetric data are from NOAA. Data from coring sites 3 and 4 are not included in this discussion but the numbering system is maintained for consistency with Colman et al. (2002)

earlier geophysical investigations of Quaternary sedimentation in the bay (Colman and Hobbs 1987, 1988; Colman and Halka 1989a, 1989b). Preliminary reports of coring results have included Kerhin et al. (1998), Cronin (2000), and Baucom et al. (2001).

The focus of the investigation of long sediment cores presented here (Fig. 2) is geochemical and paleo-environmental aspects of the transition interval from fluvial

Fig. 2 Core lithologies and radiocarbon sample depths and ages (cal a B.P.)



to estuarine conditions in Chesapeake Bay. Typical estuarine sediments recovered in cores consisted of dark gray to black silt and clay with varying amounts of fine to medium sand and shells, especially shells of the opportunistic bivalve *Mulinia lateralis*, as well as *Crasostrea virginica* and *Mercenaria mercenaria*. The shallow-water core (site 5) was generally coarser grained and more bioturbated than the cores from the deep axial channel (sites 1 and 6).

Methods

Sediment cores were collected in 1998 and 1999 at the locations shown in Fig. 1. One core (site 5, 7.9-m modern water depth) was collected from a dredging barge using a crane to deploy a modified Mackereth-type piston corer configured to penetrate and recover as much as 12 m of sediment. The other two cores (site 1 and site 6, at 25- and 26-m water depths, respectively) were obtained from the R/V *Marion Dufresne* with its Calypso piston corer configured to sample >20 m of sediment. Coring operations and the physical properties

of the sediments are described in more detail by Cronin (2000) and Baucom et al. (2001); basic core lithologies are shown on Fig. 2. Cores were collected in fiberglass or PVC liners and they were cut into 1.5-m lengths and capped immediately after retrieval of the corer. Refrigerated cores were split lengthwise and bulk-sampled in either 2- or 3-cm slices. Subsamples for geochemical analysis were collected from discrete depths immediately prior to bulk sampling. Spacing of analyzed samples was either 3 or 10 cm. Samples were dried at 60°C and crushed with a mortar and pestle. Isotope samples were then acidified with 10% HCl to remove any carbonate, dried again, and disaggregated.

Carbon bulk and isotopic compositions were determined using a Carlo Erba NC2500 elemental analyzer coupled to a Finnigan Delta Plus isotope ratio mass spectrometer at the USGS facilities in Reston, Virginia. Calibration for the total organic carbon (TOC) determinations was accomplished using the National Research Council of Canada Certified Reference Material PACS-2, a marine sediment with values of 3.32 mass% C and 0.27 mass% N. Carbon-isotope compositions were measured on CO₂ and are reported in the δ nota-

tion, as permil (‰) relative to Vienna Pee Dee Belemnite (VPDB). The $\delta^{13}\text{C}$ values are calibrated on a scale defined by USGS-24 (graphite) with $\delta^{13}\text{C} = -15.9\text{‰}$. Analytical precision is estimated to be $\pm 0.2\text{‰}$ (1σ).

For authigenic rhenium analyses, dried and powdered sediment samples (~ 0.1 g) were leached with 1.5 ml 4 N Seastar HNO_3 for one hour at room temperature to extract authigenic rhenium. Sample vials were centrifuged and the supernatant was filtered using dedicated syringes and 0.45- μm cartridge filters. The leachate was combined with 0.1 g of 1.8 ppb ^{187}Re -enriched spike solution, diluted with Milli-Q water 3–6 \times , and transferred to autosampler vials. Samples were analyzed by inductively-coupled plasma mass spectrometry

(ICP-MS) using a Finnigan Element 2 instrument equipped with a Cetac Technologies MCN6000 desolvating nebulizer and an autosampler. Original sample concentrations were calculated from measured ratios of ^{187}Re : ^{185}Re by the standard isotope-dilution formula.

Radiocarbon ages indicated for cores 1, 5, and 6 (Fig. 2) were measured by accelerator mass spectrometry (Colman et al. 2002). Ages were corrected for secular variation using CALIB 4.1 (Stuiver et al. 1998). Carbonate sample ages were reservoir corrected using a standard marine reservoir age of 400 years verified for Chesapeake Bay by analyses of archived oyster shells of known age (Colman et al. 2002). Calibrated radiocarbon ages are reported using “cal a” or “cal ka” notation,

Table 1 Chesapeake Bay and Delaware radiocarbon data

Depth below mean sea level (m)	^{14}C age	Analytical error	Cal a B.P. ^a	Cal error + (older)	Cal error– (younger)	Medium
Site 5, core RR98-9 (Colman et al. 2002)						
–14.4	5,340	40	5,690	100	90	Oyster
–15.0	6,060	55	6,460	160	110	Oyster
–15.9	7,180	40	7,640	60	70	Oyster
Site 6, core MD99-2209 (Colman et al. 2002)						
–43.2	6,660	45	7,200	120	80	Oyster
Chesapeake Bay entrance (Harrison et al. 1965)						
–21.0	8,135	160	9,073	362	442	Oyster
Core MD99-2208 (Vogt et al. 2000)						
–17.8	7,740	55	8,505	95	95	Oyster
Tangier Sound (Hobbs 1988)						
–15.3	7,390	115	8,195	194	196	Wood
James Estuary (Ellison and Nichols 1976)						
–2.5	1,310	160	1,210	325	286	Basal peat
–5.0	2,700	160	2,819	350	461	Basal peat
–7.3	4,880	140	5,621	291	304	Basal peat
Rappahannock Estuary (Ellison and Nichols 1976)						
–0.9	535	95	552	118	176	Basal peat
–1.3	980	90	879	181	178	Basal peat
–4.5	3,345	120	3,589	286	247	Basal peat
–5.7	2,810	160	2,958	405	251	Basal peat
–9.2	5,780	210	6,595	460	422	Basal peat
Delaware radiocarbon database (Ramsey and Baxter 1996)						
–1.2	1,650	70	1,547	160	160	Basal peat
–2.5	2,070	80	2,026	158	158	Basal peat
–2.6	1,770	60	1,685	138	138	Basal peat
–2.8	2,560	95	2,575	220	221	Basal peat
–3.1	2,260	70	2,208	158	159	Basal peat
–3.6	2,420	95	2,528	220	220	Basal peat
–3.6	3,270	70	3,500	140	140	Basal peat
–3.8	1,950	55	1,865	129	129	Basal peat
–3.9	2,820	70	2,929	153	153	Basal peat
–4.7	2,880	70	3,029	184	185	Basal peat
–5.0	2,490	80	2,549	190	190	Basal peat
–5.2	2,945	95	3,104	240	241	Basal peat
–5.3	3,314	63	3,522	129	130	Basal peat
–6.3	3,780	170	4,131	443	443	Basal peat
–6.4	3,430	70	3,660	182	183	Basal peat
–6.4	3,980	105	4,398	256	256	Basal peat
–7.7	4,090	100	4,596	246	246	Basal peat
–10.9	4,490	80	5,093	223	223	Basal peat
–12.8	5,600	110	6,424	241	241	Basal peat
–13.4	5,345	110	6,110	207	207	Basal peat
–14.7	6,170	115	7,022	244	245	Basal peat
–20.8	7,500	135	8,243	252	253	Basal peat
–26.3	10,800	300	12,561	851	851	Basal peat

^aAges and 2σ errors recalibrated using CALIB 4.3 for all data except from Colman et al. (2002), and Delaware basal peat

indicating years or thousands of years, respectively, before 1950 A.D. Other literature ages (Table 1) are either (1) reported in the notation format in which they were published (cal a B.P.), or (2) recalibrated in the case of older ages using CALIB 4.3 and reported as “cal a” or “cal ka” also. For recalibration of ages without reported $\delta^{13}\text{C}$ values, estimated values of 0, -15, and -25 were used for shell, salt-marsh peat, and wood samples, respectively.

Results

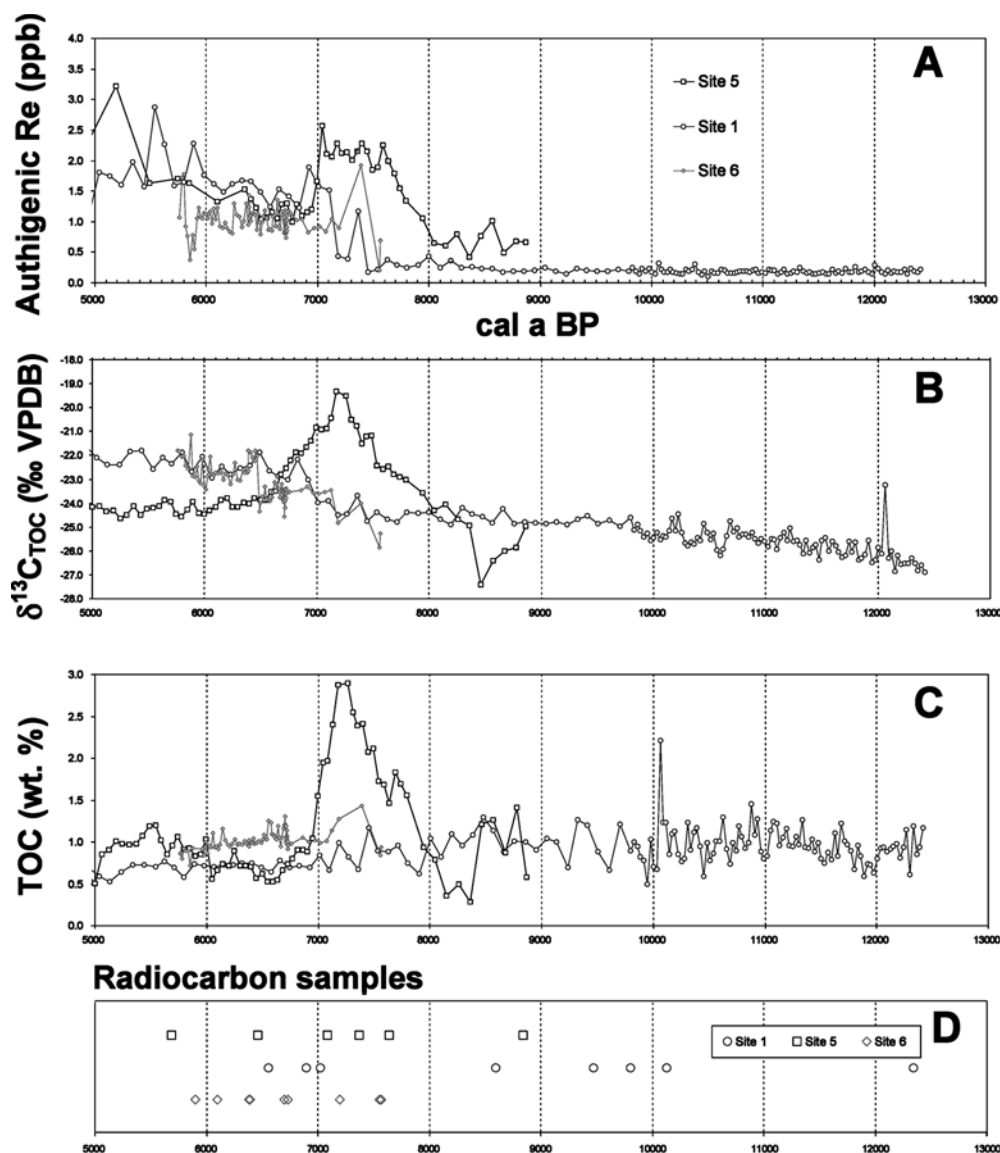
Geochemical data show consistent timing and patterns of environmental change in the paleo-Chesapeake in spite of wide separation of core sites (~100 km). Profiles of authigenic rhenium, organic carbon isotopes, and TOC for the three long cores (8 to 21 m) are shown in Fig. 3A–C. Age and mass accumulation models for the three cores were developed from extensive radiocarbon

analyses (Fig. 2, $n = 23$ between 5 and 13 cal ka; Colman et al. 2002). Here we focus on the interval from 6.5 to 8.5 cal ka.

One geochemical indicator which is sensitive to salinity changes is the conservative trace element rhenium. Rhenium is highly enriched in seawater relative to crustal abundances. Concentrations of dissolved Re in river waters are typically at least an order of magnitude lower than in the ocean (Colodner et al. 1993). Rhenium is a conservative element in seawater, meaning that its concentration is consistent from basin to basin and that its concentration varies linearly with salinity. In sediments, authigenic rhenium content is controlled by three factors in the Chesapeake system: salinity of overlying waters, oxygen concentration and associated redox state of sediments, and contributions from erosion of rhenium-rich Miocene deposits in the area.

Authigenic rhenium in dried Chesapeake Bay sediments rose from low values around 0.2 ± 0.1 ppb (Fig. 3A), indicating deposition in freshwater conditions

Fig. 3A–D Plots of **A** authigenic rhenium, **B** $\delta^{13}\text{C}_{\text{org}}$, **C** TOC, and **D** radiocarbon samples for sites 1, 5, and 6



(Colodner et al. 1993), to higher but fluctuating values around 1.3 ± 1.0 ppb, indicating deposition in estuarine to marine conditions. This transition in the site 1 core occurred abruptly between 7.1 and 7.5 cal ka, after 5,000 years of low and stable values prior to this rise. At site 5 values rose more gradually, starting at about 8 cal ka, and reached a plateau at 2.1 ppb by 7.6 cal ka. They dropped at about 6.9 cal ka to values similar to those recorded in the other cores, and then rose again later. The bottom of the site 6 core just reaches the transition interval. The rhenium pattern is similar to the site 1 pattern of an abrupt rise, although there are only two pre-rise or early-rise data points. The values following the rise are lower than at site 1, consistent with the expected lower salinity at this more upstream site.

Measured $\delta^{13}\text{C}$ of sedimentary organic matter from two cores shows a change from typical terrestrial or fluvial values around -25‰ VPDB to more marine planktonic values of -22‰ between 7.6 and 6.5 cal ka. In one core (site 5) the value rose to a high of -19‰ at 7.1 cal ka, followed by stabilization around -24‰ by 6.5 cal ka. A coincident peak in TOC (3%) is also seen in this core. No comparable $\delta^{13}\text{C}$ or TOC anomalies are seen in either of the cores from deeper water (sites 1 and 6).

Discussion

The geochemical results described above indicate clearly that a rapid transition from fresh to brackish conditions took place in Chesapeake Bay between 7.4 and 8.2 cal ka. The specific mechanism, however, which produced this initial change is less obvious. The cause of the apparent, subsequent return to progressively fresher conditions in the bay, based particularly on fossil oyster occurrences described below, also requires an explanation consistent with the first mechanism and with other data sets of eustatic sea level.

The geochemical proxies measured at the three core sites provide independent evidence of the timing and nature of flooding of the paleovalley which became Chesapeake Bay. The clearest evidence comes from the rhenium data in the deeper water cores (sites 1 and 6). In these locations, particularly at site 1, we interpret the rhenium signal as being predominantly driven by salinity changes, with an abrupt shift to higher salinity at around 7.5 cal ka. This is also reflected to a lesser extent by the $\delta^{13}\text{C}$ data, where the transition begins at around the same time but proceeds more gradually.

Rhenium data for core site 5 are less clear, however, and appear to show a primary salinity signal modified by the reducing conditions developed in organic-rich sediments during an interval of salt-marsh deposition between 7.9 and 6.6 cal ka (see discussion below). Another complicating factor which cannot be ruled out at site 5 is the addition of reworked rhenium from older Miocene sediments. The essential absence of reworked organic carbon, as indicated by consistent radiocarbon ages and

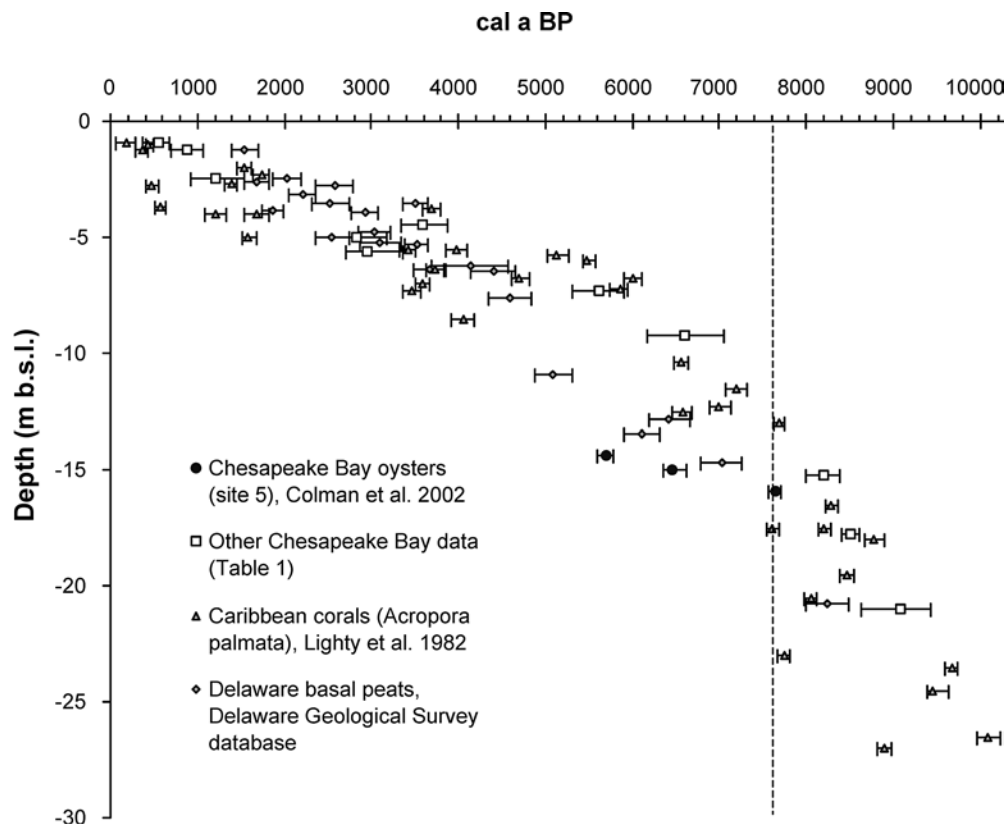
a relatively small offset between carbonate and bulk TOC ages (Colman et al. 2002), however, suggests that substantial oxidation of Miocene sediments took place during reworking; this would also strip residual authigenic rhenium prior to redeposition. In addition, rhenium levels in freshwater sediments deposited before the change to brackish water (site 1 older than 8 ka) are extremely uniform in spite of changes in sedimentation rate. This indicates that the leachable contribution from freshwater authigenic and/or reworked rhenium is low (average = 0.19 ppb, $n=110$) and constant (standard deviation = 0.04 ppb).

As described above, the core from site 5 shows a transient $\delta^{13}\text{C}$ peak as heavy as -19‰ centered at 7.1 cal ka. The organic carbon in these sediments is more enriched in ^{13}C than even fully marine particulate organic carbon. This suggests contribution of a source of carbon which is isotopically unusually heavy. One likely source would be salt-marsh vegetation reworked and deposited as the intertidal zone passed through the paleo-elevation of the core site during inundation. Due to physiology which excludes salt, biomass from salt-marsh plants such as *Spartina alterniflora* has been shown to have $\delta^{13}\text{C}$ values of -13 to -10‰ (Haddad and Martens 1987). The decline of the $\delta^{13}\text{C}$ peak starting at 7.1 cal ka and continuing to around 6.6 cal ka is consistent with drowning or transgression of the marsh with rising sea level. Morris et al. (2002) suggest that salt-marsh drowning typically occurs when sea-level rise exceeds 12 mm year^{-1} .

The $\delta^{13}\text{C}$ peak at site 5 coincides with a TOC peak (3%) about three times higher than the typical value in all cores (1%). This is also consistent with proximity to a salt marsh during deposition, even though peat as such was not recovered. Additionally, the site 5 core shows greater sedimentological variability over its length than either of the deep-water cores, consistent with its location relative to the shore. The $\delta^{13}\text{C}$ peak occurs within a sedimentological interval in the core (Fig. 2) interpreted as having been deposited in restricted estuarine conditions (Baucom et al. 2001); such an interval is not well developed in the site 1 and site 6 cores.

In the site 5 core, the interval of the $\delta^{13}\text{C}$ excursion is underlain by articulated oyster shells dated at 7.6 cal ka (Baucom et al. 2001; Colman et al. 2002) at a depth of about 16.5 m below modern sea level (Figs. 2 and 3B). The oysters overlie coarse fluvial sand and gravel. This oyster-on-sand contact marks the initial transition from fresh to brackish conditions at this site. The $\delta^{13}\text{C}$ peak in the site 5 core is bracketed above (15 m below sea level, m b.s.l.) by less concentrated oysters which give an age of 6.5 cal ka (Fig. 2). As part of another study, a similar core was collected from a buried paleochannel about midway between sites 1 and 5 in 10 m of water. This core produced a basal oyster age of 8.2 cal ka for a bed at 17.8 m b.s.l. (Fig. 4; Vogt et al. 2000). The basal age of the site 6 core, which penetrated the upper few centimeters of an oyster bed, is 7.6 cal ka at 43 m b.s.l. (Fig. 2).

Fig. 4 Comparison of sea-level data from Chesapeake Bay site 5 with data derived from corals (Lighty et al. 1982, ages recalibrated), basal salt-marsh peat from the Delaware radio-carbon database (Ramsey and Baxter 1996; elevations adjusted to depth below mean sea level using Lewes, Delaware tide gauge datum information), and other published Chesapeake Bay data (Table 1). The *dashed line* corresponds to the timing of the rapid sea-level rise event (CRE 3) proposed by Blanchon and Shaw (1995)



Extensive areas of fossil oyster beds older than 3.5 ka are present in northernmost Chesapeake Bay and in upstream areas of a major tributary, the Potomac River (Callender et al. 1984; C. Judy, written communication). Salinities are too low in these areas for modern oysters to survive (< 11 ppt), indicating that saltwater intruded farther up into the estuary and its tributaries in the past than it does at present.

The location of the estuarine mixing zone can be dated and mapped using occurrences of fossil oysters such as those in Chesapeake sediments. Other estuaries, such as Mobile Bay (Ryan and Goodell 1972) in Alabama and the Hudson River estuary (Bell et al. 2000) in New York, show similar patterns of upstream occurrence of fossil oysters (> 5.7 and > 6.2 ka, respectively). Younger fossil oysters are found progressively oceanward in Chesapeake Bay and the other estuaries, tracking retreat of the mixing zone as fluvial sediments filled the estuary heads during the period of relatively slow sea-level rise (< 3 mm year $^{-1}$) which began around 6 ka.

Eustatic sea-level change is an obvious driver of transition of coastal river valleys to estuaries during the mid-Holocene. Sea level can be constrained from Caribbean coral and basal salt-marsh peat from Delaware (Fig. 4), along with Chesapeake Bay oyster and salt-marsh data spanning the interval of interest. Anomalous high subsidence reported for the Chesapeake Bay region based on tide gauge data appears to be a relatively recent phenomenon. Age-depth data for 7–8 ka

oyster and peat deposits in and around the bay do not generally indicate measurable depression below expected depths based on non-Chesapeake sea-level curves.

Published data from other Chesapeake studies constrain sea level in areas beyond our immediate coring region (Fig. 4, Table 1). Hobbs (1988) dated a piece of wood from a long core from the eastern side of the Virginia part of the bay (approximately 37° 47.7'N, 75° 51.7'W) and obtained a ^{14}C age of 7.4 ka. The sample was collected from below a fossil oyster reef at a depth of 15.25 m b.s.l., 15 cm above the basal Holocene unconformity. Ellison and Nichols (1976) cored and dated peat in several Chesapeake Bay subestuaries, including an unusually thick salt-marsh peat in the Rappahannock Estuary (Figs. 1 and 4, Table 1). They obtained a radiocarbon age of 6.6 cal ka for a basal peat collected from 9.2 m below the marsh surface. Microfossils from the site 1 core at depths up to 38 m b.s.l. indicate several, brief episodes of temporary saltwater incursion may have taken place, starting as early as about 10 cal ka (Willard et al. 2000), possibly produced by storm surges. Sediments containing freshwater diatoms overlie each of these early brackish intervals.

The pattern of rapid advance of oysters upstream followed by gradual retreat, as well as the abrupt changes in proxy salinity indicators can be explained in several ways. One scenario which is consistent with these observations is the occurrence of a severe and extended drought. Decreased discharge of freshwater from tributaries would cause intrusion of brackish water into areas

which were usually fresh. Although such a scenario is possible, the drought would need to be of considerable duration to maintain the situation; otherwise, the system would eventually revert to fresher conditions as the drought ended. No independent evidence of such a drought has yet been reported. Data of Webb et al. (1993) indicate a general increase in precipitation and runoff over the Holocene for the northeast U.S. This could explain the seaward retreat of oysters instead of, or in combination with, sediment infilling at the head of the estuary and in tributaries. Increasing precipitation alone, however, is unlikely to have totally surpassed sea-level rise in Chesapeake Bay.

A second explanation for the Chesapeake data is that sea level rose monotonically ($< 10 \text{ mm year}^{-1}$) but that the sea level crossed a break in slope (from steeper to flatter) in the topography being flooded. This would produce acceleration in both areal and upstream extent of inundation with no change in absolute rate of rise. In combination with the topographic threshold, the system would have also surpassed a hydrographic threshold. Saltwater intrusion into an estuary is a function of both freshwater discharge into the estuary and cross-sectional area of the estuary (Savenije 1993). The expansion of the nascent Chesapeake beyond the confines of the narrow Cape Charles paleovalley of the Susquehanna River (Colman et al. 1990) would have markedly increased the cross-sectional area of the estuary without changing freshwater discharge. Crossing this hydrographic threshold would have resulted in extensive and sustained upstream intrusion of saltwater. The oceanward retreat of oysters observed during continued but more gradual rising of sea level since that time indicates that the system was re-establishing near-equilibrium conditions which were swamped by an initial inundation.

All Chesapeake cores described here show that the shift to brackish conditions around 7.4 to 8.2 cal ka was (1) virtually simultaneous (given uncertainties in measurement, calibration, and reservoir correction of radiocarbon ages) over an upstream distance of at least 100 km, and (2) permanent. The sustained transition at site 1, based on geochemistry (Fig. 3) and microfossils (Willard et al. 2000), began around 7.4 cal ka, or 35 m b.s.l. Oysters overlying the basal Holocene unconformity or fluvial sediments in two cores gave similar ages – site 5 = 7.6 cal ka, 16.5 m b.s.l.; core MD99-2208 = 8.2 cal ka, 17.8 m b.s.l. (Vogt et al. 2000). The base of a third core (site 6) which penetrated the top of an oyster bed presumed to overlie the basal Holocene unconformity or early Holocene fluvial deposits gave an age of 7.6 cal ka.

The site 6 basal oysters are approximately the same age as the other two basal oysters from shallow-water cores, but they are not interpreted as sea-level indicators because they are found approximately 43 m b.s.l. or 26 m below the level of oysters in the other cores. This 26-m depth range for approximately contemporaneous oysters indicates that care must be taken in using oysters from a setting like Chesapeake Bay as sea-level

indicators. Taken in the chronological and vertical context of other sea-level data (Fig. 3), it is clear that the site 6 oysters cannot have lived close to sea level. It must also be true, however, that shallower oysters from the other sites could have lived at some depth below sea level. So, in the most conservative interpretation, they can only be used to constrain lower limits of sea level at the time of their formation. The fact that the transition to brackish conditions in the deep-water cores (site 6 and site 1) did not take place much earlier than the transition in the shallow-water cores presents an interesting problem.

The permanent fresh-brackish transition appears to have been delayed until sea level rose gradually beyond about 18 m below modern sea level. This corresponds to both the transition depth at site 5 and the depth of the modern rim of the relict axial basin in the Chesapeake (Fig. 1). Because the thalweg of the Susquehanna paleovalley lies at more than 50 m b.s.l. in Chesapeake Bay, this implies that a physical or hydrological barrier must have been present to restrict the influx of saltwater until after 8.2 cal ka. There is no evidence of an early to mid-Holocene sill between the modern bay mouth and site 1. Therefore, these data support the second explanation above – sea-level rise passing a topographic and hydrographic threshold – for the delay in transition to brackish conditions. The narrow, pre-8.2 cal ka Chesapeake would have functioned as a fjord-like and tidal-fresh extension of the Susquehanna River until the critical sea level was reached.

A final explanation for the advance-retreat pattern in oyster beds and the sharp geochemical transition is that they were produced by a sudden (< 200 years) and large (> 4 m) jump in sea level – similar to that postulated from Caribbean relict-reef data by Blanchon and Shaw (1995). Larcombe et al. (1995) reported evidence of a sea-level rise of 30 mm year^{-1} around the same time (post-8.2 cal ka) in Australia. Most coral-derived records of sea level from the Pacific Ocean, however, do not show this event (e.g., Tahiti, Bard et al. 1996, Montaggioni et al. 1997; Papua New Guinea, Chappell and Polach 1991), although the larger depth ranges of Pacific reef-crest corals may not be sensitive enough to register an event of less than 6 m (Blanchon 1998; but see Montaggioni and Bard 1998).

Recently, Blanchon et al. (2002) have reported additional discoveries of relict reefs and submerged wave-cut notches in the Caribbean which support and refine previous interpretations of a meter-scale jump starting at 7.6 cal ka. A rapid-rise event at this time, although not necessarily as large or as rapid, is also consistent with data from the inner Atlantic shelf of the U.S. reported by Thieler et al. (1999). Other indications of rapid or coincident sea-level rise may include catastrophic flooding of the Black Sea by overtopping of the Bosphorus spillway (“Noah’s Flood”, 7,150 ^{14}C a B.P.; Ryan et al. 1997; Ryan and Pitman 1998; but see also Aksu et al. 2002), more gradual flooding of Ancyclus Lake to form the Litorina/Baltic Sea (7,500 to 7,300 ^{14}C a B.P.,

Bianchi et al. 2000; about 8,350 to 7,860 ^{14}C a B.P., Sohlenius et al. 2001), and sudden transition from restricted to open estuarine conditions in Galveston Bay around 7.2 cal ka (Anderson et al. 2001). An anomalously rapid rise is not required by the Chesapeake Bay data, but neither is it precluded.

Conclusions

Geochemical and oyster evidence from both deep-water and shallow-water sediment cores constrain the timing of the fresh-brackish transition in northern Chesapeake Bay to between 7.4 and 8.2 cal ka. This transition took place when sea level rose beyond about 18 m below modern sea level, consistent with Caribbean coral data and Delaware basal peat data. The delay in saltwater intrusion into the bay may imply the existence of a topographic and hydrographic threshold which was not exceeded by previous sea-level rise. Alternatively, the sudden change could be accounted for by a rapid jump in sea level. It is not possible to differentiate between these two mechanisms with the present data but additional, detailed coring studies in Chesapeake Bay and other estuaries could distinguish between the two scenarios.

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