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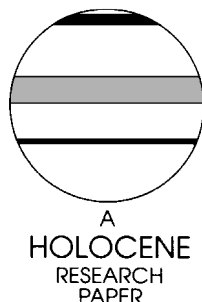
Late-Holocene climate and ecosystem history from Chesapeake Bay sediment cores, USA

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Abstract: Palaeoclimate records from late-Holocene sediments in Chesapeake Bay, the largest estuary in the USA, provide evidence that both decadal to centennial climate variability and European colonization had severe impacts on the watershed and estuary. Using pollen and dinoflagellate cysts as proxies for mid-Atlantic regional precipitation, estuarine salinity and dissolved oxygen (DO) during the last 2300 years, we identified four dry intervals, centred on AD 50 (P1/D1), AD 1000 (P2/D2), AD 1400 (P3) and AD 1600 (P4). Two centennial-scale events, P1/D1 and P2/D2, altered forest composition and led to increased salinity and DO levels in the estuary. Intervals P3 and P4 lasted several decades, leading to decreased production of pine pollen. Periods of dry mid-Atlantic climate correspond to 'megadroughts' identified from tree-ring records in the southeastern and central USA. The observed mid-Atlantic climate variability may be explained by changes in atmospheric circulation resulting in longer-term, perhaps amplified, intervals of meridional flow. After European colonization in the early seventeenth century, forest clearance for agriculture, timber and urbanization altered estuarine water quality, with dinoflagellate assemblages indicating reduced DO and increased turbidity.

Key words: Palaeoclimate, climatic change, estuarine sediments, drought, pollen, dinoflagellate cysts, anthropogenic impacts, Chesapeake Bay, late Holocene.

Introduction

The combined influence of natural climatic variability and human-induced ecosystem degradation introduces complexities for management and restoration of coastal ecosystems of the world. Whereas many studies of coastal ecosystems have focused on impacts of cultural eutrophication (Jorgensen and Richardson, 1996), land-use changes and sediment influx (Howarth *et al.*, 1991) and hydrological changes (Swart *et al.*, 1996a; 1996b), growing evidence suggests that estuarine ecosystems are also influenced by climatic variability (i.e., Peterson *et al.*, 1995; Cronin *et al.*, 2000). With few exceptions, separating anthropogenic effects from those due to climatic causes has remained problematic due in part to the limited historical record available for most ecosystems (e.g., Johannessen and Dahl, 1996a; 1996b; Gray and Abdullah, 1996; Swart *et al.*, 1996a; 1996b; Halley and Roullet, 1999).

Sedimentary records are useful to separate human impacts on ecosystems from those due to climatic variability and other 'natural' factors. For example, Nagy and Alve (1987) and Alve (1991) were able to distinguish the impact of both climate and pollution in Oslo and Drammens Fjords in Norway on the basis of foraminifera from sediment cores.

In the current paper, we examine the late Holocene (2300 yr BP to present) record of Chesapeake Bay and the adjacent terrestrial ecosystem in its watershed through the study of fossil dinoflagellate cysts (henceforth referred to as dinocysts) and pollen from sediment cores. These cores preserve high-resolution records of both late-Holocene palaeoclimate and ecosystem changes since European colonization (henceforth referred to as colonization) beginning in the seventeenth century. A number of changes in the bay and its watershed have been attributed to large-scale eighteenth- and nineteenth-century land clearance (Brush, 1984; DeFries, 1986) and twentieth-century nutrient influx (Cooper and Brush, 1991; Karlson *et al.*, 2000; Zimmerman and Canuel, 2000). However, prior studies of Chesapeake Bay have limited data for the period pre-dating colonization and low temporal resolution since colonization. A thick (10 m to >20 m) Holocene sequence of sediments dated by radioisotopes (Cronin *et al.*, 1999; Colman *et al.*, 2002) provides an excellent record of climatic variability over decadal (Cronin *et al.*, 2000) to millennial (Willard and Korejwo, 2000) timescales.

Our two primary goals in this study are to use terrestrial (pollen) and estuarine (dinocyst) indicators to determine the regional palaeoclimate history of the mid-Atlantic region over the past 2300 years and to test the hypothesis that anthropogenic factors in the Chesapeake watershed have had a predominant

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influence on the functioning of the bay's ecosystem since the seventeenth century. To our knowledge, this is the first high-resolution palaeoenvironmental study from estuarine sediments integrating data from both terrestrial and estuarine indicators to separate anthropogenic impacts from those caused by long-term natural climatic processes. The resulting details on ecosystem response to natural climatic variability provide a context for development of restoration targets for this greatly altered ecosystem.

Regional setting, the Chesapeake Bay watershed and estuary

Chesapeake Bay is 320 km long, covers an area of 6500 km², and drains a watershed of 166 000 km² (Figure 1). The modern bay formed between 9 ka and 7 ka when postglacial sea-level rise flooded a dendritic river system of the palaeo-Susquehanna River (Colman and Mixon, 1988; Willard *et al.*, 2000). The Chesapeake Bay watershed mainly occupies parts of Maryland, Virginia, the District of Columbia, Pennsylvania and New York. Forest composition and distribution in the watershed is controlled by climatic, geomorphic and physical factors, such as moisture availability, soil type and elevation. The bay itself lies within the Atlantic Coastal Plain, a region characterized by unconsolidated sediments, little topographic relief and dominance of oak-pine forests in the south and oak-chestnut and tulip poplar forests in the north (Braun, 1950; Brush *et al.*, 1980). The Piedmont Province, extending from the Fall Line in the east to the Catoctin Mountains in the west, has greater topographic relief and is underlain by igneous and metamorphic rocks. Vegetation in the relatively mesic eastern part of the province consists primarily of oak-

hickory forests, grading toward oak-dominated forests in the drier west (Braun, 1950; Brush *et al.*, 1980). The Valley and Ridge Province, to the west, is underlain by different lithologies with varying hydrologic characteristics. Generally, oak-chestnut forests dominate the dry ridges, and sugar-maple and basswood assemblages dominate the wetter sites in the valleys and in the northern part of the watershed (Brush *et al.*, 1980; Hack and Goodlett, 1960).

Chesapeake Bay is a partially mixed estuary with a strong north-south salinity gradient and a seasonal pycnocline formed by inflowing dense saline water from the Atlantic Ocean overlain by less saline water influenced by fresh water flowing into the bay from its tributaries. The major tributaries feeding the bay are the Susquehanna (48% of total inflow), Potomac (33%), James (13%), Rappahannock (3%) and Patuxent (1%) Rivers along the western shore and the Choptank (1%) and Nanticoke (1%) Rivers along the eastern shore (Schubel and Pritchard, 1986). Large seasonal and interannual variability in bay salinity, temperature and dissolved oxygen are caused by regional precipitation and river discharge from the watershed (Malone, 1991; Cronin *et al.*, 1999), nutrient dynamics (Boynton and Kemp, 1985), physical processes in the estuary (Boicourt, 1992) and other ecological processes (Smith *et al.*, 1992).

The Chesapeake Bay region lies in a jet stream transition zone (Figure 2), and shifts in the configuration of the jet strongly influence storm frequency and intensity, precipitation and temperature in the region (Vega *et al.*, 1998; 1999). In northern parts of the bay and watershed, precipitation is controlled primarily by atmospheric circulation patterns over the North Atlantic, with meridional flow patterns over the ocean resulting in decreased storm frequency over the eastern USA, decreased precipitation and increased bay salinity (Henderson-Sellers and Robinson, 1986). To a lesser extent, precipitation in the region is also influenced by events in the tropical Pacific Ocean, with meridional flow across the USA correlated slightly with decreases in precipitation (Vega *et al.*, 1999). The transitional position of the bay relative to the jet stream and the combined influence of North Atlantic and tropical Pacific Ocean events on the regional climate makes the Chesapeake Bay watershed a particularly sensitive area to small changes in atmospheric circulation patterns resulting from climatic variability.

Materials and methods

Core sites were chosen from mesohaline regions of the bay, which are sensitive to long-term changes in salinity (Cronin *et al.*, 2000) and dissolved oxygen (Cooper and Brush, 1991; Karlsten *et al.*, 2000; Zimmerman and Canuel, 2000). Three cores were taken during 1996 by the R/V *Discovery* (sites PTMC-3, PRCK-1, PTXT-2; Kerhin *et al.*, 1998; Cronin *et al.*, 1999), and a fourth core was taken on the IMAGES V cruise of the *Marion-Dufresne*

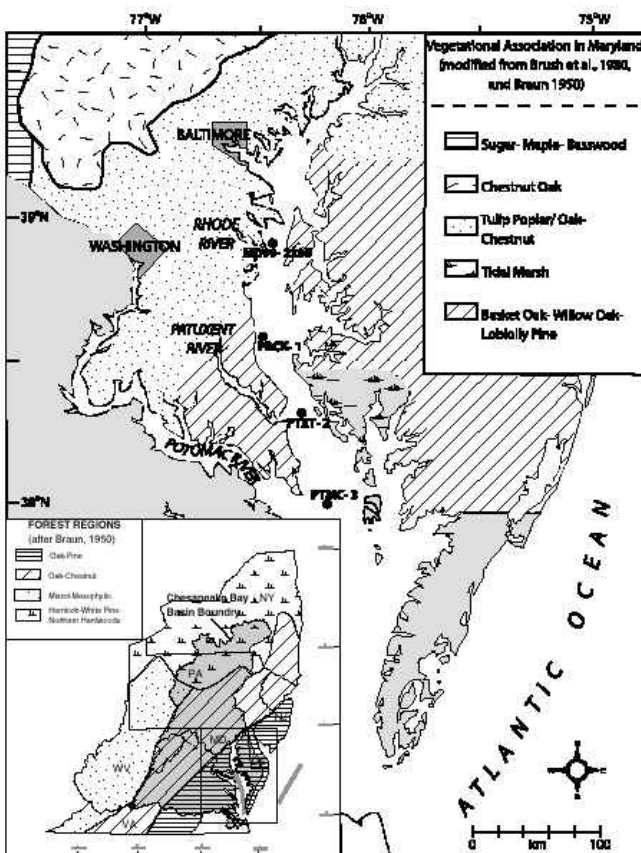


Figure 1 Inset map of Chesapeake Bay watershed boundaries and forest regions (after Braun, 1950); larger map shows coring sites in Chesapeake Bay and distribution of generalized forest types in Maryland (after Braun, 1950; Brush *et al.*, 1980).

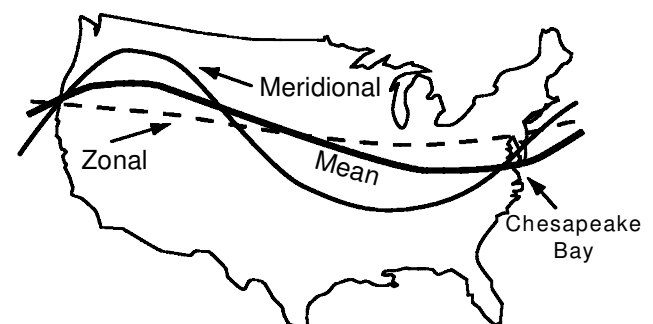


Figure 2 Simplified path of mean position of polar jet stream showing deviations toward meridional and zonal flow.

in 1999 (core MD99-2209; Chesapeake Shipboard Scientific Staff, 2000) (Figure 1; Table 1). The *Discovery* cores were taken with a 3-inch piston corer; the *Marion-Dufresne* core was taken with a modified Calypso piston corer.

Site selection was based on several criteria. First, geophysical data (CHIRP, side-scan sonar) indicate a thick Holocene sequence in these regions (Colman and Halka, 1989; Hagen and Vogt, 1999; Halka *et al.*, 2000; Vogt *et al.*, 2000). Second, these sites cover an area of strong depth- and latitude-controlled gradients in salinity and dissolved oxygen (DO). Two sites (MD99-2209 and PTMC-3) represent deep-channel (25 m water depth) locations, influenced by seasonal hypoxia. Cores from sites PRCK-1 and PTXT-2 were collected in shallower water (~11 m water depth) on the flanks of the main channel, near the seasonal pycnocline above the oxicleine layers in most years. Third, these cores have excellent stratigraphic, sedimentological (Kerhin *et al.*, 1998; Baucom *et al.*, 2000) and chronological data (Cronin *et al.*, 1999; 2000; Colman *et al.*, 2002; Table 1), and pollen signatures from all cores provide a robust signal of regional vegetational variability.

Pollen and dinoflagellate cysts were isolated from sediments using standard palynological preparation techniques (Traverse, 1988; Willard and Korejwo, 2000). Pollen and dinocysts were isolated from 2 cm intervals in the core. Initial sample intervals were selected at 10 cm spacing, with samples added subsequently to improve temporal resolution. For each sample, one tablet of *Lycopodium* spores was added to 5–7 grams of dried sediment for calculation of pollen concentration (pollen/gram dry sediment). Samples were processed with HCl and HF to remove carbonates and silicates, acetolyzed (1 part sulphuric acid: 9 parts acetic anhydride) in a boiling water bath for 10 minutes, neutralized, and treated with 10% KOH for 10 minutes in a water bath at 70°C. After neutralization, residues were sieved with 149 µm and 10 µm nylon mesh to remove the coarse and clay fractions, respectively. When necessary, samples were swirled in a watch glass to remove mineral matter. After staining with Bismarck Brown, palynomorph residues were mounted on microscope slides in glycerin jelly. At least 300 pollen grains and 300 dinocysts were counted from each sample to determine percent abundance and concentration of palynomorphs. Confidence limits for *Pinus* and *Quercus* percentages were calculated using binomial standard errors as outlined in Buzas (1990). Mann-Whitney tests were used to determine whether abundance of indicator taxa varied significantly both within and among cores. Pollen data from surface samples and sediment cores are available from the North American Pollen Database (NAPD) at the World Data Center for Paleoclimatology in Boulder, CO (<http://www.ngdc.noaa.gov/paleo/pollen.html>), and both pollen and dinoflagellate cyst data are available at the USGS Chesapeake Bay website (<http://geology.er.usgs.gov/eespteam/ches/>).

Dinocyst assemblages were quantified from MD99-2209 and PTMC-3-2 to estimate changes in Chesapeake Bay environmental parameters from ecological data on modern dinocyst species and the modern analogue technique (MAT) (Overpeck *et al.*, 1985). Our modern dinocyst data set consists of percentage data for 16 species common to fossil and modern samples from 107 coretop sites along the eastern US coast. To identify common groupings of modern dinocyst assemblages, we performed a Q-mode cluster analysis with UPGMA and the Pearson correlation coefficient. We compared fossil and modern assemblages using the squared chord distance dissimilarity index (SCD). The 10 coretop samples with the lowest SCD values (i.e., with the greatest similarity between modern and fossil assemblages) were used to calculate palaeo-DO values for each fossil sample. Calculations using the top 5 and the top 15 modern analogue samples generally yielded similar results to those based on 10 samples. SCD values typically were

<0.10, indicating that they are close analogues (Overpeck *et al.*, 1985).

Stratigraphy and age

Holocene sediments in Chesapeake Bay comprise the Cape Charles channel fill, which unconformably overlies fluvial sands and gravels deposited during low sea level of the last glacial period (Colman and Mixon, 1988). The early-Holocene infilling of the bay during the final stages of deglaciation (10–7 ka) has been documented in analyses of the *Marion-Dufresne* cores collected on the IMAGES V cruise (Vogt *et al.*, 2000). The current study focuses on the past 2300 years. Postcolonial sediments were dated using chronologies based on ²¹⁰Pb (lead-210), ¹³⁷Cs (cesium-137) and total lead, as well as pollen biostratigraphy. Precolonial sediments were dated using accelerator mass spectrometry (AMS) radiocarbon dates from shells (Table 1); recent work indicates that total organic carbon (TOC) dates are consistently 1500–2000 years too old, so they were not included in calculation of age models (Cronin *et al.*, 2000; Colman *et al.*, 2002). Uncorrected radiocarbon ages from molluscs and foraminifers were calibrated using the CALIB 3.0 program of Stuiver and Reimer (1993), and all radiocarbon ages discussed herein are calibrated.

The most important pollen type used to date postcolonial events in Chesapeake sediments is *Ambrosia* (ragweed), an early successional plant that rapidly (≤ 1 yr) occupies cleared sites (Bazzaz, 1974; Keever, 1983). As summarized in Brush (1984) and illustrated in Figure 3, *Ambrosia* pollen abundance increased greatly after early colonial land clearance began in the seventeenth century. In these cores, we used ¹³⁷Cs, ²¹⁰Pb and, in MD99-2209, initial increases in total lead above background levels to develop age models for sediments deposited in the last ~120 years and to calibrate the three peaks consistently present in the *Ambrosia* curve (A1, A2, A3; Figure 4). In core PTMC-3, ²¹⁰Pb dating places the maximum *Ambrosia* abundance (peak 1) before AD1900 (± 10 yrs) (Figure 5a), and analyses of ²¹⁰Pb and total lead in a replicate core for MD99-2209 (core RD 98; Zimmerman, 2000; Zimmerman and Canuel, 2003) date the upper limit of peak 1 at AD 1910 (± 10 yrs) (Figure 5b). Based on these dates and the great increase in acreage cleared for lumber production beginning around 1880 (Figure 4), we estimate an age range of AD 1880–1910 (± 10 yrs) for *Ambrosia* peak 1. The magnitude of *Ambrosia* peak 2 is unclear from these samples, but it is dated at AD 1940–50 (Figure 4), based on age models constructed from ²¹⁰Pb and ¹³⁷Cs. *Ambrosia* peak 3 was dated at 1963–75 (± 5 yrs); the peak is preceded by the ¹³⁷Cs peak (AD 1963–64), and the total lead peak (AD 1975) corresponds to the *Ambrosia* maximum. During this interval, urban acreage expanded (Figure 4), and agricultural practices and land use were changed.

Dating the initial increase in *Ambrosia* to >2% is problematic because the time of most rapidly changing land-use practices (eighteenth century) is at the lower and upper limits of confident ²¹⁰Pb and radiocarbon dating, respectively. Therefore, dating must rely on a combination of historical accounts of land clearance and radiometric dates. Radiocarbon dates near the initial *Ambrosia* increase in these cores provide an estimate of AD 1700 as the earliest date for the rise (Figures 3, 8 and 9). Based on these data and the occurrence of eighteenth-century land-use changes summarized below, we estimate that the initial rise in *Ambrosia* to >2% occurred between AD 1700 and 1750. Land-use changes in the early to mid-eighteenth century include: settlement and development of new towns (Gottschalk, 1945); a shift from tobacco to grain production in the coastal region; and improvements in farm machinery and increases in tilling depths (see summary in Brush and Hillgartner, 2000).

Table 1 Locality, water depth, and age information for cores MD99-2209, PRCK 1-2, PTXT 2-P-5 and PTMC 3-P-2 (see Colman *et al.*, 2002; Cronin *et al.*, 2000)

Site and core no.	Latitude (°N)	Longitude (°W)	Water depth (m)	Lab. no.	Depth of C14 date (cm)	Material dated	$\delta^{13}\text{C}$	^{14}C age (conventional)	1σ	Calibrated age (yr BP)	$\pm 2\sigma^*$ (yr BP)
Rhode River											
MD99-2209	38° 53.18'	76° 23.68'	26	OS-21226	296	bivalve	-0.87	610	30	270	300-150
RD 98-1 ^a	38° 53.202'	76° 23.502'	26.5	OS-19215	340	bivalve	-0.87	725	55	340	270-460
MD99-2209	38° 53.18'	76° 23.68'	26	OS-21381	369	bivalve	-0.57	745	35	410	450-510
MD99-2209	38° 53.18'	76° 23.68'	26	OS-21382	455	bivalve	-0.68	1150	40	680	760-640
RD 98-1 ^a	38° 53.202'	76° 23.502'	26.5	OS-19214	457	bivalve	-1.03	1150	85	675	550-870
MD99-2209	38° 53.18'	76° 23.68'	26	OS-21227	485	bivalve	-1.29	1240	30	770	870-710
MD99-2209	38° 53.18'	76° 23.68'	26	OS-21383	576	bivalve	-0.9	1600	35	1165	1230-1070
MD99-2209	38° 53.18'	76° 23.68'	26	OS-21384	665	bivalve	-1.73	2050	40	1610	1700-1520
MD99-2209	38° 53.18'	76° 23.68'	26	OS-21228	733	bivalve	-0.77	2210	35	1810	1880-1720
MD99-2209	38° 53.18'	76° 23.68'	26	OS-21229	780	bivalve	-0.74	2500	35	2140	2280-2070
Parker Creek											
PRCK-1-2	38° 32.8657'	76° 28.7112'	10.7	WW-1584	171	molluscs	NA	1550	50	1082	968-1207
PRCK-1-2	38° 32.8657'	76° 28.7112'	10.7	157685	295-300	molluscs	-2.1	2120	40	1710	1610-1810
PRCK-1-2	38° 32.8657'	76° 28.7112'	10.7	WW-1583	317	molluscs	NA	6770	70	7237	7143-7375
Patuxent River											
PTXT-2-P-5	38° 20.0007'	76° 18.5801'	12	WW-1816	298-304	<i>Mulinia lateralis</i>	NA	750	60	403	271-490
PTXT-2-P-5	38° 20.0007'	76° 18.5801'	12	WW-1817	388-392	<i>Mulinia lateralis</i>	NA	1030	50	608	517-665
Potomac River											
PTMC-3-P-2	38° 01.6118'	76° 13.1938'	23.1	OS 15679	81	<i>Mulinia lateralis</i>	0.01	540	30	150	0-260
PTMC-3-P-2	38° 01.6118'	76° 13.1938'	23.1	WW 1284	141	<i>Mulinia lateralis</i>	0.1	540	50	150	0-280
PTMC-3-P-2	38° 01.6118'	76° 13.1938'	23.1	OS 15680	161	<i>Mulinia lateralis</i>	-0.29	885	35	500	450-540
PTMC-3-P-2	38° 01.6118'	76° 13.1938'	23.1	OS 15681	211	<i>Mulinia lateralis</i>	0.01	1150	25	675	650-720
PTMC-3-P-2	38° 01.6118'	76° 13.1938'	23.1	WW 1589	225	<i>Mulinia lateralis</i>	0	990	40	550	510-640
PTMC-3-P-2	38° 01.6118'	76° 13.1938'	23.1	IS 17242	229	<i>Elphidium</i>	-1.72	1230	30	750	680-830
PTMC-3-P-2	38° 01.6118'	76° 13.1938'	23.1	OS 15689	297	<i>Mulinia lateralis</i>	0.1	1530	70	1060	920-1230
PTMC-3-P-2	38° 01.6118'	76° 13.1938'	23.1	OS 17508	331	<i>Elphidium</i>	-2.41	2450	256	2080	1530-2710
PTMC-3-P-2	38° 01.6118'	76° 13.1938'	23.1	OS 17241	392	<i>Elphidium</i>	-1.94	2400	85	2000	1820-2270

^o One sigma counting errors.

*Upper and lower limits of range based on two-sigma errors in calibration.

^a From Zimmerman and Canuel (2000).

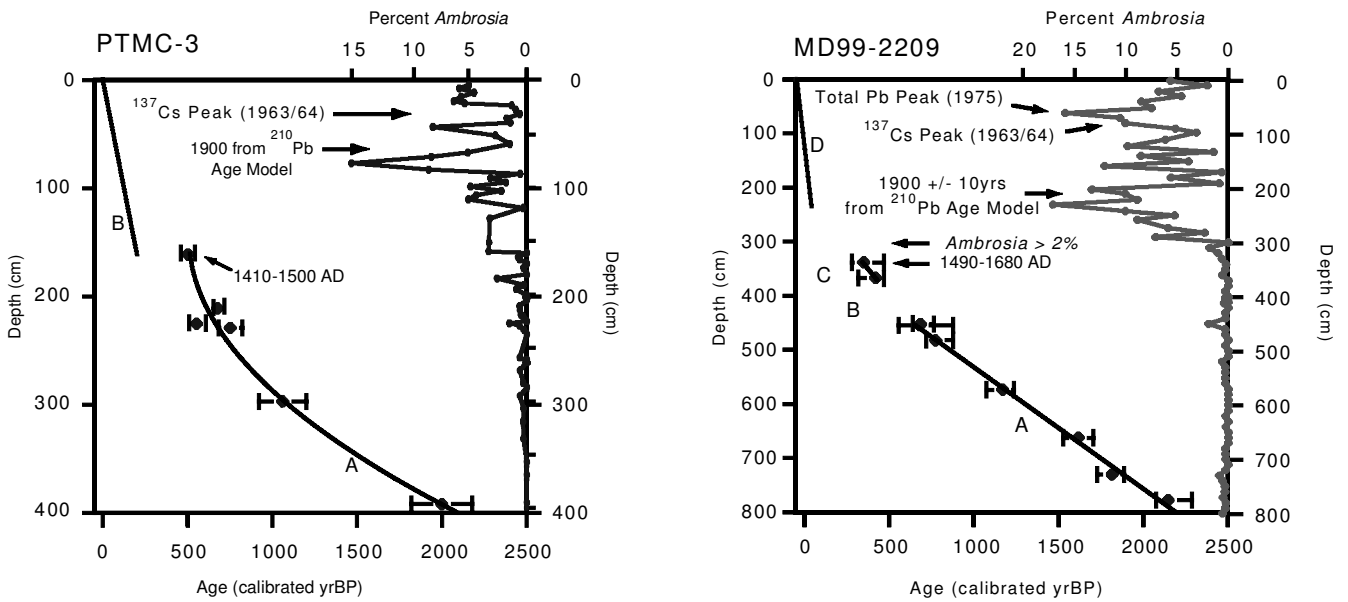


Figure 3 Age (calibrated yr BP) versus depth plots for radiocarbon dates and ^{210}Pb age models in cores PTMC-3 and MD99-2209. All coefficients of determination (r^2) are given at 95% significance levels. For core PTMC-3, two segments were constructed, one based on radiocarbon dates and another on ^{210}Pb data. For segment A (160–450 cm), age = $(0.012 \cdot \text{depth}^2) + (0.97 \cdot \text{depth}) - 41.675$ ($r^2 = 0.86$); for segment B (0–160 cm), age = $\text{depth} \cdot 1.25$. The early colonial *Ambrosia* increase occurred at 155 cm, overlying a depositional hiatus. For core MD99-2209, four separate age models were generated for different segments of the core. For segment A (370–800 cm), age = $(4.101 \cdot \text{depth}) - 1097.918$ ($r^2 = 0.99$); for segment B (330–370 cm), age = $(2.75 \cdot \text{depth}) - 555.750$ ($r^2 = 1.00$); for segment C (230–330 cm), age = $(2.549 \cdot \text{depth}) - 489.893$ ($r^2 = 0.98$); segment D (0–230 cm), age = $(0.492 \cdot \text{depth}) - 1.5$ ($r^2 = 0.99$) (from Zimmerman, 2000; Zimmerman and Canuel, 2003). *Ambrosia* curves also are plotted versus depth, with critical dates from short-lived radioisotopes (^{210}Pb , ^{137}Cs and total lead) indicated.

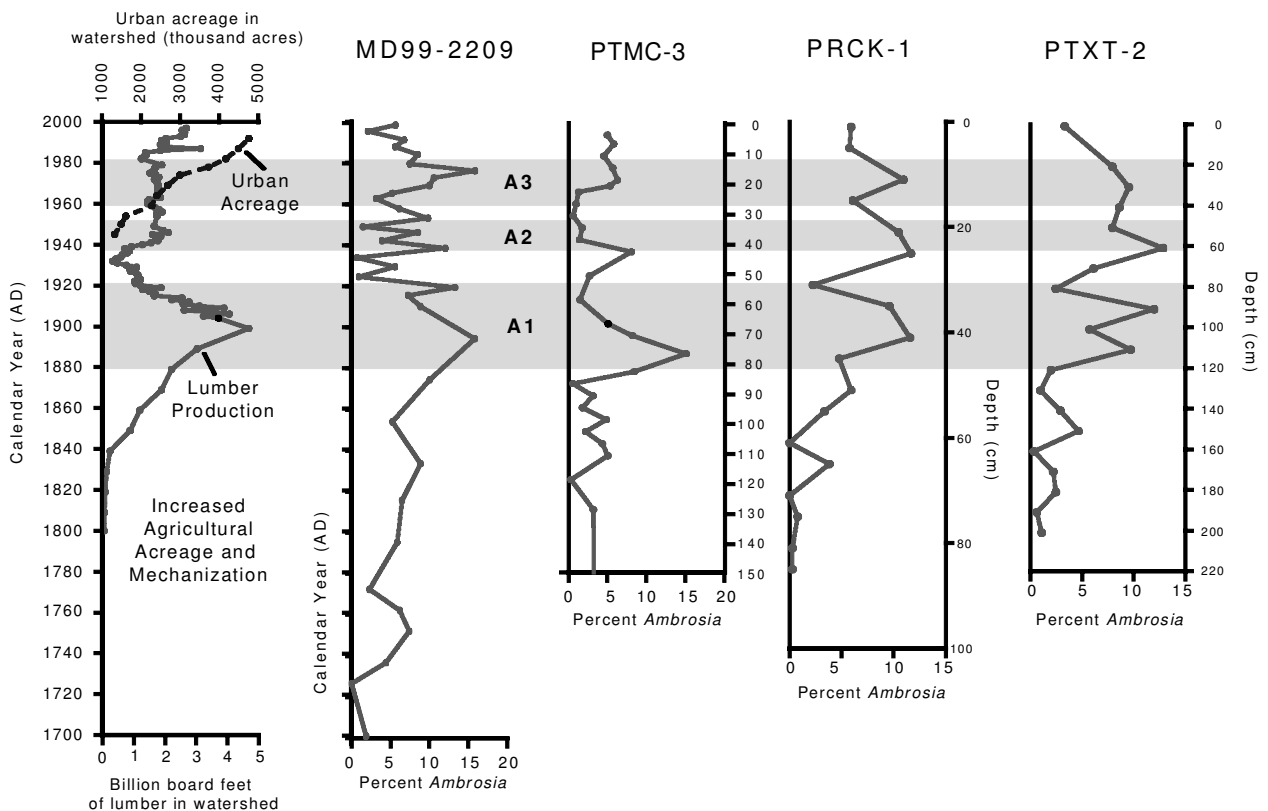


Figure 4 Timber production in Chesapeake Bay watershed (compiled from Steer, 1948; US Department of Commerce, 1946–98), urban acreage in Chesapeake Bay watershed (Krupa and Daugherty, 1996) and *Ambrosia* abundance (with peaks A1–A3) in cores MD99-2209, PTMC-3, PRCK-1 and PTXT-2. Note that the PTXT-2 core is a replicate core for the one shown in Figure 9; in this core, the upper part was analysed to identify the agricultural horizon.

Results

Calibration of palaeoecological proxies – pollen

Pollen preserved in sediments provides a distinctive signature of the source vegetation, and temporal changes in pollen abundance

have traditionally been interpreted as signifying changes in vegetational composition on centennial to millennial timescales. However, pollen production also influences pollen abundance over annual to decadal periods. We assessed both the ability of pollen assemblages to reflect forest composition using surface samples

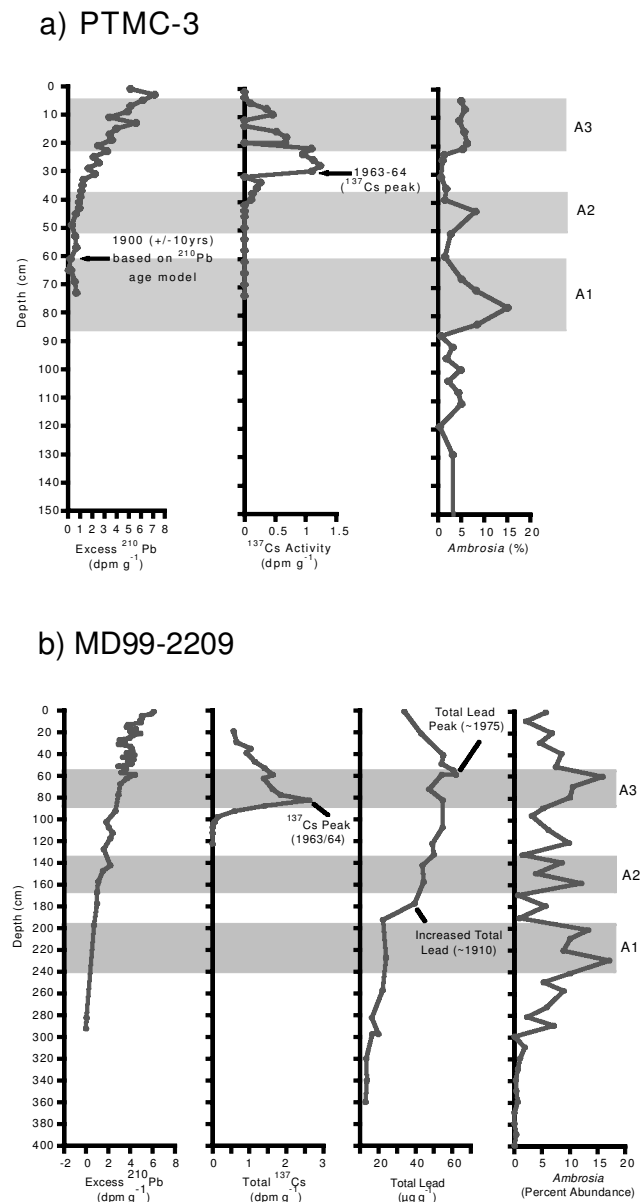


Figure 5 Concentration of short-lived radioisotopes (^{137}Cs , ^{210}Pb , and total Pb) and *Ambrosia* abundance versus depth in cores (a) PTMC-3 and (b) MD99-2209. Radioisotope values for core MD99-2209 were generated from a replicate core (RD98) by A. Zimmerman (Zimmerman, 2000; Zimmerman and Canuel, 2003). Pollen and radioisotope measurements from PTMC-3 were made on the same core.

and the interannual variability of pollen production in the region using the dominant taxa in the watershed, *Quercus* (oak) and *Pinus* (pine).

Forest composition in the bay area reflects a hydrologic gradient from oak-chestnut forests on well-drained soils in the north to oak-pine-hickory forests on more poorly drained soils in the south (Brush *et al.*, 1980). Oak-chestnut forests (Braun, 1950; Greller, 1988), also described as a tulip poplar association by Brush *et al.* (1980), dominate the northern coastal plain (Figure 1). These are primarily mixed deciduous forests, with common *Quercus alba* (white oak), *Carya* (hickory/pecan), *Liriodendron* (tulip poplar), *Acer* (maple) and *Nyssa* (gum). Prior to its demise from chestnut blight in the early twentieth century, *Castanea* (chestnut) was dominant or codominant with oaks in these forests (Greller, 1988). Oak/pine/hickory forests in the south are dominated by *Pinus taeda* and *P. echinata* (loblolly and short-leaf pines, respectively) and *Quercus virginiana* (live oak), grading southward into loblolly pine forests (Braun, 1950). The boundary

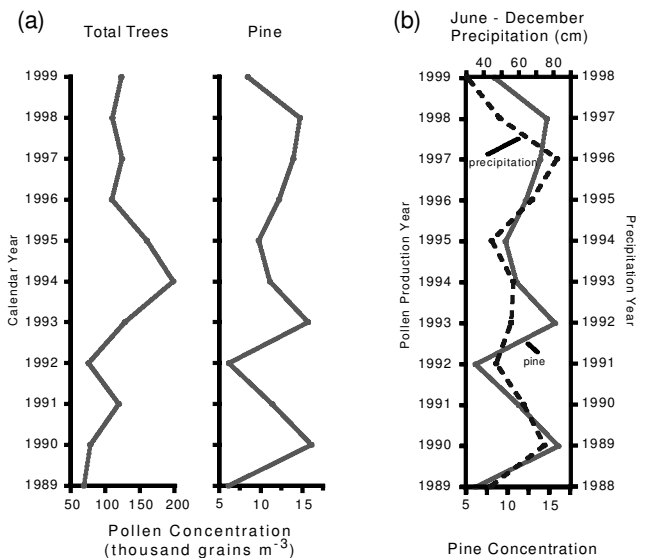


Figure 6 (a) Concentration of tree and *Pinus* pollen (grains l^{-3}), measured by a volumetric pollen sampler, Walter Reed Army Medical Center, Washington, DC, between 1989 and 1999 (data provided by S. Kosisky). Although *Pinus* pollen is a small proportion of total tree pollen counted during the year, it and total tree pollen fluctuated by a factor of two to four during the period of record. (b) *Pinus* pollen concentration (grains l^{-3}) from 1989 to 1999 (solid line) and June–December precipitation (cm) the previous year (1988–1998) (dashed line). Precipitation data obtained from: <http://www.erh.noaa.gov/er/lwx/climate/dca/dcaprecip.txt>.

between the two forest types corresponds generally to the northernmost natural occurrences of southern pines at about 38°N latitude. (Bartlein *et al.*, 1986; Hocker, 1956).

We compared pollen distribution in surface sediments (0–2 cm) from the mainstem of Chesapeake Bay with forest composition of the surrounding watershed. *Pinus* pollen abundance is correlated well with latitude ($r^2 = 0.595$, at 0.05 significance level), and decreases to the north. *Quercus* pollen abundance increased northward, but the correlation is poorer ($r^2 = 0.398$, at 0.05 significance level), due to its common presence in both forest types. Isopolls from limited terrestrial sites in North Carolina, Maryland and New Jersey show a similar latitudinal pattern in *Pinus-Quercus* abundance (Bernabo and Webb, 1977; Davis and Webb, 1975; Delcourt and Delcourt, 1984; Delcourt *et al.*, 1984), as do surface sediments from the Potomac River (Brush and DeFries, 1981; DeFries, 1986). These comparisons indicate that changes of about 10% in *Pinus* abundance represent differences in its abundance in forests. Subdominant wind-pollinated species, such as *Carya* and *Liquidambar*, vary little in abundance throughout bay sediments, and several common taxa (i.e., *Liriodendron*, *Acer*, *Nyssa* and *Ilex* (holly)) are poorly represented, either because they are insect-pollinated or produce small quantities of pollen (Brush and DeFries, 1981; Traverse, 1988).

We evaluated short-term variability in pollen production through analysis of atmospheric pollen count data generated by allergists at Walter Reed Army Medical Center in Washington, DC. Concentration of tree and herb pollen (grains m^{-3}) were obtained three days a week from 1989 through 1999 from an volumetric rotating-arm impaction sampler placed at an elevation of 37.5 m on an unobstructed roof (Kosisky and Carpenter, 1997; S. Kosisky, personal communication). Because the sampler is situated above tree canopies, it captures the regional pollen rain, analogous to pollen rain deposited in the estuary. Indeed, highest pollen concentrations are recorded in the spring (Kosisky and Carpenter, 1997), corresponding to the time of maximum pollen concentration in the water (Brush and Brush, 1994). Interannual variation in total atmospheric pollen concentration varied two- to fourfold during the 11 years of record, with particular variability

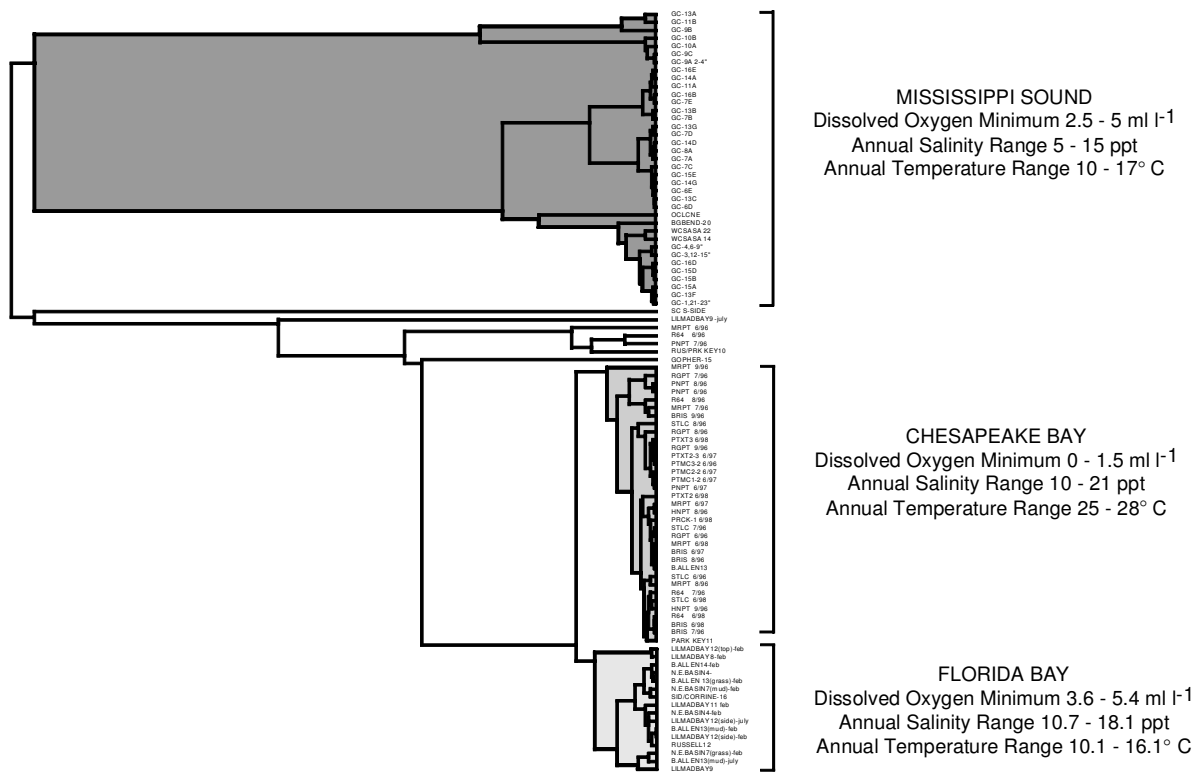


Figure 7 Dendrogram from Q-mode cluster analysis (using UPGMA and Pearson correlation coefficient) of dinoflagellate cyst data from surface samples in Chesapeake Bay, Florida Bay and Mississippi Sound. Dinocyst data and environmental parameters for each site are available from the USGS Chesapeake Bay website (<http://geology.er.usgs.gov/kespteam/ches/bayhome.html>). Salinity data for the Florida Bay samples were obtained from the USGS website on Florida Bay (www.flaccolist.usgs.gov); mean and minimum dissolved oxygen for these sites were measured in 7/98, 2/99 and 7/99 (Jeffery Stone, personal communication). Data on salinity and D/O for Chesapeake Bay were obtained through the Chesapeake Information Management System (CIMS) (www.chesapeakebay.net/cimsindex.htm). The maximum and minimum D/O, S and T values for surface and bottom water were calculated by taking the mean maximum and minimum values for a five-year period between 1 January 1990 and 31 December 1995. Environmental data for the Mississippi Sound coretop samples were obtained from the hydrologic and climatic atlas compiled by Eleuterius and Beaugez (1979). Seasonal D/O (summer mean and summer minimum), salinity (fall and spring means) and temperature (summer and winter bottom means) data for each dinoflagellate coretop site were taken from isopleths in the Eleuterius and Beaugez atlas.

in *Pinus* pollen concentration (Figure 6a). When translated to relative (percent) abundance, the interannual variability of *Pinus* pollen production is between 2% and 11%, similar to subdecadal- to decadal-scale variability.

We compared concentrations of *Pinus* and *Quercus* with temperature and precipitation records for the previous June through December. We selected this time interval because production of catkins (pollen cones) requires adequate soil moisture from the time of catkin initiation (~June) until pollen ripening (~December) the previous year (Boyer, 1981). Although no correlation existed between *Quercus* abundance and precipitation ($r^2 = 0.01$, at a 0.05 significance level), *Pinus* pollen production was positively correlated with June–December precipitation ($r^2 = 0.411$, at a 0.05 significance level) (Figure 6b). This indicates that variation in precipitation during the last half of the previous year explains more than one-third of *Pinus* pollen production in the Washington, DC, area; maximum production occurs after relatively wet summers and falls, as was suggested by Boyer (1981). Therefore, on short (subdecadal to decadal) timescales, variation in *Pinus* pollen abundance of the order of 10% or more in Chesapeake Bay sediment cores likely reflects differential pollen production tied to variation in regional precipitation; on longer time scales (multidecadal to centennial), it represents changes in forest composition reflecting long-term changes in precipitation and, possibly temperature.

Calibration of palaeoecological proxies – dinoflagellate cysts

Phytoplankton taxa occupy a critical position in the Chesapeake Bay ecosystem because of their role as primary producers for

higher trophic levels and in oxygen dynamics of the bay. Along with diatoms, many photosynthetic (autotrophic) dinoflagellates (organic-walled, unicellular protists) are primary algal producers in Chesapeake Bay (Marshall and Alden, 1988), and dinoflagellate species that produce cysts often are fossilized in Chesapeake Bay sediments. Because of their sensitivity to water quality (turbidity, dissolved oxygen), salinity and temperature (Wall *et al.*, 1977; Turon, 1984; Edwards and Anderle, 1992), dinoflagellate species are increasingly used in palaeoceanography (Harland, 1983), and palaeoecological studies of ocean margin ecosystems (de Vernal *et al.*, 1989; Ellegaard, 2000).

We compiled ecological preferences of estuarine dinocysts from three estuarine regions in the eastern USA, Chesapeake Bay, Florida Bay (Brewster-Wingard *et al.*, 1996) and Mississippi Sound (Edwards and Willard, 2001). Stratification of the deep channel of Chesapeake Bay results in seasonal oxygen depletion (minimum DO from 0 to 1.75 ml l⁻¹) that may last through the summer months depending on climatically driven fluctuations in freshwater inflow and wind-driven mixing (Seliger and Boggs, 1988; Malone, 1991; Cronin *et al.*, 1999). Annual temperature range in this temperate estuary is large, ranging from 1–2°C (winter) to 29–30°C (summer), as is salinity range (3–24 ppt). Florida Bay is a well-oxygenated, shallow-water, subtropical bay with DO levels from 3 to 7 ml l⁻¹, an annual temperature range of 10–12°C (winter) to 32–33°C (summer) and salinity range of 10–41 ppt. Mississippi Sound varies between a partially and well-mixed estuary and occasionally becomes highly stratified. In general, the waters are well oxygenated (minimum DO 2.5 to 5 ml l⁻¹) with a relatively small salinity range (5–15 ppt) and annual

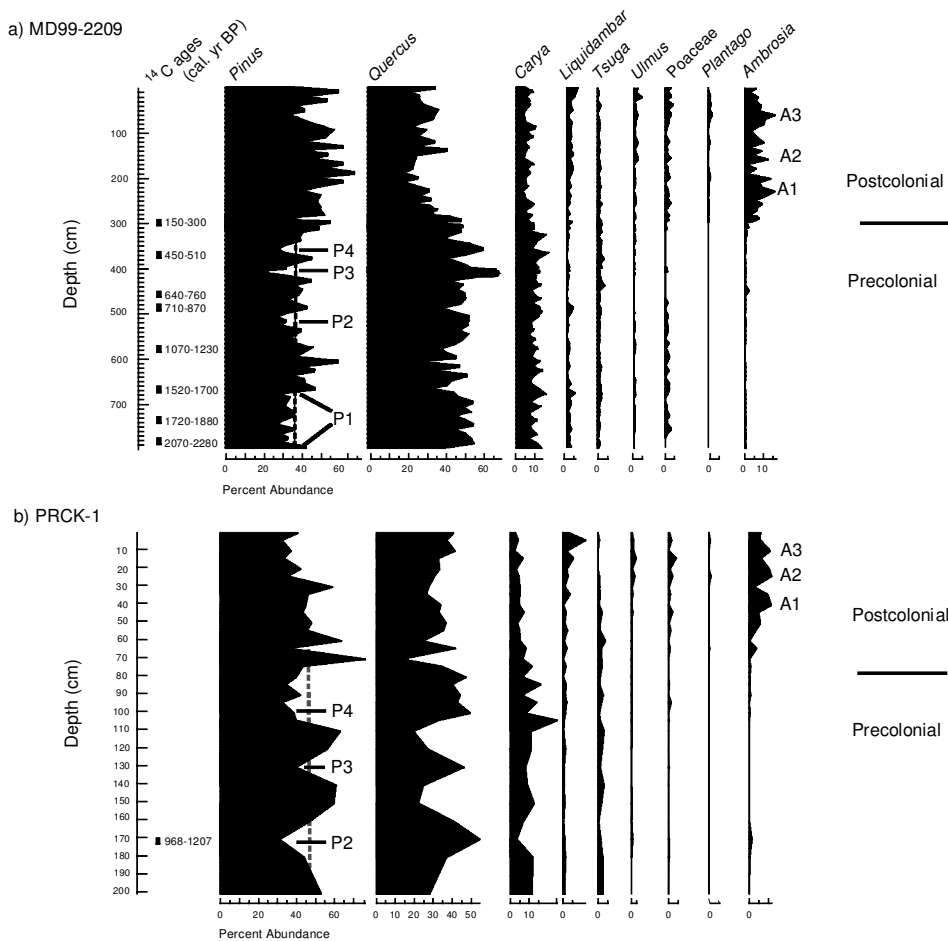


Figure 8 Percent abundance of pollen of major plant groups, Chesapeake Bay. Dashed line indicates average precolonial abundance. (a) Core MD99-2209, near mouth of Rhode River. (b) Core PRCK-1, near mouth of Parker Creek.

temperature ranging from 10–12°C (winter) to 27–28°C (summer).

Using Q-mode cluster analysis (with UPGMA and Pearson correlation coefficient) we identified groupings of modern samples from the three regions based on dinocyst census data. We also estimated environmental preferences of dinocyst species from conditions at 107 surface sample sites covering a range of salinity and dissolved oxygen regimes: maximum and minimum salinity (S) range from 15 to 40 ppt and 3 to 30 ppt, respectively, and median and minimum dissolved oxygen (DO) from 3.5 to 7.5 and 0 to 5.5 ml l⁻¹, respectively. The modern samples separated into three clear groupings, generally corresponding to the three estuaries (Figure 7).

The two primary factors affecting assemblages appear to be annual salinity range and annual minimum DO. The first-order separation of Chesapeake Bay and Florida Bay samples from Mississippi Sound reflects the dominance of *Spiniferites* spp., which here includes primarily *S. bulloideus* and *S. ramosus* (but excludes *S. mirabilis*); *Operculodinium centrocarpum* is common (up to 40%) in bay assemblages. The abundance of these taxa reflects the broad salinity tolerance characteristic of such cosmopolitan species (Wall *et al.*, 1977). Chesapeake Bay and Florida Bay assemblages are separated primarily on the basis of the strong dominance of *Spiniferites* spp. and *S. mirabilis* in Chesapeake Bay compared to abundance of species more typical of neritic to oceanic waters (*Nematosphaeropsis*, *Lingulodinium machaerophorum* (Wall *et al.*, 1977)) in Florida Bay. *Spiniferites mirabilis*, a species tolerant of reduced levels of DO (Harland, 1983; Turon and Londeix, 1988) is most abundant (up to 15%) in Chesapeake Bay sediments in areas that undergo a wide annual range of DO and salinity. Mississippi Sound assemblages are dominated

strongly by *Polysphaeridium zoharyi*, a species restricted to tropical/subtropical estuaries that tolerates only small salinity fluctuations (Wall *et al.*, 1977). In summary, *S. mirabilis* is a good indicator of seasonal hypoxia and broad salinity fluctuations, whereas *Operculodinium centrocarpum*, *Nematosphaeropsis* and *Lingulodinium* are characteristic of higher salinities and DO levels.

Late-Holocene pollen and dinocyst assemblages

Late-Holocene pollen and dinocyst spectra are divided clearly into pre- and postcolonial units by increases in abundance of *Ambrosia* and *Spiniferites mirabilis* (Figures 8–10). Postcolonial pollen assemblages are distinguished by increased abundances of *Ambrosia*, *Poaceae* and *Pinus* pollen, consistent presence of *Plantago*, and decreased abundance of *Quercus* and *Carya* pollen (Figures 8 and 9). Non-dominant trees, such as *Liquidambar*, *Ulmus* and *Nyssa*, each doubled their pollen abundance during the transition from pre- to postcolonial time. Postcolonial dinocyst assemblages are characterized by a higher abundances of *Spiniferites mirabilis* and *Spiniferites* spp. (primarily *S. bulloideus* and *S. ramosus*). Although present, typical marine species, including *P. zoharyi*, *L. macherophorum*, (<5%), and *N. labyrinth* were less abundant than in precolonial sediments (Figure 10).

Precolonial pollen assemblages reveal both spatial and temporal patterns in forest distribution. *Pinus* pollen was significantly ($p < 0.0005$) more abundant (averaging 40–47%) in southern sites (PTMC-3, PTXT-2, PRCK-1) than at MD99-2209 in the north, where it averaged 35%. The existence of a long-term (>2000 year

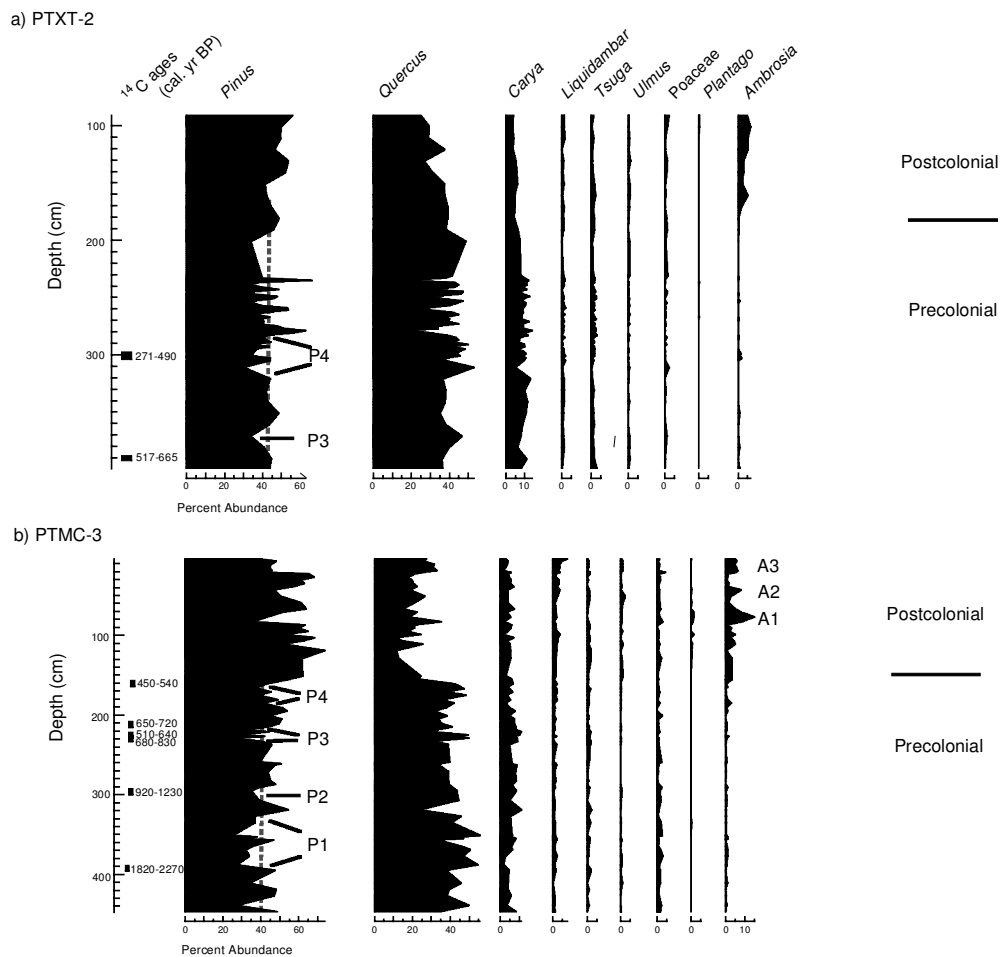


Figure 9 Percent abundance of pollen of major plant groups, Chesapeake Bay. Dashed line indicates average precolonial abundance. (a) Core PTXT-2, near mouth of Patuxent River. Note that this is a replicate core for the one shown in Figure 4 and was analysed for details of precolonial variability. (b) Core PTMC-3, near mouth of Potomac River.

transition zone between oak-chestnut and oak-pine-hickory forests indicates the inherent control of geomorphic, climatic and edaphic factors on their distribution. Throughout the late Holocene, pollen assemblages were dominated by *Pinus* and *Quercus*, with consistent, but relatively invariant, occurrence of *Carya*, *Liquidambar* and other non-dominant trees. Four sustained intervals of decreased *Pinus* abundance (P1–P4) are documented in these cores. In these intervals, *Pinus* abundance was 5–15% below the precolonial average and 20–30% less than adjacent peak abundances (Figures 8 and 9). Confidence limits calculated for *Pinus* were 4–5% (at a 0.05 significance level), so changes of 20–30% are significant. Interval P1, preserved in cores MD99-2209 and PTMC-3-2, lasted approximately 500 years (2100 to 1600 yr BP). Interval P2, preserved in MD99-2209, PRCK-1 and PTMC-3, appears to have lasted about 400 years (1150 to 750 yr BP). Intervals P3 and P4 each lasted less than one century, from 490 to 570 yr BP and 320 to 365 yr BP, respectively.

Precolonial dinocyst assemblages in both MD99-2209 and PTMC-3 were dominated (58% and 44%, respectively) by *Spiniferites* spp., primarily including *S. bulloideus* and *S. ramosus*. *Operculodinium centrocarpum* was subdominant, and *Polysphaeridium zoharyi*, *Lingulodinium machaerophorum*, *Nematospheropsis* spp. and *Tuberculodinium* spp. were common components. *Spiniferites mirabilis* was present consistently, with two intervals of sustained lower-than-average abundance (Figure 10). The first (D1) lasted about 350 years (2100 to 1750 yr BP), and the second (D2) lasted about 450 years (1200 to 750 yr BP).

Discussion and conclusions

Our results have implications for understanding patterns and causes of late-Holocene climatic variability in eastern North America. Several dry periods ranging from decades to centuries in duration are evident in Chesapeake Bay records. Multidecadal events include intervals P4 and P3 (Figure 11), which are preserved in all four sediment cores. Interval P4 (~ AD 1525–1650) consists of several decadal-scale dry intervals (Figure 12), including the most severe mid-Atlantic droughts of the last 800 years, which affected the Roanoke Island (1587–89) and Jamestown (1606–12) colonies of Virginia (Stahle *et al.*, 1998). Interval P3 (~ AD 1320–1400) corresponds to fourteenth-century dry conditions in North Carolina (Stahle and Cleaveland, 1992). The diminished percentages of *Pinus* in these sedimentary records offer strong evidence that multiyear droughts had great effects on pollen production in sensitive tree taxa.

Periods of lower-than-average precipitation spanning several centuries (intervals P2/D2 and P1/D1) apparently altered forest composition in the Chesapeake Bay region. Interval P2/D2 (~ AD 800–1200) corresponds to the ‘Mediaeval Warm Period’, which has been documented as drier than average by tree-ring (Stahle and Cleaveland, 1994) and pollen (Willard *et al.*, 2001) records from the southeastern USA. Interval P1/D1 (~200 BC–AD 300) represents the first documentation of mid-Atlantic or southeastern USA dry conditions; however, based on reconstructions of past air mass regimes, Bryson and Wendland (1967) suggested

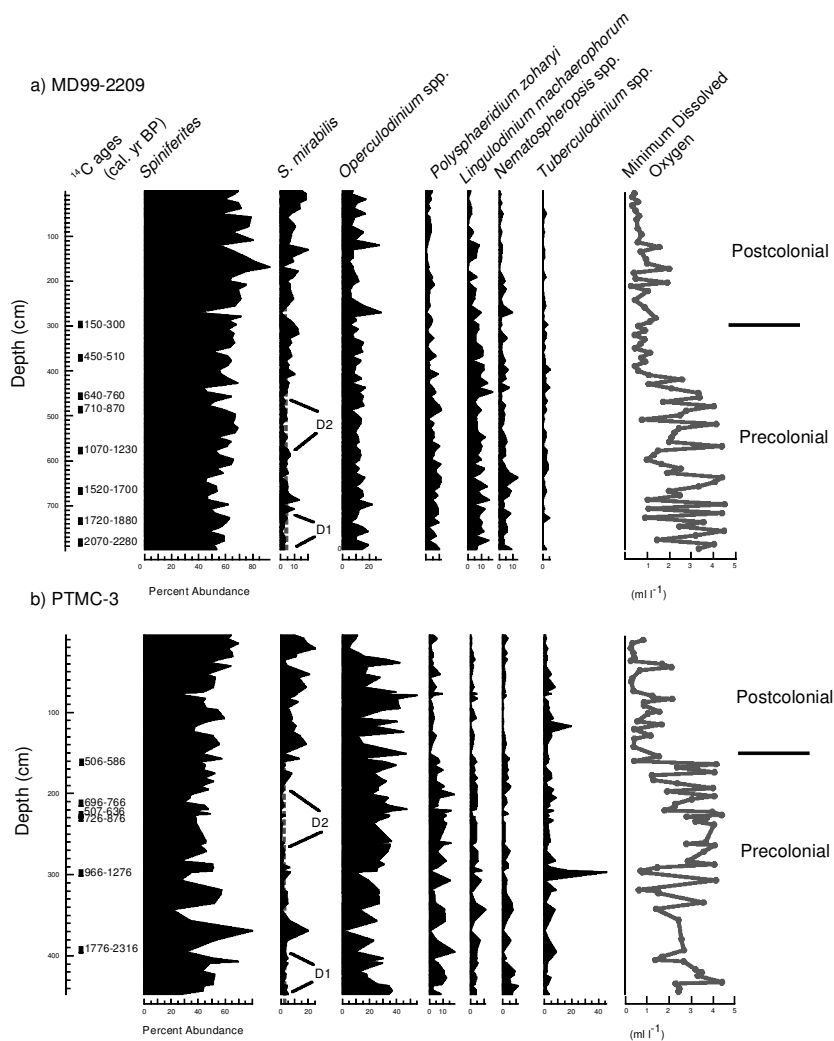


Figure 10 Percent abundance of major dinocyst taxa, Chesapeake Bay. Dashed line indicates average precolonial abundance. Estimates of minimum annual dissolved oxygen are calculated from minimum DO values for 10 closest modern analogues. (a) Core MD99-2209, near mouth of Rhode River. (b) Core PTMC-3, near mouth of Potomac River.

that drought conditions probably occurred in the mid-Atlantic region during this time.

Mid-Atlantic dry periods generally correspond to central and southwestern USA 'megadroughts', described by Woodhouse and Overpeck (1998) as major droughts of decadal or more duration that probably exceeded twentieth-century droughts in severity. Droughts in the late sixteenth century that lasted several decades, and those in the 'Mediaeval Warm Period' and between ~AD 50 and AD 350 spanning a century or more have been indicated by Great Plains tree-ring (Stahle *et al.*, 1985; Stahle and Cleaveland, 1994), lacustrine diatom and ostracode (Fritz *et al.*, 2000; Laird *et al.*, 1996a; 1996b) and detrital clastic records (Dean, 1997).

Modern atmospheric conditions during which meridional flow dominates the mid-Atlantic climate are conceptually similar to the dry intervals documented during the last 2300 years of Chesapeake Bay record. During such times, weakening of the polar jet across the North Atlantic allows greater input of dry polar air to the mid-Atlantic region and blocks influx of warm tropical air northward along the coast; enhancement of the ridge and trough system across the continent draws Arctic air to the mid-Atlantic (Yarnal and Leathers, 1988). Great Plains climate patterns are also influenced by the position of the North American jet stream (Mo *et al.*, 1997; Castro *et al.*, 2001). When a strong ridge is positioned over the Rocky Mountains (as occurs during times of enhanced meridional flow), Great Plains sites on the subsiding branch of the ridge will have dry climates. The wavelength and amplitude of the ridge and trough system determines whether precipitation

is also reduced in the Chesapeake Bay region, because the exact position of the jet stream determines precipitation totals (Yarnal and Leathers, 1988). The apparently coeval dry periods in the central and eastern USA documented here may have resulted from the coincidence of enhanced meridional flow both over North America and the western Atlantic Ocean. In modern records, such changes in upper-level atmospheric flow patterns have been attributed to changes in the position of the intertropical convergence zone (Castro *et al.*, 2001; Harman, 1991; Mo *et al.*, 1997), sea-level pressure anomalies over the Atlantic and Pacific Oceans (Leathers *et al.*, 1991; Wallace and Gutzler, 1981) or tropical sea-surface temperature anomalies (Montroy, 1997); the longer-term changes documented here may represent amplifications of these phenomena.

Chesapeake Bay sediments provide an exceptionally detailed record of postcolonial forest cover and bay ecosystem. Pollen evidence indicates that the naturally occurring north-south gradient from oak-chestnut to oak-pine-hickory forests was eliminated by colonial land clearance. About a decade after the first *Ambrosia* maximum (~AD 1890), *Pinus*, which typically is the first tree established in cleared fields in the region, became more abundant (Figure 13). This pattern is consistent with evidence for earliest seeding of southern pines at 10–20 years of age (Iverson *et al.*, 1999). Non-dominant trees, such as *Liquidambar*, *Ulmus* and *Juglans*, became more abundant in the 1940s. These taxa typically colonize under pines and other early successional trees before being outcompeted by other hardwoods (Iverson *et al.*, 1999).

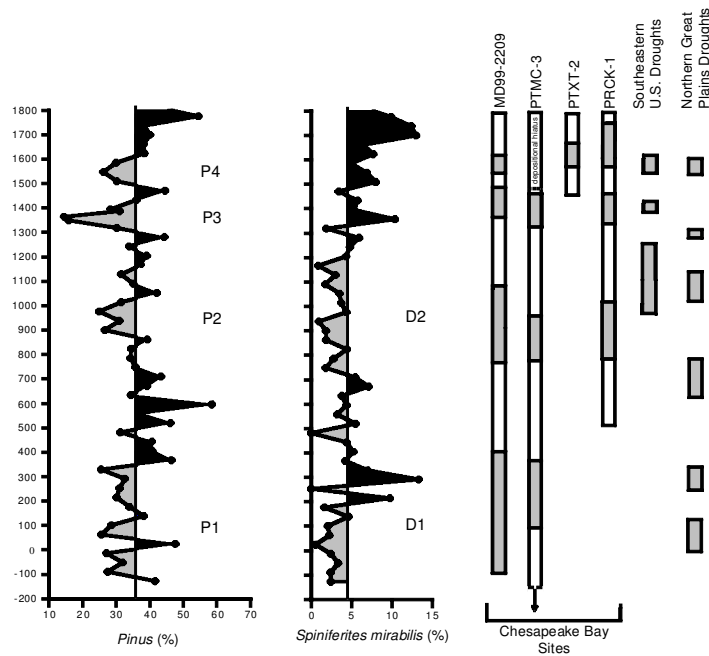


Figure 11 Percent abundance of *Pinus* pollen and *Spiniferites mirabilis* during the last 2200 years, core MD99-2209, Chesapeake Bay, near mouth of Rhode River. Intervals P1–P4 and D1–D2 indicate intervals discussed in text based on pollen and dinocyst assemblages. Major dry periods preserved in cores MD99-2209, PTMC-3, PTXT-2 and PRCK-1 are indicated, along with the duration of each at the site. Duration of droughts recorded elsewhere in the USA were obtained from the following – southeastern USA: Stahle *et al.*, 1988; 1998; 2000; Stahle and Cleaveland, 1992; northern Great Plains: Fritz *et al.*, 2000; Laird *et al.*, 1996a; Stahle *et al.*, 2000.

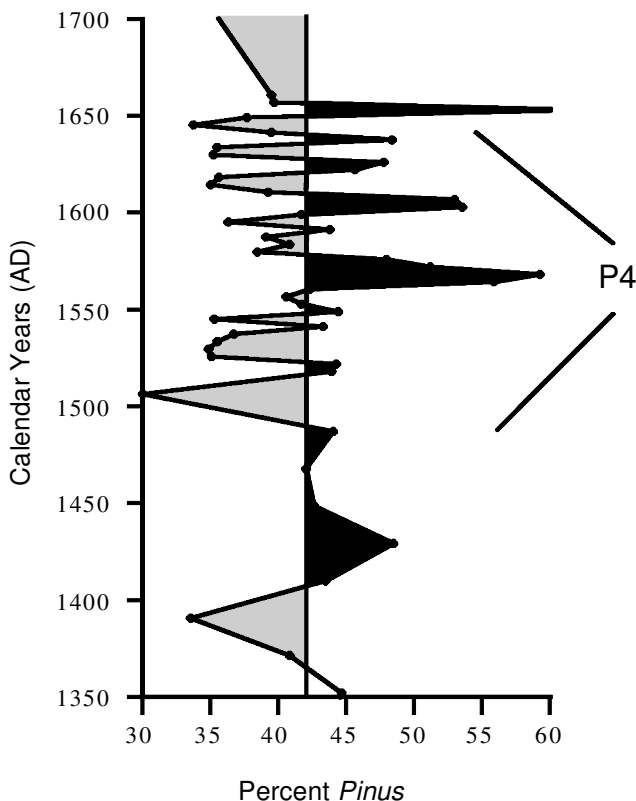


Figure 12 Percent abundance of *Pinus* pollen between AD 1350 and 1700, site PTXT-2, showing detail of decadal dry periods within interval P4.

Greater abundance of weedy species since 1960 corresponds to modern agricultural practices and urbanization, as has been documented at different times in Europe (Iversen, 1941; Odgaard and Rasmussen, 1998).

Two phases of anthropogenic influence on phytoplankton are evident, one beginning in the late nineteenth century and another after 1950. Following peak lumber harvesting between 1880 and

1910, sedimentation rates increased two- to fourfold (Brush, 1984; Colman *et al.*, 2002; Cronin *et al.*, 1999); several dinocyst taxa nearly disappeared (*Lingulodinium*, *Nematophaeropsis labyrinthica*, *Tuberculodinium* and *Achomosphaera*), and *Spiniferites* became dominant. Although climate-driven declines in both salinity and dissolved oxygen levels affected phytoplankton as early as the sixteenth to seventeenth centuries, long before peak land clearance (Figure 10), late-nineteenth-century land clearance exacerbated the decline. After 1950, dinocyst assemblage diversity decreased, reflecting water-quality changes associated with increased urbanization, greater hypoxia (Karlsen *et al.*, 2000) and increased agricultural nutrient input (Jaworski *et al.*, 1997). When viewed in the context of natural, late-Holocene terrestrial and estuarine ecosystem variability, it is clear that land-use changes of the last centuries had unprecedented impacts on regional forests and the bay. It remains unclear to what degree restoration efforts will return forests and water quality to their precolonial conditions.

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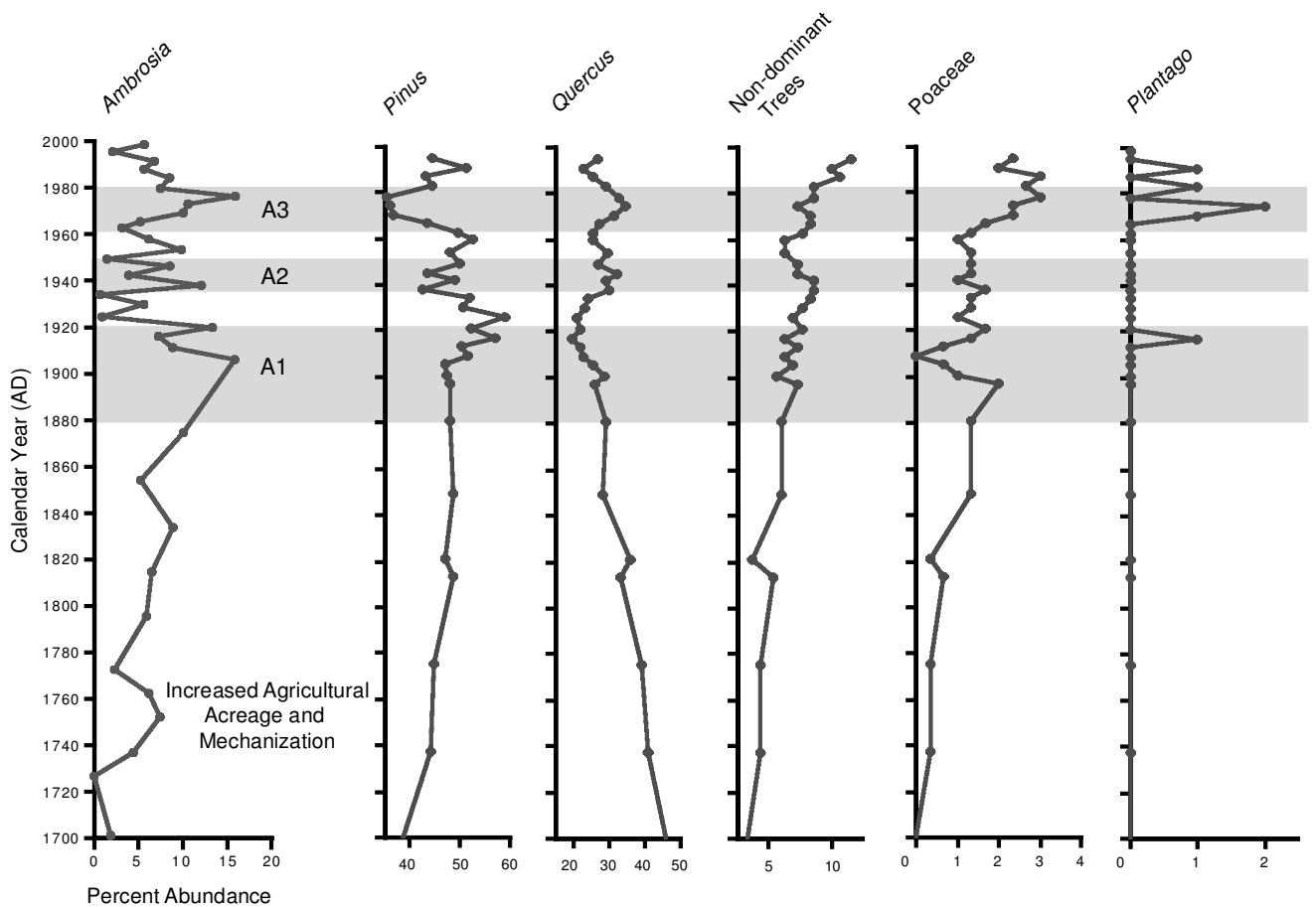


Figure 13 Abundance of pollen of major plant groups and the dinoflagellate cyst *Spiniferites mirabilis* since AD 1700, core MD99-2209, Chesapeake Bay, near mouth of Rhode River.

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