Climatic variability in the eastern United States over the past millennium from Chesapeake Bay sediments

T. Cronin
D. Willard
A. Karlsen
S. Ishman
S. Verardo
J. McGeehin — U.S. Geological Survey, Reston, Virginia 20192, USA
R. Kerhin* Maryland Geological Survey, Baltimore, Maryland 21218, USA
C. Holmes U.S. Geological Survey, St. Petersburg, Florida 33701, USA
S. Colman U.S. Geological Survey, Woods Hole, Massachusetts 02543, USA
A. Zimmerman Virginia Institute of Marine Science, Gloucester Point, Virginia 23062, USA

ABSTRACT

Salinity oscillations caused by multidecadal climatic variability had major impacts on the Chesapeake Bay estuarine ecosystem during the past 1000 yr. Microfossils from sediments dated by radiometry (^{14}C, ^{137}Cs, ^{210}Pb) and pollen stratigraphy indicate that salinity in mesohaline regions oscillated 10–15 ppt during periods of extreme drought (low fresh-water discharge) and wet climate (high discharge). During the past 500 yr, 14 wet-dry cycles occurred, including sixteenth and early seventeenth century megadroughts that exceeded twentieth century droughts in their severity. These droughts correspond to extremely dry climate also recorded in North American tree-ring records and by early colonists. Wet periods occurred every ~60–70 yr, began abruptly, lasted <20 yr, and had mean annual rainfall ~25%–30% and fresh-water discharge ~40%–50% greater than during droughts. A shift toward wetter regional climate occurred in the early nineteenth century, lowering salinity and compounding the effects of agricultural land clearance on bay ecosystems.

Keywords: Holocene climate, benthic foraminifera, Chesapeake Bay, estuarine ecosystem, eastern U.S. precipitation, eastern U.S. climate.

INTRODUCTION

Twentieth century climatic variability is caused by ocean-atmosphere interactions (Latif and Barnett, 1994; Woodhouse and Overpeck, 1998), solar variability (Crowley and Kim, 1996; Lean and Rind, 1998), volcanic processes (Robock and Mao, 1995), climatological "noise," anthropogenic perturbations to Earth's atmosphere, or a combination of factors (Rind and Overpeck, 1993). Because instrumental records rarely exceed 100 yr, paleoclimate reconstructions can establish which factors are most important over multidecadal time scales, distinguish anthropogenic and natural causes, and test climate-model simulations of decadal- and centennial-scale variability (Latif, 1998). Paleoclimatic records from polar and tropical regions suggest that decadal- and centennial-scale climate changes are characteristic of the past millennium, a period that includes the Medieval Warm Period (ninth through fourteenth centuries) and the Little Ice Age (fifteenth through nineteenth centuries) (Hughes and Diaz, 1994; Overpeck et al., 1997). Yet, except for tree-ring records (i.e., Cook and Jacoby, 1983; Stahle et al., 1998), multicentury records are sparse from most of the United States, and the ecological impacts of Holocene climate variability remain poorly known.

We investigated the past millennium of paleoclimatic history of the mid-Atlantic region preserved in the sedimentary record of Chesapeake Bay. Chesapeake Bay, the largest estuary in the United States, is a 320-km-long, 20–40-km-wide, 6500 km^2 drowned river valley (average and maximum depths of 8.5 m and 53 m, respectively) inundated by the late Quaternary sea-level rise ca. 6–8 ka (Colman and Mixon, 1988). The bay's thick Holocene section (Colman and Halka, 1989) contains micropaleontological and geochemical evidence for changes in fresh-water and sediment inflow, salinity, dissolved oxygen, and temperature (Cronin et al., 1999a). Its record of decadal- and centennial-scale variability holds promise for understanding impacts of past and climate change.

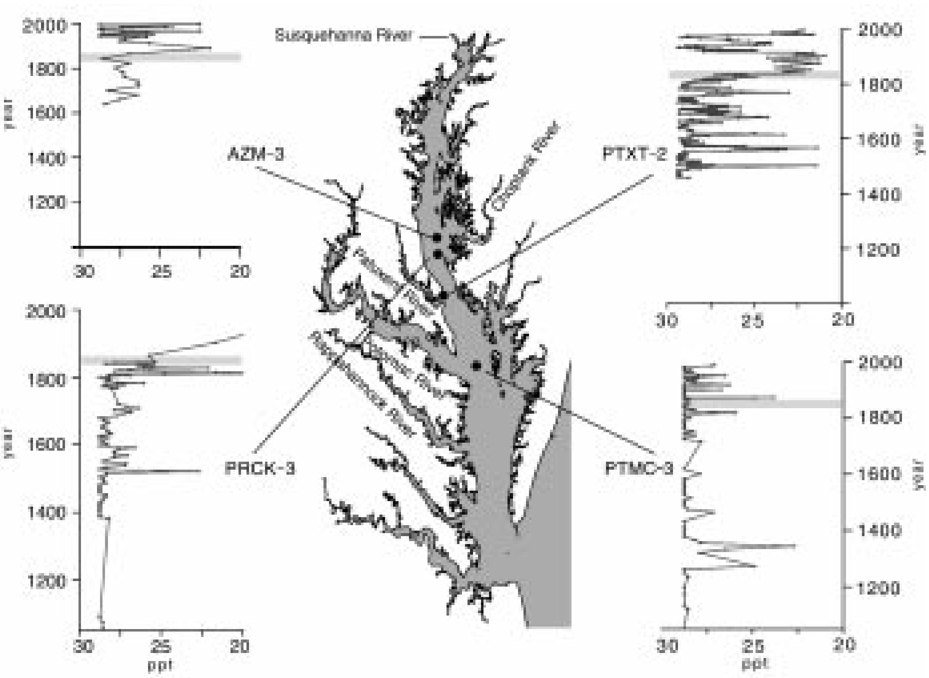

Figure 1. Cores in Chesapeake Bay. Sites PTMC-3, PRCK-3, AZM-3 are located in main channel, and PTXT-2 is near Patuxent River mouth. Paleosalinity estimates are based on percentage of *Elphidium* in Holocene sediments using salinity model in Figure 2. Ages are based on radiometric dating and pollen stratigraphy; sedimentation at PRCK-3 was irregular. Shallow-water sites (PTXT-2, AZM-3) were more sensitive to salinity variability than deeper water sites (PRCK-3, PTMC-3).

*Deceased.

Data Repository item 20004 contains additional material related to this article.

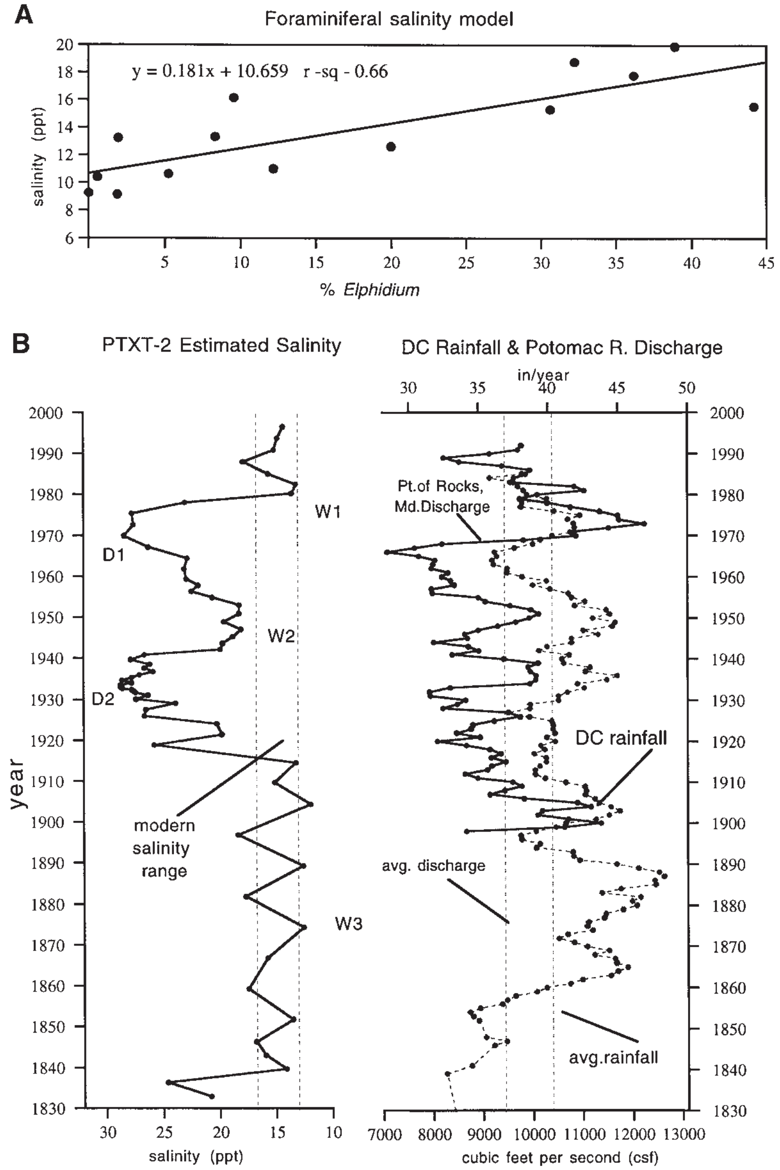

Figure 2. A: Relationship between salinity and *Elphidium* percentage in upper 1 cm of sediment in Chesapeake Bay and its tributaries in salinity measured at 10–20 ppt. B: Historical rainfall in Washington, D.C., from NOAA's National Climate Data Center (www.noaa.ncdc.gov/ghcn/ghcnV1.CLIMVIS.html) and Potomac River discharge at Point of Rocks, Maryland (U.S. Geological Survey Station 01638500), and paleosalinity at PTXT-2 based on foraminifers. D1–D2 are dry periods; W1–W3 are wet periods. Dotted lines show approximate modern range in salinity and precipitation at core site over past decade. Discrepancies between paleosalinity estimate and Washington rainfall from 1830 to 1860 may be due to age dating uncertainty or inadequate rainfall records.

MATERIAL AND METHODS

We studied cores across a salinity gradient of <10–22 ppt in the mesohaline bay (Fig. 1; Table 1[1]). Chesapeake Bay salinity is influenced by seasonal and annual fresh-water discharge, which is largely a function of precipitation in the watershed (Najjar, 1999; Cronin et al., 1999a). Monitoring records in the bay just off the Potomac River, for example, show that minimum and maximum wet-season salinities from 1984 to 1997 were 10 and 18 ppt, respectively; dry-season minima and maxima were 17 and 22 ppt, respectively. Post-1950 data show that interdecadal midbay surface-water salinity extremes ranged from maxima of 15–16 ppt during the dry 1960s to minima of <5–7 ppt during the wet 1970s. Thus, this region is sensitive to both short- and long-term fresh-water discharge variability and is well suited to record climatically driven salinity oscillations.

Lithologic and X-ray radiograph studies show that sediments in the midbay consist mostly of fine-grained organic-rich mud and fine sand (Kerhin et al., 1998). Sediment-accumulation rates and age models were obtained by using pollen stratigraphy and radiometric dating. Prior studies (Brush et al., 1982; Brush, 1989; Cooper and Brush, 1991) provide a framework for post-colonial pollen stratigraphy. Two events are especially useful markers: the appearance of abundant (>10% of the total assemblage) *Ambrosia* (ragweed) between ca. A. D. 1800 and 1850 (the agricultural horizon) and the disappearance of *Castanea dentata* (chestnut) ca. 1930. We converted 22 ^{14}C dates on shells into calendar years using the model of Stuiver and Reimer (1993). Ages for sediments deposited over the past century at sites PTXT-2 and PTMC-3 were obtained from short-lived radioisotopes ^{137}Cs and ^{210}Pb. A ^{137}Cs spike at 30 cm (ca. 1963) in PTXT-2-G-4, 27 cm in PTMC-3-P-2, and 56 cm in AZM-3 is corroborated by ^{210}Pb profiles. Errors associated with each dating method are ~±20 yr for the agricultural horizon, ~±5 yr for the ^{137}Cs spike, ~±5–20 yr for the ^{210}Pb profile, and ±62–145 yr for radiocarbon dates. Age and core site data are available from the GSA data repository (see footnote 1). From dated horizons, we computed calendar-year age models that are corroborated by instrumental and tree-ring records (see following).

We used the benthic foraminifer *Elphidium* and several ostracode species to trace salinity variability (Ellison and Nichols, 1976; Cronin et al., 1999b). A linear model to calculate paleosalinity from the relative frequencies of *Elphidium* on the basis of its frequencies in surface sediment is expressed as $S = (0.181 \times P) + 10.66$ ($r^2 = 0.66$), where S = salinity and P = percentage of *Elphidium* for salinity between 10 and 20 ppt (Fig. 2A). For P values >45%, the model extrapolates salinity on the basis of populations from the lower bay where *Elphidium* is more abundant (P = ~50%–85% of the total foraminiferal assemblage) at salinities of 20–30 ppt (Ellison and Nichols, 1976). At salinities below ~10 ppt, *Elphidium* is absent (P = 0%). Other environmental factors (i.e., dissolved oxygen, substrate) account for the rest of the variability in *Elphidium* abundance.

To cross-check this model, we compared paleosalinity trends from core PTXT-2-P-5 to instrumental records of Washington, D.C., rainfall and Potomac River discharge for the past 175 and 90 yr, respectively (Fig. 2B). For downcore analyses, ~100 foraminifers were isolated from 40–60 g at 2 cm spacing. Confidence intervals for 100 individuals are ±5% to 9%, depending on the species' relative frequency; thus, temporal changes in relative frequencies of 10% to >80% are statistically significant (Buzas, 1990). Although the earliest rainfall records may not be accurate, the decadal trends revealed by the *Elphidium* record still generally match Washington rainfall and Potomac River discharge. Three periods of reduced salinity (W1–W3) correspond to periods of greater than average discharge and rainfall (1860–1910, 1930–1950s, and post-1970). Dry periods (D1, D2) during the 1930s and 1960s cor-

[1]GSA Data Repository item 20004, Table 1, Site and geochronologic data for Chesapeake Bay cores, is available on request from Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, editing@geosociety.org, or at www.geosociety.org/pubs/drpint.htm.

respond to times of high bay salinity. Similar but damped patterns are evident at the other sites (Fig. 1). A salinity excursion of 10–15 ppt (i.e., W1 and W2) equals an increase in annual Potomac River discharge of ~2000–4000 cubic feet per second (cfs) (35.315 $ft^3 = 1\ m^3$) (25%–50%) over dry periods, and a corresponding 7–10 in/yr (18–25 cm/yr, 20%–30%) increase in rainfall. These foram-based paleosalinity variations are quantitatively similar to those measured over the past 50 yr from this region.

PAST MILLENNIUM OF CLIMATE VARIABILITY

Reconstructed salinity for the past millennium reveals several first-order trends (Fig. 1). First, over centuries, it is possible to trace a north to south and shallow to deep salinity gradient, as measured by the percentage of *Elphidium*. Farther south and deeper water core sites have on average greater proportions of *Elphidium* and thus higher estimated salinity. The shallowest site (site PTXT-2, 12 m water depth) was the most sensitive to decadal salinity variability; the deepest and most oceanward site (PTMC-3) was the least sensitive. This pattern reflects the influence of saline Atlantic water in the deep channel, even during wet periods, and the greater sensitivity of shallower regions to changes in fresh-water flow. Second, pre-eighteenth century salinity was on average several parts per thousand higher than salinity during the past two centuries; all cores show evidence for low frequencies of *Elphidium* since about 1800–1850, although this signature may partially reflect the influence of nineteenth and twentieth century land-use changes. Third, during part of the Medieval Warm Period (ca. A.D. 1250–1350), site PTMC-3 underwent a period of low salinity signifying extremely wet climate. However, further study of the Medieval Warm Period variability is needed.

Fourth, site PTXT-2 provides a detailed sedimentary record of decadal variability in Chesapeake Bay salinity over the past 550 yr, revealing 14 oscillations between high (25–28 ppt) and low (14–16 ppt) salinity since A.D. 1450 (Fig. 3, W1–W14 and D1–D14). These oscillations signify multidecadal periods of wet and dry regional climate. During the most extreme excursions (W14, W13, W12, W10, W6, W3, W1), the frequency of *Elphidium* was low, taxa tolerant of lower salinity predominated (*Ammobaculites*), and salinity fell by ~10–15 ppt. Conversely, dry intervals are prolonged periods of *Elphidium* dominance and high salinity and low rainfall sustained over several decades; these intervals correspond to periods referred to as "mega-droughts" by Woodhouse and Overpeck (1998) and "sustained severe droughts" by Meko et al. (1995). There also appears to be a quasicyclic pattern to the major salinity excursions with a period of ~60–70 yr. If the relationships among rainfall, streamflow and salinity were similar to those for the nineteenth and twentieth centuries, these salinity oscillations equate to changes in regional precipitation of ~25%–30% and streamflow, of 40%–50%. Another important characteristic of these salinity cycles is that the shift from dry to wet climate was extremely rapid, occurring over about a decade.

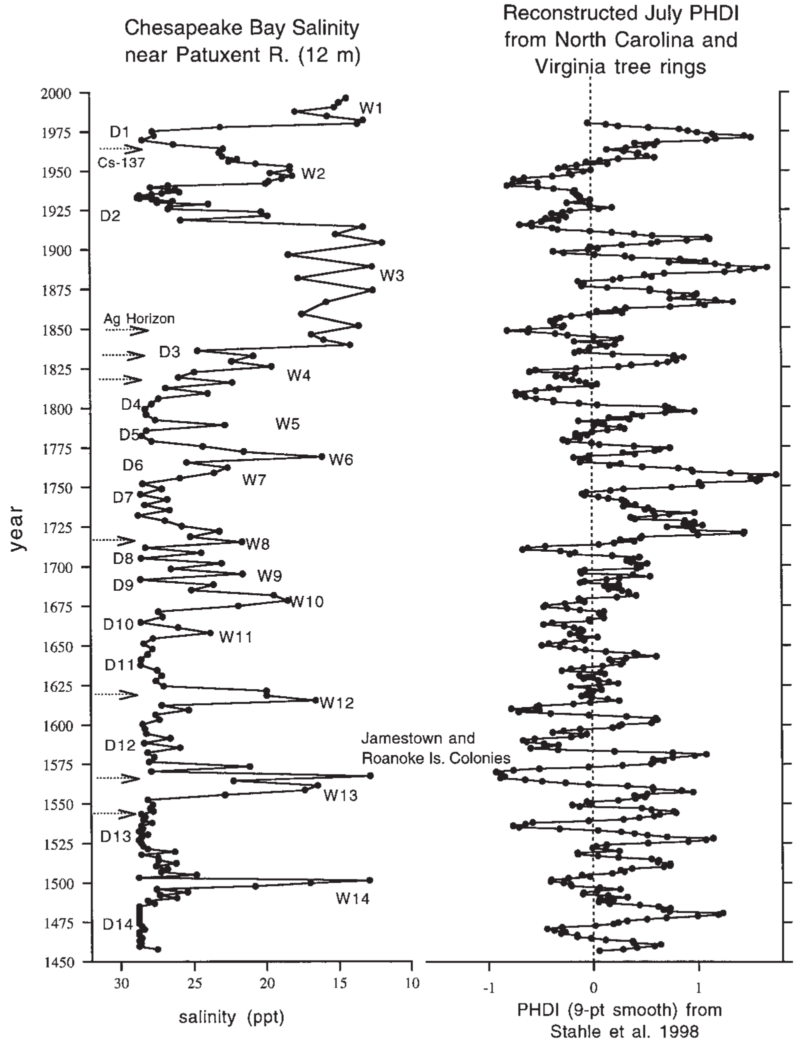

Figure 3. Comparison of 500 yr salinity record at PTXT-2-P-5 to nine point smoothed tree-ring Palmer Hydrological Drought Index (PHDI) for area (from Stahle et al., 1998). Negative (positive) PHDI values signify dry (wet) conditions. Wet periods (< 25 ppt salinity) are W1–W14; dry periods (high salinity) are D1–D14. Dashed lines connect major salinity and tree-ring events. ^{14}C dates are indicated by arrows.

Large-scale fluctuations in salinity are also indicated by ostracode assemblages. *Cytheromorpha newportensis* (salinity range ~>16–32 ppt) oscillated approximately in phase with *Elphidium*. During droughts, marine species (*Protocytheretta edwardsi*, *Actinocythereis captionis*, *Pellucistoma magniventra*) inhabited mid-Chesapeake Bay, reflecting diminished fresh-water influx; by the late 1700s, the brackish-water species *Perissocytheridea brachyforma* (<15 ppt) composed 30% to >50% of the assemblage.

MEGADROUGHTS IN THE EASTERN UNITED STATES

It is useful to compare the Chesapeake Bay salinity record to other climatic records for North America. Figure 3 compares the PTXT-2 salinity record to the Palmer Hydrological Drought Index (PHDI) for July reconstructed from tree rings from southern Virginia and northeastern North Carolina (Stahle et al., 1998). The PHDI record was smoothed for easier comparison to the sedimentary record. Although a precise correlation between paleosalinity and PHDI is difficult to achieve because of dating error, there are noteworthy similarities between the two records. First, there is strong evidence that sustained droughts during the middle to late sixteenth and early seventeenth century were more severe than twentieth century droughts. Events D14, D13, D12, and D11 represent sixteenth and early seventeenth century periods of high salinity corresponding to continental-scale droughts from Mexico, California, the Colorado Plateau, and the southeastern United States (Meko et al., 1995; Woodhouse and Overpeck, 1998). Their existence in the Chesapeake record demonstrates that decadal climate variability had a severe impact on this large estuary. These droughts probably had severe impacts on Spanish and

English colonists during the 1560s, ca. 1587–1589, and ca. 1606–1612 (Stahle et al., 1998).

Second, from the late seventeenth until the early nineteenth century, more frequent lower amplitude salinity excursions (W11–W4) signify periods of high regional precipitation and PHDI values usually >0. Stahle et al. (1988; see also Stahle and Cleaveland, 1992) found progressively higher PHDI values after A.D. 1750 in bald cypress records from the southeastern United States.

Third, Chesapeake and tree-ring records indicate that a climatic transition occurred about 1825–1840 when a prolonged low-salinity phase began (W3), and PHDI values exceeded 1 for much of the period 1825–1910 (Stahle et al., 1998). On the basis of 74 tree ring records from the eastern United States, Cook and Mayes (1987) inferred a major atmospheric reorganization about 1836 that apparently initiated this period of wet climate. They also showed that the period 1836–1879 was characterized by high tree growth in mid-Atlantic regions, also consistent with greater mean annual precipitation and low bay salinity. Thus, although land-use changes influenced parts of Chesapeake Bay (Cooper and Brush, 1991), paleosalinity and tree-ring records suggest that during the nineteenth century Chesapeake Bay was also influenced by continental-scale climatic changes.

Although it is premature to link variability in Chesapeake salinity and eastern U.S. precipitation directly to specific causes, the patterns suggest "teleconnections" that can, at least conceptually, be related to decadal-scale climatological processes operating today. For example, a positive Pacific North America atmospheric pattern causes more meridional mid-tropospheric flow in the eastern United States when the polar front and jet stream are pushed farther south than usual and the eastern U. S. coast has greater spring precipitation (Yarnal and Leathers, 1988; Henderson and Vega, 1996). Sixteenth and seventeenth century droughts might represent the opposite situation (negative Pacific North America pattern), in which zonal flow induces reduced precipitation in the watershed. Teleconnections to the North Atlantic oscillation (Hurrell, 1995) may also influence the northern part of the watershed when positive values correlate with increased precipitation (Vega et al., 1998). Simulated (Delworth et al., 1993) and observed (Kushnir, 1994) variability in North Atlantic Ocean sea-surface temperature, with spectral peaks at 40–80 yr, is similar to the 60–70 yr frequency of Chesapeake wet periods. Because the Pacific North America and North Atlantic oscillation are linked to wind-driven oceanic circulation in subtropical gyres and/or thermohaline circulation (Latif, 1998), climate variability in the Chesapeake region may ultimately be linked to oceanic factors, although this hypothesis needs to be tested. Estimates of past precipitation and river discharge changes based on paleosalinity reconstructions are roughly comparable to increases in Susquehanna River basin rainfall and discharge due to doubling of atmospheric CO_2, as predicted by regional climate (Crane and Hewitson 1998) and hydrological models (Najjar, 1999).

ACKNOWLEDGMENTS

R. Najjar, A. Vega, H. Dowsett, G. Dwyer, B. Sen Gupta, B. Cecil, D. Stahle, M. Cleaveland, and an anonymous reviewer gave comments on the manuscript. We thank H. Kitazato, J. Bernhard, M. Buzas, and B. Sen Gupta for comments on foram ecology. Funded by U.S. Geological Survey, Maryland Geological Survey, and National Science Foundation grant OCE-9521170 to Zimmerman.

REFERENCES CITED

Brush, G. S., 1989, Rates and patterns of estuarine sediment accumulation: Limnology and Oceanography, v. 34, p. 1235–1246.

Brush, G. S., Martin, E. A., DeFries, R. S., and Rice, C. A., 1982, Comparisons of ^{210}Pb and pollen methods for determining rates of estuarine sediment accumulation: Quaternary Research, v. 18, p. 196–217.

Buzas, M. A., 1990, Another look at confidence limits for species proportions: Journal of Paleontology, v. 64, p. 842–843.

Colman, S. M., and Halka, J. P., 1989, Map showing Quaternary geology of the southern Maryland part of the Chesapeake Bay: U.S. Geological Survey Miscellaneous Field Studies Map MF-1948-C, scale 1:125,000.

Colman, S. M., and Mixon, R. B., 1988, The record of major Quaternary sea-level changes in a large coastal plain estuary, Chesapeake Bay, eastern United States: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 68, p. 99–116.

Cook, E. R., and Jacoby, G. C., 1983, Potomac River streamflow since 1730 as reconstructed by tree rings: Journal of Climate and Applied Meteorology, v. 22, p. 1659–1672.

Cook E. R., and Mayes, P. 1987, Decadal-scale patterns of climatic change over eastern North America inferred from tree rings, *in* Berger, W. H., and Labeyrie, L. D., eds., Abrupt climate change; evidence and implications: NATO Advanced Study Institutes Series C, no. 216, p. 61.

Cooper, S. R., and Brush, G. S., 1991, Long term history of Chesapeake Bay anoxia: Science, v. 254, p. 992–996.

Crane, R. G., and Hewitson, B. C., 1998, Doubled CO_2 precipitation changes for the Susquehanna Basin: Down-scaling from the genesis general circulation model: International Journal of Climatology, v. 18, p. 65–76.

Cronin, T., Colman, S., Willard, D., Kerhin, R., Holmes, C., Karlsen, A., Ishman, S., and Bratton, J., 1999a., Interdisciplinary project probes Chesapeake Bay down to the core: Eos (Transactions, American Geophysical Union), v. 80, p. 237, 240–241.

Cronin, T. M., Wagner, R. S., and Slattery, M., eds., 1999b, Microfossils from Chesapeake Bay sediments: Illustrations and species database: U.S. Geological Survey Open-File Report 99-45, 159 p.

Crowley, T. J., and Kim, K.-Y., 1996, Comparison of proxy records of climate change and solar forcing: Geophysical Research Letters, v. 23, p. 359–362.

Delworth, T., Manabe, S., and Stouffer, R. J., 1993, Interdecadal variations of the thermohaline circulation in a coupled ocean-atmosphere model: Journal of Climate, v. 9, p. 2356–2375.

Ellison, R. L., and Nichols, M. M., 1976, Modern and Holocene foraminifera in the Chesapeake Bay region, *in* Schaefer, C. T., and Pelletier, B. R., eds., International Symposium on Benthonic Foraminifera of Continental Margins, Part A, Ecology and Biology: Halifax, Nova Scotia, Canada, Maritime Sediments Special Publication 1, p. 131–151.

Henderson, K. G., and Vega, A. J., 1996, Regional precipitation variability in the southern United States: Physical Geography, v. 17, p. 93–112.

Hughes, M. K., and Diaz, H. F., 1994, Was there a "Medieval Warm Period," and if so, where and when?: Climatic Change, v. 26, p. 109–142.

Hurrell, J. W., 1995, Decadal trends in the North Atlantic oscillation: Regional temperatures and precipitation: Science, v. 269, p. 676–679.

Kerhin, R. T., Williams, C., and Cronin, T. M., 1998, Lithologic descriptions of piston cores from Chesapeake Bay, Maryland: U.S. Geological Survey Open-File Report 98-787, 21 p.

Kushnir, Y., 1994, Interdecadal variations in North Atlantic sea surface temperature and associated atmospheric conditions: Journal of Climate, v. 7, p. 141–157.

Latif, M., 1998, Dynamics of interdecadal variability in coupled ocean-atmosphere models: Journal of Climate, v. 11, p. 602–624.

Latif, M., and Barnett, T. P., 1994, Causes of decadal climate variability over the North Pacific and North America: Science, v. 266, p. 634–637.

Lean, J., and Rind, D., 1998, Climate forcing by changing solar radiation: Journal of Climate, v. 11, p. 3069–3094.

Meko, D., Stockton, C. W., and Boggess, W. R., 1995, The tree-ring record of severe sustained drought: Water Resources Bulletin, v. 31, p. 789–801.

Najjar, R. G., 1999, The water balance of the Susquehanna River Basin and its response to climate change: Journal of Hydrology, v. 219, p. 7–19.

Overpeck, J., Hughen, K., Hardy, D., Bradley, R., Case, R., Douglas, M., Finney, B., Gajewski, K., Jacoby, G., Jennings, A., Lamoureux, S., Lasca, A., MacDonald, G., Moore, J., Retelle, M. Smith, S., Wolfe, A., and Zielinski, G., 1997, Arctic environmental change of the last four centuries: Science, v. 278, p. 1251–1256.

Robock, A. D., and Mao, J., 1995, The volcanic signal in surface temperature observations: Journal of Climate, v. 8, p. 1086–1103.

Stahle, D. W., and Cleaveland, M. K., 1992, Reconstruction and analysis of spring rainfall over the southeastern U.S. for the past 1000 years: American Meteorological Society Bulletin, v. 73, p. 1947–1961.

Stahle, D. W., Cleaveland, M. K., Blanton, D. B., Therrell, M. D., and Gay, D. A., 1998, The lost colony and Jamestown droughts: Science, v. 280, p. 564–567.

Stahle, D. W., Cleaveland, M. K., and Hehr, J. G., 1988, North Carolina climate changes reconstructed from tree rings: A.D. 372 to 1985: Science, v. 240, p. 1517–1519.

Stuiver, M., and Reimer, P. J., 1993, Extended ^{14}C data base and revised CALIB 3.0 ^{14}C age calibration program: Radiocarbon, v. 35, p. 215–230.

Vega, A., Sui, C. H., and Lau, K. M., 1998, Interannual to interdecadal variations of the regionalized surface climate of the United States and relationships to generalized flow patterns: Physical Geography, v. 19, p. 271–291.

Woodhouse, C. A., and Overpeck, J. T., 1998, 2000 years of drought variability in the central United States: American Meteorological Society Bulletin, v. 79, p. 2693–2714.

Yarnal, B., and Leathers, D. J., 1988, Relationships between interdecadal and interannual climatic variations and their effect on Pennsylvania climate: Association of American Geographers Annals, v. 78, p. 624–641.

Manuscript received May 10, 1999
Revised manuscript received August 12, 1999
Manuscript accepted August 17, 1999