ISOTOPIC TEMPERATURES FROM THE EARLY AND MID-PLIOCENE OF THE US MIDDLE ATLANTIC COASTAL PLAIN, AND THEIR IMPLICATIONS FOR THE CAUSE OF REGIONAL MARINE CLIMATE CHANGE

ANDREW L.A. JOHNSON,[1*] ANNEMARIE VALENTINE,[1†] MELANIE J. LENG,[2] HILARY J. SLOANE,[2] BERND R. SCHÖNE,[3] and PETER S. BALSON[4††]

[1]Department of Natural Sciences, College of Life and Natural Sciences, University of Derby, Kedleston Road, Derby DE22 1GB, UK, a.l.a.johnson@derby.ac.uk; [2]NERC Isotope Geoscience Facilities, British Geological Survey, Nicker Hill, Keyworth NG12 5GG, UK, mjl@bgs.ac.uk, hjs@bgs.ac.uk; [3]Institute of Geosciences, University of Mainz, Johann-Joachim-Becher-Weg 21, 55128 Mainz, Germany, schoeneb@uni-mainz.de; [4]British Geological Survey, Nicker Hill, Keyworth NG12 5GG, UK

[*] Corresponding author

[†] Current address: Student Advice and Support Services, Loughborough University, Loughborough LE11 3TU, UK, drmitzyvalentine@gmail.com

[††] Current address: 6 Cherryholt Lane, East Bridgford NG13 8BN, UK, cragpad@btinternet.com

RRH: PLIOCENE MARINE TEMPERATURES

LLH: JOHNSON ET AL.

Keywords: sclerochronology, paleotemperature, paleoclimate, paleoceanography, Pliocene

ABSTRACT

Mean seasonal extreme temperatures on the seafloor calculated from the shell $\delta^{18}O$ of the scallop *Placopecten clintonius* from the basal part of the early Pliocene Sunken Meadow Member (Yorktown Formation) in Virginia are very similar to those from the same horizon at the latitude of Cape Hatteras in North Carolina (~210 km to the south). The lowest and highest temperatures calculated from each shell (using $\delta^{18}O_{seawater}$ = +0.7‰) give mean values for winter and summer of 8.4 ± 1.1 °C (± 1σ) and 18.2 ± 0.6 °C in Virginia, and 8.6 ± 0.4 °C and 16.5 ± 1.1 °C in North Carolina (respective median temperatures: 13.3 °C and 12.6 °C). Patterns of ontogenetic variation in $\delta^{18}O$, $\delta^{13}C$ and microgrowth increment size indicate summer water-column stratification in both areas, with summer surface temperatures perhaps 6 °C higher than on the seafloor. The low winter paleotemperatures in both areas are most simply explained by the greater southward penetration of cool northern waters in the absence of a feature equivalent to Cape Hatteras. The same current configuration but a warmer general climate can account for the high benthic seasonal range (over 15.0 °C in some cases) but warmer median temperatures (15.7-21.3 °C) derived from existing $\delta^{18}O$ data from scallops of the higher Yorktown Formation (using $\delta^{18}O_{seawater}$ = +0.7‰ for the upper Sunken Meadow Member and $\delta^{18}O_{seawater}$ = +1.1‰ for the mid-Pliocene Rushmere, Morgarts Beach and Moore House members). Existing $\delta^{18}O$ data from the infaunal bivalve *Mercenaria* of the Rushmere Member yields a similarly high median temperature (21.6 °C) but a low seasonal range (9.2 °C), pointing to the periodic influence of warm currents, possibly at times when the Gulf Stream was exceptionally vigorous.

INTRODUCTION

The Pliocene (5.33–2.58 Ma) contains the most recent interval (~3.3-3.0 Ma) in which global mean surface temperature was significantly higher than present (by 1.9-3.6ºC; Masson-Delmotte et al., 2013). This interval – the mid-Pliocene or, more strictly, mid-Piacenzian Warm Period (both abbreviated to MPWP) – has been the focus of study for nearly 30 years by the Pliocene Research, Interpretation and Synoptic Mapping (PRISM) group of the United States Geological Survey (Dowsett et al., 2016). It has been used extensively as a test-bed for numerical models of an Earth with relatively high atmospheric CO_2 because concentrations of this greenhouse gas were well above pre-industrial interglacial values according to most reconstructions (e.g., Masson-Delmotte et al., 2013; Martínez-Boti et al., 2015), yet many other large-scale aspects of paleogeography (e.g., continental positions, orography, ocean current patterns) were similar to now. Model outputs for the MPWP are consistent with proxy estimates of temperature at the global scale and for many regions. However, for certain parts of the North Atlantic, proxy estimates are substantially higher (Dowsett et al., 2012, 2013), indicating either inadequacies in the models (including the boundary conditions used) or the proxy data. At some relatively high-latitude (>45 ºN) sites in the North Atlantic, congruent evidence of substantial warming (mean annual sea surface temperature >5 ºC above present) is available from multiple proxies (foraminiferal assemblage composition, foraminiferal Mg/Ca ratios and alkenone unsaturation index; Dowsett et al., 2012), suggesting that it is the model estimates that are inaccurate. The high-latitude warmth has been ascribed (e.g., Dowsett et al., 1992; Cronin and Dowsett, 1996) to stronger northward transfer of heat by ocean currents than now, but model outputs do not support this (Fedorov et al., 2013; Zhang et al., 2013).

Enhanced ocean transport of heat during the Pliocene has also been inferred from proxy temperature data for lower latitudes in the North Atlantic region, on the Middle Atlantic Coastal Plain of the USA. At present, mean winter minimum and mean summer maximum sea surface temperatures lie in the ranges 5.0-10.0 ºC and 22.5-27.5 ºC, respectively, at coastal to outer shelf locations off northern North Carolina (north of Cape Hatteras) and Virginia (Table 1, stations ORIN7, DUCN7, 44006, 44014, CHLV2, CBBV2, KPTV2, OCIM2, 44009). However, during deposition of the upper (Rushmere, Morgarts Beach and Moore House) members of the Pliocene Yorktown Formation, biotic assemblage evidence (see below) points to much warmer conditions in this area: specifically, winter minimum temperatures above 10 ºC. Greater warmth north of Cape Hatteras during the interval concerned, which overlaps the MPWP (Fig. 1), has been ascribed to more vigorous northward flow of warm currents, supplying more heat (Cronin and Dowsett, 1996; Knowles et al., 2009; Williams et al., 2009; Winkelstern et al., 2013). However, Ward et al. (1991) attributed the greater warmth simply to higher sea level and the absence of a barrier equivalent to modern Cape Hatteras, allowing free passage of warm waters northwards across the shelf. If this explanation is correct, it has implications not only for our understanding of the northward expansion of warm conditions on the Atlantic Coastal Plain but also in the wider North Atlantic: elevated temperatures at higher latitudes might reflect a more northward trajectory rather than increased strength of warm-current flow. However, both of these explanations would be called into question if the higher temperatures recognised north of Cape Hatteras are not the result of greater warm-current influence but of generally warmer conditions, caused by some other factor (e.g., increased atmospheric CO_2).

We attempt to identify the cause of mid-Pliocene warming north of the latitude of Cape Hatteras using estimates of seasonal marine paleotemperature from the oxygen isotopic ($\delta^{18}O$) composition of bivalve shells from the Yorktown Formation. Most of the data derives from scallops, which are a propitious subject because in early ontogeny they typically grow rapidly and throughout the year, thus preserving an easily recoverable record of the full range of seasonal temperature variation (e.g., Johnson et al., 2009; Chute et al., 2012), and kinetic and 'vital' effects seem to be at most small (e.g., Barrera at al., 1990; Hickson et al., 1999; Owen et al., 2002a, b). In addition, most scallops live only in fully marine conditions, thus reducing uncertainties about the value to use for $\delta^{18}O_{seawater}$ in the isotopic temperature equation. We present new data from the genus *Placopecten* of the early Pliocene Sunken Meadow Member in Virginia and North Carolina, using this to formulate a model of climate and water circulation for a time when coastline geometry was much like that in the mid-Pliocene (i.e., no 'Cape Hatteras') yet sea temperatures were lower according to independent evidence. We then derive a prediction for benthic seasonal temperature range if the subsequent warming of marine climate in the area was due to an increase in the influence of warm currents. We test this hypothesis through a reanalysis of existing $\delta^{18}O$ data from other scallops (*Chesapecten* and *Carolinapecten*) of the Yorktown Formation, both from a horizon in the Sunken Meadow Member above that of the *Placopecten* shells and from all three of the higher (mid-Pliocene) members of the Yorktown Formation. Finally, we review existing mid-Pliocene isotopic temperature data from the infaunal (non-scallop) bivalve *Mercenaria* and estimates of mid-Pliocene seasonal temperature range from bryozoan zooid-size variation, identifying and attempting to explain discrepancies with the scallop data and then making recommendations for further research. Since all the data discussed is in the form of ontogenetic or astogenetic profiles from mineralised, accretionarily-produced skeletal material, the investigation constitutes a case-study in sclerochronology (Schöne and Gillikin, 2013).

ISOTOPIC TEMPERATURES FROM *PLACOPECTEN CLINTONIUS* OF THE BASAL SUNKEN MEADOW MEMBER

Background Information

Stratigraphy.—The Sunken Meadow Member is the lowermost of the four members constituting the Yorktown Formation (Fig. 1), occurring in southeast Virginia and northeast North Carolina and averaging about 3 m thick. It mainly consists of fine-grained quartz sands, but towards the west the basal part is a medium to coarse sand (with a coarse lag deposit at the very base), while the finer deposits to the east are glauconitic in the north and phosphatic in the south; an abundant and diverse marine fauna occurs throughout (Ward and Blackwelder, 1980; Ward et al., 1991). Evidence of coeval marine sedimentation extending into South Carolina exists in the form of lag deposits at the base of younger units (Ward et al., 1991). Other marine deposits present farther south and considered to be of approximately the same age are 'Unit 11' of the lower Tamiami Formation, which occurs in southwest Florida (Williams et al., 2009), and the Wabasso Formation, which occurs in the subsurface of eastern Georgia (Huddlestun, 1988). The latter falls within planktonic foraminiferal biozone PL1 (of Berggren, 1973), which is dated at 4.9-3.7 Ma. The Sunken Meadow

Member itself is generally placed (e.g., Dowsett and Wiggs, 1992) within planktonic foraminiferal biozone N19 (of Blow, 1969), which is dated at 4.8 to ~3.5 Ma (essentially within the Zanclean) in the calibration of Berggren et al. (1985). Krantz (1991) tentatively suggested a more precise date of 4.5-4.4 Ma for the Sunken Meadow Member by relating the transgression associated with its deposition to a phase of warming and global ice-volume reduction identified in the deep-ocean $\delta^{18}O$ record (more fully and recently documented by Lisiecki and Raymo, 2005). The transgression was preceded by a phase of non-deposition or erosion (associated with a sea-level lowstand), such that the Sunken Meadow Member rests unconformably on either the Cobham Bay Member of the Eastover Formation, whose age is no younger than ~4.9 Ma (Krantz, 1991), or on the Miocene Pungo River Formation (Ward and Blackwelder, 1980).

Hydrography.—The transgression associated with the Sunken Meadow Member pushed the coastline up to 150 km west of its present position in southeast Virginia and northeast North Carolina (Ward et al., 1991; Krantz, 1991). Figure 2B shows its position according to Ward et al. (1991, fig. 16-4A). If Pliocene non-marine deposits in Maryland and Delaware correlate with the Sunken Meadow Member, the coastline may have been somewhat farther south in this area, nearer the Virginia state line, as reconstructed by Pazzaglia (1993). In a seismic study in eastern Albemarle Sound, North Carolina (north of Cape Hatteras), Mallinson et al. (2005) identified south- to southwestward-dipping clinoforms within lower Pliocene clastic sediments. They considered that these might represent a delta advancing from the northeast (implying that the coastline was recurved southward to this area) but also accepted that they might represent advancing shelf bedforms, an interpretation supported by foraminiferal evidence from the adjacent Mobil#1 well (Zarra, 1989). The eastward fining within the Sunken Meadow Member mentioned above is inconsistent with the existence of a delta advancing from the northeast but in full agreement with the position inferred by Ward et al. (1991) for the Sunken Meadow coastline in northern North Carolina (Fig. 2B). Farther south, the Mid-Carolina Platform High (centred on the North/South Carolina state line and a persistent influence on Cenozoic sedimentation; Riggs and Belknap, 1988) probably caused a reduction in water depth, but the lag deposits referred to above argue against emergence. The coastline therefore lay some distance inland of its present position (Fig. 2B). In summary, while the exact form of the coastline during Sunken Meadow deposition may have differed slightly from the configuration in Figure 2B, its shape was only mildly curvilinear, lacking an eastward protrusion as large as modern Cape Hatteras (whose tip lies in North Carolina at ~35°N, some 140 km south of the North Carolina-Virginia state line). Consequently neither northward- nor southward-flowing currents would have been deflected to the east as they are now (Fig. 2A).

At present, the Gulf Stream is a very warm and rapid western boundary current flowing northward above the Florida-Hatteras Slope to Cape Hatteras, where it diverges from the continental margin. Closer to the coast (on the shelf) current flow is weaker and more variable in direction, but still generally towards the north (Bumpus, 1973; Atkinson et al., 1983). This flow has been termed the Carolina Coastal Current (Cronin, 1988). At times surface intrusions from the Gulf Stream bring very warm water onto the North Carolina shelf south of Cape Hatteras (Atkinson, 1977). Farther south in the South Atlantic Bight (SAB; Cape Hatteras to Cape Canaveral) there also occur intrusions of deeper, relatively cool (and more nutrient-rich) Gulf Stream water onto the shelf, but these are relatively infrequent in the northern SAB (Atkinson et al., 1983). In the southern part of the Middle Atlantic Bight (MAB; Cape Hatteras to Cape Cod), there is typically a weak southward flow on the shelf (Bumpus, 1973), termed the Virginia Coastal Current (Cronin, 1988). This is approximately paralleled by a similar surface current above the upper and middle zones of the continental slope, in the western part of the Slope Sea between the shelf and Gulf Stream (Csanady and Hamilton, 1988; Böhm et al., 2006).

In the absence of a feature analogous to Cape Hatteras during Sunken Meadow deposition, it is reasonable to assume that there was no region of very steep latitudinal gradient in sea-surface temperature (SST) as there is now adjacent to the Cape because of the meeting of warm northward-flowing and cool southward-flowing water. However, even if the Gulf Stream was as strong as at present (suggested by evidence of early Pliocene submarine erosion on the Florida-Hatteras Slope; Pinet and Popenoe, 1985), it is not certain that warm water would have extended north into the area of present-day Virginia. The main current might still have turned eastwards, in the manner of all western boundary currents at 30-40° from the equator, so surface intrusions onto the Virginia shelf would have been rare. Also, any equivalent of the Carolina Coastal Current might not have been strong enough to displace cool southward-flowing waters, given that with the present coastal configuration northern shelf waters penetrate as far south as Cape Fear (Fig. 2B) about 10% of the time through wind-forcing (Pietrafesa et al., 1994). In fact, the Gulf Stream may have been less strong than now (i.e., more like the Brazil Current and East Australian Current, which are relatively weak western boundary currents) during deposition of the Sunken Meadow Member, due to incomplete development of the Central American Isthmus (Cronin and Dowsett, 1996; Schmidt, 2007). Hence, supply of warm water by currents into the area of present-day Virginia is still less certain. Ward and Blackwelder (1980) considered that at this time a structural high in the area of Cape Fear (essentially part of the Mid-Carolina Platform High) was sufficiently elevated to prevent northward-flowing warm waters entering the area of deposition of the Sunken Meadow Member, and Cronin and Dowsett (1996) took the view that the shelf off the eastern US was influenced only by southward-flowing cool currents in the early Pliocene.

Ward et al. (1991) considered that deposition of the Sunken Meadow Member took place in mid-shelf water depths of 20-40 m under the influence of upwelling (from the evidence of glauconite and phosphate). Purdy et al. (2001) and Fierstine (2001) also inferred upwelling but greater water depths (>50 m and >100m, respectively) from the fish fauna. The generally fine-sand sediments suggest neither strong wave action nor vigorous tidal currents, so it is possible that there was significant thermal stratification of the water column in summer, as now at similar latitudes and depths off eastern North America. For instance, at 30 m depth in the modern MAB (at approximately the latitude of the North Carolina-Virginia state line), water temperature is about 6 ºC less than at the surface in summer (Winkelstern et al., 2013). Spring freshwater run-off (reducing the salinity and hence density of surface waters) assists this stratification but also causes a temporary reduction in the salinity of bottom waters (from about 34.5 psu to 32.5 psu) as far offshore as the outer shelf (Krantz et al., 1988, fig. 1). In the modern SAB, bottom salinity is somewhat higher and essentially constant (at about 36 psu) through the year (Krantz et al., 1988, fig. 1). It is difficult to determine from first principles whether, in the absence of a feature analogous to Cape Hatteras, the area of deposition of the Sunken Meadow Member would have more resembled the modern MAB or SAB in respect of the influence of freshwater run-off. However, it seems improbable that there were significant differences over the area of deposition.

Marine Climate.—The classification of terrestrial climates has been extensively discussed and several schemes have been developed (e.g., 'K□ppen-Geiger', 'K□ppen-Trewartha'; Belda et al., 2014), with each climate defined by precise criteria. Marine climate (usually considered in terms of seasonal minimum and maximum temperatures in surface or shallow subsurface waters) has received less attention and the divisions recognised have been poorly and differently defined. For 'inner sublittoral' bottom waters of the modern eastern US shelf, Hazel (1971, 1988) recognised a 'mild temperate' marine climate north of Cape Hatteras (to 38 °N, and by implication to beyond 39 °N), defined by winter minimum temperatures in the range 2.5-5.0 °C and summer maximum temperatures in the range 20.0-22.5 °C. Winter minimum surface temperatures are actually higher than 5 °C (though less than 10 °C) at coastal to outer shelf locations up to 180 km north of Cape Hatteras (Table 1, stations ORIN7, DUCN7, 44006, 44014, CHLV2, CBBV2) and at depths of a few tens of metres temperatures are probably a degree or two warmer (though still less than 10 °C; Winkelstern et al., 2013). Summer maximum surface temperatures exceed 22.5 °C in this area and farther north (Table 1) but at a few tens of metres depth they may be within the range for a mild temperate climate given by Hazel (1971, 1988), or even the redefined range given by Krantz (1990; 17.5-20.0 °C), due to thermal stratification (Winkelstern et al., 2013). Hazel (1971, 1988) recognised a 'subtropical' marine climate at present immediately south of Cape Hatteras (to 35 °N, and by implication to beyond 33 °N), defined by winter minimum and summer maximum temperatures in the ranges 12.5-15.0 °C and 27.5-30.0 °C, respectively. His diagrams (1971, fig. 6; 1988, fig. 8) belie the figures given for summer maximum temperature, suggesting a range of 25.0-27.5 °C. This is the range given by Krantz (1990) for summer maximum temperature in a subtropical marine climate (with temperatures approaching 30 °C in shallow water), and the summer maximum surface temperature at a mid-shelf location about 8 km south of the latitude of Cape Hatteras (27 °C: Table 1, station DSLN7) is in agreement with this, as is the winter minimum surface temperature at this location (15 °C) with the winter minimum range given by Hazel (1971, 1988) for a subtropical marine climate. Farther south (to beyond 33 °N), summer maximum surface temperature is in the range 27.5-30.0 °C, even at offshore locations, and winter minimum surface temperature is below 12.5 °C at most coastal locations and above 15 °C at most offshore locations, resulting in a much smaller seasonal range offshore than is seen anywhere north of Cape Hatteras (Table 1). Ward et al. (1991) termed the modern marine climate north of Cape Hatteras 'cool temperate' but were in agreement with Hazel (1971, 1988) that the 'subtropical' zone to the south is not separated by a zone of 'warm temperate' conditions, characterised by Krantz (1990) as having winter minimum and summer maximum temperatures in the ranges 10.0-12.5 °C and 22.5-25.0 °C, respectively. Such a zone has, however, been recognised in southeast Virginia and northeast North Carolina during deposition of the upper (Rushmere, Morgarts Beach and Moore House) members of the Yorktown Formation (see below). By contrast, mollusk assemblages indicate a cool (= mild) temperate marine climate in this area during deposition of the underlying Sunken Meadow Member (Ward et al., 1991). Ostracod assemblages (Hazel 1971, 1988) support this in the sense that the summer maximum temperatures implied (no higher than 20 °C) are within the redefined range given by Krantz (1990) for a mild temperate marine climate. However, the winter minimum temperatures implied (no lower than 12.5 °C) are above the range for a mild temperate marine climate, and Hazel (1988) described the marine climate represented by the ostracod assemblages as warm temperate. Existing isotopic ($\delta^{18}O$) temperatures for the Sunken Meadow

Member (Krantz, 1990) support a mild temperate designation, maximum values (from profiles exhibiting a summer) being within the redefined range given by Krantz (1990) and minimum values being within the expanded range described above. The temperature estimates of Krantz (1990) are considered further below in the light of research into the oxygen-isotopic composition of ambient water.

Materials and Methods

We used complete and apparently pristine valves of *Placopecten clintonius* (Fig. 3A) from the basal 30 cm of the Sunken Meadow Member (L.W. Ward, personal communication, 2011). Four specimens from locations on the James River, Virginia, were provided from the collections of the Virginia Museum of Natural History: three (VA1, VA2, VA4; VMNH 93624, 93625, 93626, respectively) from Grove Wharf (1 in Fig. 2B) and one (VA3; VMNH 93627) from Claremont (2 in Fig. 2B), which lies just upstream of the Sunken Meadow type locality (Ward and Blackwelder, 1980). Two specimens from Lee Creek Mine, North Carolina (3 in Fig. 2B), were collected by A.L.A. Johnson (NC1, NC2; University of Derby, Geological Collections, 53346, 53347, respectively) and a further two specimens from this location were provided from the collections of the Florida Museum of Natural History (NC3, NC4; UF 261869, 261868, respectively). Lee Creek Mine is about 210 km south of the Virginia locations.

After washing the valves in tap-water to remove NH_4Cl (see below) and cleaning following the method adopted by Valentine et al. (2011), a hand-held drill with a 0.5 mm bit was used to extract samples of calcite powder from the outer layer, starting from a position near the dorsal margin (umbo) and continuing to the ventral margin, except in VA4 (sampled only to 56 mm of the total height of 62 mm) and NC4 (sampled only to 60.5 mm of the total height of 110 mm). Grooves 0.1-0.5 mm in depth and 5-60 mm in length were cut parallel to microgrowth-increment boundaries (Fig. 3B) in order to yield sufficient material for analysis and possible repeat analysis. Samples from VA1, VA2, NC1 and NC2 were extracted at height intervals averaging 1.3-1.5 mm ('fine' sampling) and analysed at the Stable Isotope Facility, British Geological Survey, Keyworth, using an Isoprime dual inlet mass spectrometer coupled to a Multiprep system. Powder samples were dissolved with concentrated phosphoric acid in borosilicate wheeton vials at 90 °C. Samples from VA3, VA4, NC3 and NC4 were extracted at height intervals averaging 2.3-2.5 mm ('coarse' sampling) and analysed at the Institute of Geological Sciences, University of Mainz, using a Thermo Finnigan MAT 253 continuous flow-isotope ratio mass spectrometer coupled to a Gasbench II. Powder samples were dissolved with concentrated phosphoric acid in helium-flushed borosilicate exetainers at 72 °C. Both laboratories calibrated their $\delta^{18}O$ and $\delta^{13}C$ data against NBS-19 and their own Carrara marble standard. Internal precision (1σ) for both laboratories are <0.05 for $\delta^{18}O$ and $\delta^{13}C$. Isotope values were calculated against the Vienna Pee Dee Belemnite (VPDB), Craig corrected, and reported in delta notation and given as parts per mil (McKinney et al., 1950).

The living descendant of *P. clintonius*, *P. magellanicus*, is a stenohaline marine form which has been shown to yield accurate information on seasonal temperatures from $\delta^{18}O$ of serial ontogenetic samples (Krantz et al., 1984; Tan et al., 1988; Chute et al., 2012). In conformity with the work on *P. magellanicus*, temperatures were calculated using the equation for calcite (1) of Epstein et al. (1953).

$$T = 16.5 - 4.3(\delta^{18}O_{calcite} - \delta^{18}O_{seawater}) + 0.14(\delta^{18}O_{calcite} - \delta^{18}O_{seawater})^2 \quad (1).$$

Values for $\delta^{18}O_{seawater}$ (measured against VSMOW) must be adjusted downward for correspondence with the VPDB scale used for $\delta^{18}O_{calcite}$. An adjustment of 0.22‰ was used by Chute et al. (2012) in recent work on *P. magellanicus*. However, the internationally agreed figure is 0.27‰ (Gonfiantini et al., 1995), and this has been used herein, in conformity with recent work on Pliocene *Mercenaria* from the Middle Atlantic Coastal Plain (Winkelstern et al., 2013). Various initial values of $\delta^{18}O_{seawater}$ were used in accordance with the differing estimates available (see below).

Before the cleaning and isotopic sampling procedure outlined above, the valves were coated with a sublimate of NH_4Cl (which enhances the visibility of surface details) and digitally photographed. Images were then imported into the measurement software Panopea© (2004, Peinl and Schöne) and both the position (shell height) of growth breaks and the size of microgrowth increments (Fig. 3B) determined along the axis of maximum growth, except in the umbonal region where (to varying heights) the record of growth had been effaced by abrasion. Microgrowth-increment profiles appear to offer insights into the degree of mixing of the water column and therefore assist interpretation of oxygen-isotope temperature data from bivalves: specifically, whether or not summer values correspond to surface temperatures or are an underestimate as a result of the development of thermal stratification (Johnson et al., 2009). While benthic temperature data is useful in its own right, informed estimation of the corresponding surface temperature is worthwhile for comparison with model outputs.

Results

Oxygen Isotopes.—Nearly all $\delta^{18}O$ values (Fig. 4) are positive and, as expected, show cyclic patterns in the case of every shell, assumed to reflect seasonal temperature variation. There is some fairly low amplitude 'noise' (e.g., from about 30-50 mm in VA3 and 20-40 mm in NC2; Fig. 4C, 4F), possibly reflecting the local occurrence of cement-lined micro-borings, as identified in Pliocene *Aequipecten opercularis* (Johnson et al., 2009). Only a few major anomalies are evident. VA3 (Fig. 4C) shows two extremely low values at the end of ontogeny which are very different from adjacent values and more than 1‰ lower than the minimum value of the preceding summer in this shell and all summer minima in other shells. VA4 (Fig. 4D) shows a single-point positive excursion of similar magnitude at a shell height of 16 mm. The anomalous $\delta^{18}O$ values in VA3 and VA4 are not accompanied by anomalous $\delta^{13}C$ values, but in NC2 (Fig. 4F) a single-point positive excursion of about 1‰ in $\delta^{18}O$ at 43.8 mm is matched by a slightly smaller one in $\delta^{13}C$. The values shown for NC2 are the result of resampling over a zone from 41.7-53.9 mm shell height after similar anomalies had been identified in the original profiles. Re-detection rules out contamination or instrumental malfunction. Diagenesis is an unlikely explanation (also for the anomalously high $\delta^{18}O$ value from VA4) because this typically causes a reduction in $\delta^{18}O$ (Tucker and Wright, 1990), but may account for the anomalously low values from VA3. The lack of concomitant shifts in $\delta^{13}C$ in VA3 could reflect the generally low amount of carbon (relative to oxygen) in porewaters (Tucker and Wright, 1990). Whatever their cause, it is appropriate to exclude from further analysis all the above values (signified by stars in Fig. 4). Abrupt excursions to lower $\delta^{18}O$ values in the late ontogeny (shell height >80 mm) of the large specimens NC1 and NC2 (Fig. 4E, 4F, respectively) might be thought to represent anomalies. However, comparison with the large set of oxygen-isotope profiles from modern *P. magellanicus* provided by Chute at al. (2012) suggests that they constitute summer records from individuals whose growth rate had declined, perhaps to zero for some intervals. The higher extreme values (i.e., lower peaks) than those in summer records from early ontogeny are an expected result of this (through time-averaging in sampling), as are the lower extreme values (i.e., shallower troughs) of intervening winter records in comparison with early ontogeny. Such ontogenetic reduction in the amplitude of seasonal oxygen-isotope cycles has been widely recognised in bivalves and makes it wise to concentrate on the first few years of growth in any attempt to document the full range of seasonal temperature fluctuation. For this reason, only information from the first and second winters and summers is incorporated in subsequent analysis.

Assignment of portions of the curves to summers or winters is generally uncontroversial but in certain cases either lateral truncation of the profile or 'noise' makes it impossible to determine the relevant seasonal extreme with accuracy, and in a few instances 'noise' is sufficient to raise doubts about the seasonal assignment. Thus, truncation at the ventral margin makes it impossible to say whether the lowest values recorded for Summer 2 in VA1 and VA2 (Fig. 4A, 4C, respectively) are representative of the extreme conditions in the summers concerned. Similarly, truncation at the dorsal end of the profile makes it impossible to say whether the highest value recorded for Winter 1 in NC3 (Fig. 4G) is representative of the extreme conditions in the winter concerned. With respect to 'noise', while the portion of the profile from NC2 (Fig. 4F) corresponding to Winter 1 is clear, identification of the exact position and value for the winter extreme is made problematic by high-frequency variability; it could indeed be argued that the two-point increase in $\delta^{18}O$

defining Summer 1 at the dorsal end of this profile together with the one-point decrease in $\delta^{18}O$ defining Winter 1 at the dorsal end of the profile from VA4 (Fig. 4D) are both representative of 'noise'. However, in at least the latter case the value for the winter extreme identified is similar to that of another winter extreme from the same shell. Uncertainties over seasonal recognition and the most appropriate value for seasonal extremes could be addressed by applying a smoothing function (e.g., Wang et al., 2015) but this would lead to an underestimation of seasonal range in cases where the extreme values measured are probably not representative of the extreme conditions experienced by the animal – i.e., where an extreme value derives from a sample close to a growth-break (blue triangles in Fig. 4). Such cases are not uncommon: the Summer 1 extremes in VA1, VA3, NC1 and NC4, the Summer 2 extremes in VA2, VA3, NC1 and NC3, and the Winter 2 extremes in VA3 and NC3 are from samples taken 0.5-1.5 mm from a major or moderate growth break, and there are other cases where seasonal extremes are from samples a little farther from growth breaks. Under these circumstances it seems best not to apply a smoothing function but to take the most extreme value measured for a given season as representative of the extreme conditions experienced. This has the additional benefit of enabling like-with-like comparison with data from previous isotopic studies of Yorktown Formation bivalves (Krantz, 1990; Goewert and Surge, 2008; Winkelstern et al., 2013). While the most extreme conditions experienced by the organism in a given winter or summer are probably not recorded in cases where the relevant measured extreme derives from a sample close to a growth break, it is unlikely that the measured value is seriously unrepresentative in years 1 and 2 because 'spring' and 'fall' growth breaks at this age (e.g., Fig. 4A-C, 4G) are rarely associated with any steepening of the $\delta^{18}O$ profile, implying that growth interruptions were generally brief.

Table 2 shows the extreme $\delta^{18}O$ values for the winter and summer periods identified. The 13 winter values range from +1.80‰ to +2.73‰. Of the six lowest values, four (+1.90‰, +1.91‰, +2.01‰, +2.15‰) are second values from shells showing another winter value of +2.25‰ or more. Of these four, one (Winter 1: NC3) is from the dorsal end of a profile and may well reflect truncation, while the other three (Winter 2: VA3, NC1, NC2) might be early manifestations of ontogenetic reduction in growth-rate (combined with temporary cessation of growth in VA3), leading to greater time-averaging in sampling. The mean of all the winter values (+2.24 ± 0.3‰; ± 1σ) may therefore be a less accurate guide to typical extreme winter temperature on the seafloor than the mean of the highest winter value from each shell (+2.42 ± 0.23‰). The highest winter values of the coarsely sampled shells are all less than those of the finely sampled shells, and the lowest and third lowest winter values recorded (+1.80‰ and +2.00‰) are from the most coarsely sampled shell (VA4), suggesting that higher values went undetected in coarse sampling. Notwithstanding the small sample size (4), the mean of the highest winter value from the finely sampled shells (+2.61 ± 0.10‰) is therefore probably the very best guide to typical extreme winter temperature on the sea floor. Whether comparing all the winter values from the Virginia and North Carolina shells (means: +2.29 ± 0.34‰, +2.20 ± 0.25‰, respectively), the highest winter values (means: +2.44 ± 0.30‰, +2.40 ± 0.12‰, respectively) or the highest winter values from finely sampled shells (means: +2.71 ± 0.02‰, +2.51 ± 0.04‰), the means from the North Carolina shells are consistently a fraction lower, suggesting a corresponding slight difference in winter benthic temperature.

The 14 summer values for $\delta^{18}O$ in Table 2 range from +0.98‰ to −0.13‰. The highest value (Summer 1: NC2) may well be representative of ‘noise’ rather than summer conditions. The high Summer 2 values from VA3 (+0.62‰) and NC2 (+0.67‰) are from samples respectively on or quite close (3.3 mm) to a growth break (Fig. 4C, 4F) and therefore may not give an accurate picture of the extreme summer conditions experienced. The same could apply to the high Summer 1 value (+0.69‰) from NC4, from a sample close (1.5 mm) to a growth break (Fig. 4H), and to those lower summer values which are from samples close or quite close to growth breaks (Summer 1: VA1, VA3, NC1; Summer 2: VA2, NC1, NC3) and/or at the end of laterally truncated profiles (Summer 2: VA1, VA2). However, while these values may not give a wholly accurate picture of extreme conditions in the summers concerned, it is unlikely that they are grossly misleading because they span a range (−0.13‰ to +0.69‰) which is similar to that spanned by values which are neither from samples close to growth breaks nor from the ends of truncated profiles (Summer 1: VA2, NC3; −0.03‰, +0.56‰, respectively). It is also unlikely that sampling strategy has influenced the results much because for North Carolina shells the mean of summer values from coarsely sampled shells (+0.52 ± 0.15‰) is only slightly higher than from finely sampled shells (+0.37 ± 0.23‰; +0.52 ± 0.33‰ if the questionable very high value from NC2 is included) and the same is true for Virginia shells (coarse: +0.24 ± 0.31‰; fine: +0.17 ± 0.23‰). The mean of all summer values is +0.31 ± 0.27‰ (+0.36 ± 0.31‰ if the questionable value from NC2 is included), and the mean of North Carolina values (+0.45 ± 0.21‰; +0.52 ± 0.27‰ if the questionable value from NC2 is included) is somewhat higher than the mean of Virginia values (+0.20 ± 0.27‰), suggesting a corresponding small difference in summer seafloor temperature.

Carbon Isotopes.—Most $\delta^{13}C$ values (Fig. 4) are positive but, with the exception of VA4 (Fig. 4D), all the profiles show more or less pronounced ontogenetic trends to lower values such that those from late ontogeny are often negative. Superimposed on the ontogenetic reduction, and displayed in every profile, is a cyclicity in $\delta^{13}C$ which parallels that in $\delta^{18}O$ except in the early ontogeny of VA1, VA2, VA4 and NC3 (Fig. 4A, 4B, 4D, 4G, respectively), where there are slight offsets or additional oscillations. The range of mean $\delta^{13}C$ values from North Carolina shells (NC1: +0.44 ± 0.66‰, NC2: +0.77 ± 0.54‰, NC3: +0.08 ± 0.49‰, NC4: +0.65 ± 0.58‰) is less than from Virginia shells (VA1: +1.01 ± 0.56‰, VA2: +1.09 ± 0.50‰, VA3: -0.04 ± 0.44‰, VA4: +0.50 ± 0.29‰) and the mean of the mean values from North Carolina shells (+0.49 ± 0.26‰) is also less than from Virginia shells (+0.64 ± 0.45‰).

Microgrowth Increments.—Trendlines (five-point averages) are included with the raw increment data in Figure 4. These show low amplitude, high frequency variation, particularly in the North Carolina shells, but in the Virginia shells VA1 and VA2 (Fig. 4A, 4C, respectively) a rather higher amplitude, lower frequency oscillation of about six cycles per annum is evident. Superimposed on this in VA2, also discernible in VA3 (Fig. 4C), and of high amplitude in VA4 (Fig. 4D) is an approximately annual cycle of increment size variation, which is also evident in the North Carolina shells NC1, NC3 and NC4 (Fig. 4E, 4G, 4H, respectively), and of high amplitude in NC3. The cycles are typically offset somewhat from those of $\delta^{18}O$ variation. NC2 (Fig. 4F) shows a supra-annual overall pattern of increment size variation.

Discussion

Shell Preservation.—While the anomalous $\delta^{18}O$ values discussed above are only explicable by diagenesis in the case of one shell, it might be argued that the covariation between $\delta^{18}O$ and $\delta^{13}C$ noted in every shell is evidence of pervasive alteration through interaction with meteoric waters, since these typically have low $\delta^{18}O$ and low $\delta^{13}C$ signatures. However, the wavelengths of $\delta^{18}O$ cycles from *P. clintonius* (typically 40-50 mm between Summer/Winter 1 and 2, decreasing thereafter; Fig. 4) are like those derived from modern *P. magellanicus* at a similar stage in ontogeny, and which undoubtedly relate to seasonal changes in ambient temperature (Krantz et al., 1984; Chute et al., 2012). The covariance of $\delta^{13}C$ with $\delta^{18}O$ must therefore relate to environmental changes which follow the seasonal temperature cycle rather than to diagenesis. Parallelism between $\delta^{18}O$ and $\delta^{13}C$ profiles has been noted in modern *P. magellanicus,* as has ontogenetic reduction in $\delta^{13}C$ like that seen in *P. clintonius* (Krantz et al., 1987, 1988), so there can be little doubt that the isotopic composition of samples from the latter is essentially original.

Paleohydrography.—$\delta^{18}O$ from bivalves provides a record of temperature in the benthic environment, but an estimate of SST is required for purposes of comparison with numerical models. In the shallow shelf (0-20 m), winter benthic temperature is usually almost identical to SST, and summer surface and seafloor temperatures are typically also about the same. In the mid-shelf (20-40 m), winter benthic temperature is usually only a little different from SST (typically a degree or two higher; e.g., Winkelstern et al., 2013) but, as indicated above, summer benthic temperature can be substantially lower than SST as a result of incomplete mixing down of warm, low density surface waters. The amount of difference is dependent on the intensity of summer heating, the depth of water and the degree of agitation by wave and current action.

Attempts have been made to characterise hydrographic setting in terms of the $\delta^{13}C$ of bivalves. Arthur et al. (1983) documented an antiphase relationship between $\delta^{13}C$ and $\delta^{18}O$ in a modern specimen of *Spisula solidissima* from 10 m depth and an in-phase pattern in two further specimens from 45 m depth, below the summer thermocline. They considered that this difference reflected removal of ^{12}C from surface waters by photosynthesis during the summer (giving high $\delta^{13}C$ alongside low $\delta^{18}O$ in the 10 m specimen) and supply of ^{12}C to deep benthic waters during the summer by oxidation of sedimented organic matter (giving low $\delta^{13}C$ alongside low $\delta^{18}O$ in the 45 m specimens). The in-phase pattern occurs in modern *Placopecten magellanicus* from 57 m (Krantz et al., 1987, 1988) so on the basis of its pervasive occurrence in the analysed *P. clintonius* one might infer a similar sub-thermocline setting for these shells. However, an antiphase pattern was recorded by Johnson et al. (2009, fig. 5) in a modern *Aequipecten opercularis* specimen from a sub-thermocline setting at 50 m (Gulf of Tunis, Mediterranean Sea), and in-phase $\delta^{13}C/\delta^{18}O$ variation was recorded by Krantz et al. (1987, figs. 4, 5) in modern *S. solidissima* from 14 m, a depth almost certainly above the summer thermocline. Possible explanations exist for these exceptions (e.g., in summer, insufficient dissolved oxygen and insufficient nutrients, respectively); however, the important point is that patterns of $\delta^{13}C$ variation in relation to $\delta^{18}O$ are not an infallible guide to hydrographic setting.

Independent evidence may be supplied by patterns of microgrowth increment size, especially in early ontogeny. The modern sub-thermocline specimen of *A. opercularis* referred to above was only one year old at death from the evidence of its $\delta^{18}O$ profile (Fig. 5B, red line). It shows some fairly low amplitude, high frequency fluctuation in increment size but superimposed on this is a high amplitude, approximately annual cycle of variation, out of phase with that of $\delta^{18}O$ (Fig. 5A, blue line). Similar increment patterns occur in other specimens from the same location (Fig. 5A, green and yellow lines), but their relationship with $\delta^{18}O$ is not known. However, high amplitude, annual-scale variation, out of phase with $\delta^{18}O$ in early ontogeny (to an age of about 1.5 years) but in phase later, occurs widely in fossil specimens of *A. opercularis* from the inferred stratified setting of the Pliocene Coralline Crag in eastern England (Johnson et al., 2009). By contrast, sub-fossil examples from the fairly shallow (mostly <40 m), strongly tidal and hence continuously well-mixed waters of the southern North Sea show no annual cycle or only a low amplitude one, usually in phase with $\delta^{18}O$ in early ontogeny (Fig. 5B). The majority of the investigated specimens of *P. clintonius* show an annual-scale variation in increment size. This is of high amplitude and out of phase with $\delta^{18}O$ in some from both Virginia and North Carolina (markedly so in the early ontogeny of VA2 and NC3; Fig. 4B, 4G, respectively). Thus, by analogy with increment patterns in *A. opercularis*, it can be deduced that seasonal stratification occurred in both areas. Microgrowth increment evidence therefore supports the argument based on the pattern of $\delta^{13}C$ variation in relation to $\delta^{18}O$. The inference of seasonal stratification is consistent with the 20-40 m depth estimate of Ward et al. (1991), since within this range on the modern shelf adjacent to the studied areas water temperature is notably cooler than at the surface in summer (see above).

Seafloor and Surface Paleotemperatures.—Isotope-derived temperatures are dependent on the value selected for $\delta^{18}O_{seawater}$ (Equation 1). Previous work on Pliocene scallops of the Middle Atlantic Coastal Plain (Krantz, 1990; Goewert and Surge, 2008) used a variety of mainly somewhat negative values based on the fact that global ice volume was generally lower than now. Modelling of regional variation, taking account of differences in evaporation and precipitation, yields positive $\delta^{18}O_{seawater}$ for the area in question: +0.7‰ for the early Pliocene and +1.1‰ for the mid-Pliocene (Williams et al., 2009). Our preferred estimates of temperature for the early Pliocene Sunken Meadow Member are based on calculations using the former value. However, we also supply the results of calculations using the latter value and, for reference, the most extreme negative value used in earlier work on material from the Sunken Meadow Member (−0.4‰; Krantz, 1990). Figure 6 shows profiles of isotope-derived temperature from each shell using the three values of $\delta^{18}O_{seawater}$, with the preferred data shown by a thicker line. Summer and winter extreme values are listed in Table 2. While salinity, and hence $\delta^{18}O_{seawater}$, might not have been constant through the year, any significant departures from normal would probably have been in the spring (from freshwater run-off; see above), so the calculated summer and winter extreme temperatures do not need any adjustment for short-term salinity variation.

Using the preferred value for $\delta^{18}O_{seawater}$ (+0.7‰), all the shells except VA4 (the most coarsely sampled) yield at least one winter minimum temperature below the lower limit for a warm temperate marine climate (10 °C), and the finely sampled shells all yield such a temperature even using the more positive value for $\delta^{18}O_{seawater}$ (+1.1‰). Using the preferred value for $\delta^{18}O_{seawater}$, VA3 shows a second winter minimum temperature below 10 °C but second minima from other shells (NC1, NC2, NC3) are slightly above 10 °C. For NC3 (Winter 1) truncation of the profile is a likely explanation but for NC1 and NC2 (both Winter 2) neither this, nor the existence of growth breaks, can be invoked. While a general slowing of growth may have been contributory in the case of NC1 and NC2, it seems more reasonable to conclude that the evidence from these shells of slightly warmer winters in some years reflects a slightly higher mean benthic winter temperature in North Carolina than Virginia.

Nevertheless, the temperature (9.3 ± 0.9 °C; ± 1σ) calculated from the mean of all winter (extreme) $\delta^{18}O$ values from North Carolina shells using the preferred value for $\delta^{18}O_{seawater}$ is, like the corresponding temperature calculated from Virginia values (9.0 ± 1.3 °C), still below the lower limit for a warm temperate marine climate. As argued above, the most accurate indication of typical extreme winter temperature on the seafloor may be supplied by the mean of the highest winter $\delta^{18}O$ value from each of the finely sampled shells. The temperatures so-derived for North Carolina (8.2 ± 0.2 °C) and Virginia (7.4 ± 0.1 °C) are even farther below the lower limit for a warm temperate marine climate. Given the estimated depth of the shells and the likelihood of a small insulating effect from the overlying water, winter surface temperature was probably a degree or two lower than seafloor temperature and thus very firmly within the mild/cool temperate range over the whole of the studied area.

Using the preferred value for $\delta^{18}O_{seawater}$ (+0.7‰), all 14 values for summer seafloor temperature are below the lower limit (22.5 ºC) of a warm temperate marine climate yet within the range (upper boundary: 21-23.5 °C; Dickie, 1958) tolerated by modern *P. magellanicus*. The lowest value (Summer 1 of NC2) may be an artefact of noise and a significant underestimate of the maximum temperature experienced. Some other values are probably also underestimates of benthic temperature due to the effects of growth breaks, coarse sampling and truncation of the data series, but are unlikely to be seriously misrepresentative. The temperatures calculated from the mean of all summer $\delta^{18}O$ values from Virginia shells, North Carolina shells and the two sets combined (excluding Summer 1 of NC2 in the last two cases) are, respectively, 17.5 ± 1.2 °C, 16.4 ± 0.9 °C and 17.0 ± 1.2 °C. While these figures are probably fairly accurate for the seafloor, they are likely to be significant underestimates of SST, by an amount in the order of 6 °C on the basis of the likely temperature/depth profile in summer. Adding this amount to all the individual summer values (excluding Summer 1 in NC2) yields temperatures above the lower limit for a warm temperate marine climate for five of the seven Virginia estimates and for three of the six North Carolina estimates. Adding 6 °C to the temperature calculated from the mean of all summer δ^{18}O values for each of Virginia and North Carolina (excluding Summer 1 in NC2) yields for the former a temperature above (23.5 °C) and for the latter a temperature fractionally below (22.4 °C) the warm temperate range. It is implausible that summer SST was actually lower in North Carolina than Virginia (farther north) so it may be that the shells from the former area occupied slightly deeper water and that a larger 'stratification factor' should have been applied.

While there is a solid basis for favouring those temperature estimates for the Sunken Meadow Member based on a value for $\delta^{18}O_{seawater}$ of +0.7‰, it cannot be denied that some uncertainty attaches to the absolute values obtained. Very little uncertainty attaches to estimates of seasonal range (i.e., relative temperature) because this parameter is only slightly affected by the value adopted for $\delta^{18}O_{seawater}$ (e.g., only a 1.2 ºC difference between the range estimates using $\delta^{18}O_{seawater}$ = −0.4‰ and +1.1‰ for the largest range in $\delta^{18}O_{calcite}$ observed, in VA1). Using a $\delta^{18}O_{seawater}$ value of +0.7‰ and calculating mean benthic seasonal range from the temperatures specified by the means of the most extreme winter and summer δ^{18}O values from each shell (for winter: 8.4 ± 1.1 °C in Virginia, 8.6 ± 0.4 °C in North Carolina, 8.5 ± 0.9 °C in both areas combined; for summer: 18.2 ± 0.6 °C in Virginia, 16.5 ± 1.1 °C in North Carolina, 17.3 ± 1.2 °C in both areas combined) yields figures of 9.8 ºC for Virginia, 7.9 ºC for North Carolina and 8.8 ºC for both areas combined. These figures, which can be regarded as maximum estimates, are not very different from those (8.5 ºC, 7.1 ºC, 7.8 ºC, respectively) calculated from the means of all winter and summer δ^{18}O values (excluding Summer 1 in NC2) from the Virginia, North Carolina and combined shells, which can be regarded as minimum estimates. From the fact that in the present-day MAB, shelf surface temperature is higher than at 30 m by about 6 ºC in summer and lower than at 30 m by about 2 ºC in winter (Winkelstern et al., 2013), it can be reasonably surmised that the seasonal range in SST when the analysed *P. clintonius* specimens were alive was about 8 ºC more than the benthic estimates derived from them, i.e., about 17 ºC in Virginia and 15.5 ºC in North Carolina.

The figures for absolute summer and winter seafloor temperatures derived above are very similar to those obtained by Krantz (1990) through an isotopic study of the scallop *Chesapecten jeffersonius* from the upper Sunken Meadow Member. However, the temperatures from *C. jeffersonius* become significantly higher than those from *P. clintonius* when calculated using the preferred value for $\delta^{18}O_{seawater}$ (see below). The estimate for summer temperature obtained by Hazel (1971, 1988) from analysis of Sunken Meadow ostracod assemblages is consistent with the benthic summer figures from *P. clintonius*, but his estimate for winter temperature (at least 12.5 °C) is markedly higher than the winter estimate from *P. clintonius.* Possibly ostracod assemblage composition is controlled more by summer temperature than winter and gives inaccurate indications of the latter.

Paleoclimate and Paleoceanography.—The estimates derived above for absolute surface temperature in summer and winter, and for surface seasonal range, are very comparable with present-day figures for offshore locations in the MAB but differ in all respects from present-day figures for offshore locations in the SAB, including a location immediately south of the latitude of Cape Hatteras (Table 1). The data derived from *P. clintonius* in Virginia thus show that surface conditions there when the animals were alive closely resembled those on the adjacent shelf now, but the data derived from *P. clintonius* in North Carolina (from a location at the same latitude as Cape Hatteras) reveal that much the same conditions also existed some 210 km farther south, where now shelf surface temperatures are higher, particularly in winter.

It is possible to invent explanations involving multiple causes for the circumstances indicated by the *P. clintonius* data. For instance, one could propose a cooler general climate combined with warm-current flow farther north, the former outweighing current influence in North Carolina and balancing it in Virginia. However, an explanation involving a single cause – cool-current flow farther south (Cronin and Dowsett, 1996) – is available and should be favoured on grounds of parsimony alone. In fact there are additional grounds for favouring this explanation. Firstly, despite the barrier created by Cape Hatteras at present, northern shelf waters probably penetrate into the SAB more than 50% of the time through wind-forcing (Pietrafesa et al., 1994). Secondly, the cool surface waters of the Slope Sea adjacent to the shelf north of Cape Hatteras have a higher nutrient content and primary productivity than the warm Gulf Stream farther offshore (Fig. 7); flow of these waters farther south to replace the Gulf Stream at the shelf edge (i.e., extension of the 'slope current'; Fig. 8), with intrusions onto the shelf as occur from the Gulf Stream now, provides a plausible explanation for the diverse fish fauna of the Sunken Meadow Member in North Carolina (Fierstine, 2001; Purdy et al., 2001) and the similarly diverse bird fauna, including many seabirds dependent on a rich marine food source (Olson and Rasmussen, 2001). Phosphate once thought to be primary and related to high productivity is now considered to be reworked from Miocene deposits (Riggs et al., 2000). However, the benthic foraminifera of the Sunken Meadow in North Carolina are still consistent with high nutrient supply (Snyder et al., 2001). The present-day shelf of southeastern South America from 38-32 °S (i.e., to a latitude lower than that of Cape Hatteras) provides an illustration of the scenario envisaged. Here, very high primary productivity (>500 $mgC/m^2/day$ and locally 770 $mgC/m^2/day$) supports major secondary production, including large populations of commercially exploited fish (Bisbal, 1995). Discharge of terrestrially derived nutrients from the Rio de La Plata and Patos-Mirim Lagoon system undoubtedly contributes to the productivity. However, it is also supported by supply of nutrients from the cold equatorward-flowing waters of the Falklands/Malvinas Current. These waters are at the 150-200 m isobath north of 35 °S (equivalent to the latitude of Cape Hatteras) but are locally returned to the surface on the shelf as far north as 23 °S through offshore Ekman transport caused by winds blowing south-west (alongshore) on the western side of the South Atlantic subtropical anticyclone. Given similar equatorward penetration of cold 'surface' waters on the western side of the North Atlantic during the Pliocene in the absence of a feature analogous to Cape Hatteras, winds blowing north-east on the western side of the subtropical anticyclone would likewise have led to nutrient enrichment of shelf waters. Greater southward spread of northern surface waters provides an explanation for the grand mean $\delta^{13}C$ values from *P. clintonius* in North Carolina and Virginia which, although different and lower for the southern area, are evidently much less so than the $\delta^{13}C$ values from modern outer shelf specimens of *Argopecten gibbus* in the SAB compared to outer shelf specimens of *P. clintonius* in the MAB (Krantz et al., 1988).

Reversal of the current pattern proposed above – i.e., extension of warm-current influence into the area of Virginia – is a simple and attractive explanation for the mid-Pliocene warming evinced by the higher members of the Yorktown Formation, and, as noted above, has been widely adopted. However, it is theoretically possible that cold-current flow continued over the area of deposition but its effect on water temperature was outweighed by general climatic warming. We evaluate this possibility in the next section through an investigation of isotopic temperatures from the upper part of the Sunken Meadow Member and from the higher members of the Yorktown Formation. We take particular note of the evidence of seasonal temperature range, since this is almost independent of the value used for $\delta^{18}O_{seawater}$ and is a parameter which can determine whether warming was caused by northward extension of warm-current influence. This circumstance would effectively bring the conditions of the modern SAB into the area of Virginia. Maps of average monthly benthic temperature supplied by Atkinson et al. (1983) enable reconstruction of the present benthic seasonal range over the SAB. Oceanward of the mid-shelf, the range is less than 10 °C, and landward it is more. However, only in very nearshore locations, in water shallower than 20 m, is the benthic seasonal range above 15 °C. Hence, any significant evidence of a benthic seasonal range greater than 15 °C from the higher Yorktown Formation would argue against northward extension of warm-current influence.

ISOTOPIC TEMPERATURES FROM THE HIGHER YORKTOWN FORMATION

Data from Other Scallop Genera

Placopecten does not occur in the Yorktown Formation above the basal section of the Sunken Meadow Member. However, the extinct scallop genera *Chesapecten* and *Carolinapecten* occur in all four members. They do not occur with brackish-water taxa and it can therefore be assumed that they were stenohaline marine, tolerating only brief reductions in salinity. Krantz (1990) obtained profiles of $\delta^{18}O$ and $\delta^{13}C$ from *Chesapecten jeffersonius* of the upper part of the Sunken Meadow Member, *C. madisonius* of the Rushmere and Moore House members and *Carolinapecten eboreus* of the Morgarts Beach and Moore House members. Further profiles from *C. madisonius* of the Moore House Member were obtained by Goewert and Surge (2008). These authors listed the extreme $\delta^{18}O$ values from each specimen. We have used their data, in conjunction with appropriate figures for $\delta^{18}O_{seawater}$, to calculate maximum estimates of benthic seasonal temperature range for comparison with the test value identified above. We have employed the equation (1) and set of $\delta^{18}O_{seawater}$ values (−0.4‰, +0.7‰, +1.1‰) used for *Placopecten clintonius* to derive figures for winter and summer benthic temperature for each specimen (Table 2), although arguably for the mid-Pliocene material the low value for $\delta^{18}O_{seawater}$ should have been set at -0.6‰ for Rushmere and Morgarts Beach individuals and -0.5‰ for Moore House individuals in recognition of the use of these figures by Krantz (1990) and Goewert and Surge (2008). Following Williams et al. (2009), our preferred value for these units is +1.1‰. Although our focus is the seasonal range in benthic temperature, we comment below on the winter and summer benthic temperatures from which the ranges are derived, and also attempt to determine surface temperatures using the evidence of depth supplied by patterns of variation in $\delta^{13}C$ relative to $\delta^{18}O$, together with other indicators. The sampling of *Chesapecten* and *Carolinapecten* shells by Krantz (1990) and Goewert and Surge (2008) was in the outer shell layer but at a higher spatial resolution (1 mm or less) than our sampling of *Placopecten* shells; hence, although few in number for some horizons, the estimates of seasonal benthic (and surface) temperature for the higher Yorktown Formation are certainly at least as reliable as those for the basal part of the Sunken Meadow Member.

Upper Sunken Meadow Member.—Krantz (1990) obtained isotope profiles from three specimens of *Chesapecten jeffersonius* from two locations on the James River, Virginia, close to the collection locations of the specimens of *Placopecten clintonius* from Virginia discussed above. The two specimens (SM-CJ1, SM-CJ2) from the first location, Sunken Meadow Creek (the Sunken Meadow type locality; 4 in Fig. 2B), were collected about 1 m above the base of the member and the single specimen (KING-CJ) from the second, Kingsmill (5 in Fig. 2B), immediately below the contact with the overlying Rushmere Member, 2-3 m above the base of the Sunken Meadow Member (Ward and Blackwelder, 1980, fig. 22). The interpretation of depth and water-column structure derived for the basal Sunken Meadow Member from the pattern of variation in $\delta^{13}C$ in relation to $\delta^{18}O$ within *P. clintonius* shells can also be applied to higher horizons from the evidence of strong parallelism in the profiles derived from *C. jeffersonius* specimens KING-CJ and SM-CJ1 by Krantz (1990, figs. 3b, c). Krantz ascribed the lack of parallelism in SM-CJ2 (1990, fig. 3a) to diagenesis. All three specimens contain at least one winter record of $\delta^{18}O$. The temperatures derived from the most extreme values represented in each shell for the preferred value of $\delta^{18}O_{seawater}$ (+0.7‰; Table 2) give a winter mean of 11.4 ± 0.7 °C (± 1σ). The equivalent for summer (based on only two extreme values because SM-CJ1 lacks a full summer record) is 23.0 ± 1.0 °C. The winter mean may be a slight overestimate and the summer mean a slight underestimate because the winter value from SM-CJ2 and the summer value from KING-CJ are each from samples close to major growth breaks. Because these potential errors have opposite effects on the estimate of annual range, the figure (11.6 °C) derived for this from the winter and summer means is probably quite accurate. The figures for winter and summer mean temperature, median temperature (17.2 °C; taken as an approximation of annual mean temperature) and for annual range are all notably higher than the corresponding figures for *P. clintonius* in Virginia (respectively, 8.4 °C, 18.2 °C, 13.3 °C, 9.8 °C), and also North Carolina (8.6 °C, 16.5 °C, 12.6 °C, 7.9 °C). While the higher summer temperature and annual range might reflect closer sample spacing, this would not have led to recognition of a higher winter temperature. Thus, although one must acknowledge the rather small sample size, the data from *C. jeffersonius* in the upper part of the Sunken Meadow Member does seem to suggest warmer conditions on the seafloor. It also suggests warmer surface temperatures: about 9.5 °C in winter and 29 °C in summer from the arguments applied to *P. clintonius*, giving a median temperature (19.3 °C) which is higher than is specified by modern winter and summer surface temperatures at any location in the MAB (Table 1). While the increase in benthic seasonal range from that indicated by the earlier *P. clintonius* shells is not sufficient to meet the test criterion of a value of 15 °C, it is consistent with the idea that higher seawater temperatures were brought about by a warming of general climate rather than increased influence of warm currents from the south.

Rushmere and Morgarts Beach Members.—It is unfortunate that published isotope profiles exist for only single scallop specimens from each of these members, which were deposited as the result of a major transgression, leading to the highest stand of sea level in the late Neogene (Krantz, 1991) and the establishment of marine conditions across the Atlantic Coastal Plain from Maryland to Florida. Figure 2B shows the coastline according to Ward et al. (1991, fig. 16-4B), which is similar to the reconstruction of Rowley et al. (2013). Its only slightly curvilinear form would have presented no obstruction to northward- or southward-flowing currents, but shoals identified by Ward et al. (1991) in the area of the Mid-Carolina Platform High would have restricted the passage of warm waters into northeast North Carolina and southeast Virginia. Winkelstern et al. (2013) placed the Morgarts Beach Member within the MPWP but considered that the lower part of the underlying Rushmere Member might have been deposited before it, a deduction consistent with the 4.0-3.0 Ma age range for both members indicated by microfossil biostratigraphy (Krantz, 1991; Fig. 1). Whatever the time of onset of the transgression, assemblage evidence from ostracods (Hazel, 1971, 1988; Cronin, 1991), mollusks (Ward et al., 1991) and foraminifers (Dowsett and Wiggs, 1992) points unequivocally to warmer ('warm temperate') conditions in southeast Virginia and northeast North Carolina during deposition of the Rushmere and Morgarts Beach members than during Sunken Meadow deposition.

Rushmere sediments are closely similar to those of the Sunken Meadow Member but sometimes with a larger admixture of clay (Ward and Blackwelder, 1980), suggesting quieter, somewhat deeper conditions, although still within the mid-shelf depth range (Krantz, 1991). The analysed specimen (BB-CM; Krantz, 1990, fig. 4a) is from near Fort Boykin, Burwell Bay, on the James River, Virginia (6 in Fig. 2B). The $\delta^{13}C$ and $\delta^{18}O$ profiles show only short parallel trends and so provide no positive support for a seasonally stratified and hence fairly deep setting. Equally, however, they provide no clear support for a shallow setting. The winter and summer seafloor temperatures calculated from extreme $\delta^{18}O$ values using the preferred (mid-Pliocene) value for $\delta^{18}O_{seawater}$ (+1.1‰; Table 2) are, respectively, 13.3 °C and 29.3 °C (median: 21.3 °C). These seasonal temperatures are considerably higher than the equivalents from any Sunken Meadow specimen (*P. clintonius* or *C. jeffersonius*) and entirely in accordance with previous deductions of very warm conditions during deposition of the Rushmere Member. Even when calculated using a value for $\delta^{18}O_{seawater}$ of +0.7‰, summer temperature (27.3 °C) is still higher than from all the Sunken Meadow specimens using the same value for $\delta^{18}O_{seawater}$; however, winter temperature (11.7 °C) is only higher than from the eight *P. clintonius* specimens and one (SM-CJ1) of the three *C. jeffersonius* specimens. Assuming the depth of the specimen was similar to that of Sunken Meadow specimens, and using the argument applied to them, yields for the Rushmere Member a surface winter temperature of about 11.5 °C and surface summer temperature of about 35.5 °C (this is not implausible given a modern record of 33 °C at a location 330 km north of Cape Hatteras; NDBC, undated, station OCIM2). The benthic seasonal range (16 °C, using +1.1‰ for $\delta^{18}O_{seawater}$) supplied by the Rushmere Member specimen exceeds the 15 °C test criterion so, while one cannot draw firm conclusions from a single shell, the suggestion is of a warming of general climate rather than an increase in warm-current influence.

The analysed specimen of *Carolinapecten eboreus* from the Morgarts Berach Member (LTRUN-EB; Krantz, 1990, fig. 4b) is from Lieutenant Run, Petersburg, Virginia (7 in Fig. 2B). The setting is a barred embayment, probably quite shallow, notwithstanding the clayey silt sediment. The profile appears to record a winter minimum temperature – 12.5 °C using the preferred value for $\delta^{18}O_{seawater}$ (+1.1‰) – which is again higher than any Sunken Meadow specimen (using $\delta^{18}O_{seawater}$ = +0.7‰). Given the probable shallow setting, a similar surface temperature can be inferred. The profile lacks a full summer record but the highest temperature recorded (18.8 °C) indicates a median benthic temperature of at least 15.7 °C: higher than from the basal Sunken Meadow Member.

Moore House Member.—This unit only occurs in a small area of southeast Virginia, with age-equivalent strata known locally in the subsurface of northeast North Carolina (Ward et al., 1991). It represents a further but less extensive transgression after a fall in sea level following the Rushmere/Morgarts Beach transgression. Figure 2B shows the coastline according to Ward et al. (1991, fig. 16-4C). The possible recurvature to the east indicated in the area of modern Cape Hatteras is a reflection of the absence of any unequivocally age-equivalent marine strata farther south (Krantz, 1991). While this may be a result of erosion (Ward and Blackwelder, 1980) it is quite possible that the Coastal Plain was emergent south of the area of Moore House occurrence, creating a barrier to the entry of warm waters. Sr isotope dating gives ages (2.6-2.5 Ma) well after the MPWP (Winkelstern et al., 2013) but microfossil biostratigraphy gives an older date (no younger than 2.8 Ma; Krantz, 1990), and correlation with the eustatic sea level curve suggests an age of 3.1-3.0 Ma, i.e., within the MPWP (Krantz, 1991; Fig. 1). On the basis of the common presence of subtropical molluscan taxa at downdip (more easterly) locations, Ward et al. (1991) considered that the marine climate was even warmer than during the Rushmere/Morgarts Beach transgression. The three specimens of *Chesapecten madisonious* investigated by Goewert and Surge (2008) were collected from a sequence of glauconitic sands and shell hash at Riddick Pit, Chuckatuck, Virginia (8 in Fig. 2B), about 50 km from the contemporary shoreline. The setting was probably an offshore bar, but one permanently submerged on the evidence of the fully marine associated fauna (including common *Glycymeris*, as well as *Marvacrassatella* and *Dinocardium*; A.L.A. Johnson, personal observations, 2007). While the sedimentary evidence suggests quite shallow, and hence well-mixed, waters, the strong covariation of $\delta^{13}C$ and $\delta^{18}O$ in two of the *C. madisonius* specimens (CMAD-4, CMAD-5;

Goewert and Surge, 2008, fig. 3b, 3c) seems to imply a deeper, seasonally stratified situation. It is possible, however, that the reductions in $\delta^{13}C$ alongside the spring-summer reductions in $\delta^{18}O$ in these specimens are an ontogenetic rather than environmental effect. The two specimens of *C. madisonius* and one of *Carolinapecten eboreus* investigated by Krantz (1990) were collected from a sequence of shelly sands at Yadkin Pit, near Deep Creek, Virginia (9 in Fig. 2B), some 25 km southeast of Chuckatuck and farther from the contemporary shoreline. The $\delta^{13}C$ profiles of two specimens (YAD-EB1, YAD-CM1; Krantz, 1990, fig. 5b, 5c) are rather 'flat' and hence uninformative about hydrographic setting. That of the third (YAD-CM2; Krantz, 1990, fig. 5a) shows strong covariation of $\delta^{13}C$ and $\delta^{18}O$, which might reflect seasonal stratification but could, again, be an ontogenetic effect. The mean winter and summer seafloor temperatures calculated from all the extreme $\delta^{18}O$ values using the preferred (mid-Pliocene) value for $\delta^{18}O_{seawater}$ (+1.1‰; Table 2) are, respectively, 12.0 ± 2.2 °C and 27.6 ± 1.5 °C, similar to the corresponding temperatures obtained from the single Rushmere and Morgarts Beach specimens, and the median temperature (19.8 °C) is also similar to that from the Rushmere specimen. The winter mean is probably something of an overestimate because the raw values from all six specimens are from samples close to major growth breaks. The seasonal range calculated from the winter and summer means (15.6 °C) is therefore likely to be somewhat underestimated. It is in any case higher than the test criterion, suggesting that the warm winter and summer benthic temperatures are the consequence of a warmer general climate rather than greater influence of warm currents. Given the conflicting evidence of depth from sedimentology and $\delta^{13}C$ variation in relation to $\delta^{18}O$, seasonal surface temperatures are difficult to infer. However, it is important to point out that even if the water was shallow (<20 m) the interpretation of the benthic data in terms of general climate is valid because the source locations for the analysed specimens are at least 50 km from the contemporary shoreline, and hence not comparable with those present-day shallow settings close to the coast in the SAB where benthic seasonal temperature range exceeds 15 °C (see above). Local shallow areas distant from the SAB coastline have a much lower seasonal range in surface temperature (e.g., 11 °C at Frying Pan Shoals, about 50 km offshore from Cape Fear; Table 2, stations FPSN7, 41013) and the benthic range is almost certainly a little lower still.

When calculated separately for each of the Moore House locations, the seasonal mean temperatures for Yadkin Pit are higher than for Riddick Pit – only slightly in summer (respectively, 28.6 ± 1.3 °C and 26.5 ± 0.8 °C) but quite markedly in winter (respectively, 14.1 ± 0.6 °C and 9.8 ± 0.8 °C). Given the more offshore position of the former site, it is tempting to think of this as a reflection of Gulf Stream influence. However, the large seasonal range (14.5 °C) argues against this. It therefore seems likely that the specimens from the two locations are not exactly contemporaneous and reflect fluctuations in general climate (albeit smaller than the overall change over the duration of the Yorktown Formation) during deposition of the Moore House Member.

Overview.—The above isotopic data from scallops confirms the mid-Pliocene warming of marine climate on the eastern seaboard of the US that has been adduced from other evidence, and shows that warming also occurred in the early Pliocene. Evidence of a benthic seasonal range in excess of 15 °C from the mid-Pliocene Rushmere and Moore House members indicates that warming was brought about by a change in general climate rather than an increase in warm-current influence. Indeed, shallow or emergent areas south of the depositional basin make it hard to countenance greater warm-current influence. In so far as the benthic seasonal ranges determined for the mid-Pliocene of Virginia are as great as the present surface range in the outer MAB (Table 1, station 44014), and only a little short of the surface range nearer shore, it may be concluded that the cold-current influence on this area now, which has also been inferred for the early Pliocene (*Placopecten* data), existed additionally in the mid-Pliocene. While the envisaged current pattern is like that inferred for the early Pliocene (Fig. 8), the shelf current would have been displaced eastwards during deposition of the Moore House Member, when sea level was lower (Fig. 2).

Data from *Mercenaria*

Winkelstern et al. (2013) obtained $\delta^{18}O$ and $\delta^{13}C$ profiles from the hinge plates of six specimens of the infaunal bivalve *Mercenaria*, collected from the Rushmere Member at a location on the James River, Virginia, approximately the same as that of the *Chesapecten madisonus* specimen from the Rushmere Member discussed above (6 in Fig. 2B). Three of the specimens (FB5, FB16, FB25; Winkelstern et al., 2013, fig. 6H, 6J, 6L) show a clear antiphase relationship between $\delta^{18}O$ and $\delta^{13}C$, suggesting a shallow-water setting, unlike that inferred for the Rushmere Member in general. Winkelstern et al. (2013) did not tabulate values of maximum and minimum $\delta^{18}O$ from each shell but listed winter (15.7 °C, 15.3 °C, 20.2 °C, 16.5 °C, 18.6 °C, 15.5 °C) and summer (24.7 °C, 25.8 °C, 27.7 °C, 26.0 °C, 29.2 °C, 23.7 °C) temperatures calculated from these data using an equation appropriate for the aragonite mineralogy of *Mercenaria* (Grossman and Ku, 1986) and a value for $\delta^{18}O_{seawater}$ of +1.1‰ (reduced by 0.27‰ for temperature calculation; see above). The temperatures can thus be directly compared with those calculated from mid-Pliocene scallop data using the same value for $\delta^{18}O_{seawater}$. The summer temperature (29.3 °C) obtained from the single investigated Rushmere scallop is higher than any obtained from *Mercenaria* and the winter temperature (13.3 °C) is lower. The winter temperature (12.5 °C) from the single investigated Morgarts Beach scallop is similarly lower, as are all the winter temperatures (8.9–14.5 °C) from the six Moore House scallops. The summer temperatures from the Moore House scallops (25.3–29.8 °C) overlap the range of *Mercenaria* data but the mean of the former (27.6 ± 1.5 °C; ± 1σ) is a little higher than the mean from the *Mercenaria* data (26.2 ± 1.8 °C), while the winter mean from the Moore House scallops (12.0 ± 2.2 °C) is substantially lower than from *Mercenaria* (17.0 ± 1.8 °C). Although the *Mercenaria* data agrees with that from mid-Pliocene scallops in that it demonstrates higher winter and summer benthic temperatures (and a higher median temperature: 21.6 °C) than in the early Pliocene, it differs radically by evincing a low seasonal range (9.2 °C). Irrespective of the fact that a higher surface range can be inferred if the shells derive from a stratified setting (as seems probable for the Rushmere Member in general but is contradicted by the $\delta^{18}O$/$\delta^{13}C$ evidence from these shells), the figure for benthic range (which is well short of the test criterion of 15 °C) argues against the model of general climatic warming and continuing cold-current influence adduced above. Indeed, Winkelstern et al. (2013) argue perfectly logically from their data that the mid-Pliocene warming of marine climate on the eastern US seaboard was a consequence of increased warm-current influence.

The low benthic seasonal range identified by Winkelstern et al. (2013) results largely from the relatively high winter temperatures determined from their $\delta^{18}O$ profiles. It is possible that these are an artefact of growth in a setting influenced by winter freshwater influxes; this would have reduced shell $\delta^{18}O$, resulting in an overestimate of winter temperature. While the Rushmere Member was undoubtedly deposited in an offshore, fully marine setting, it is conceivable that some faunal elements were transported from a nearer shore, freshwater-influenced environment. Certainly *Mercenaria* is able to tolerate reduced salinity (Elliot et al., 2003) and the $\delta^{18}O/\delta^{13}C$ evidence from the Rushmere specimens is consistent with growth in a nearshore setting. Many (but not all) individuals from the shell bed concerned are disarticulated and broken (I.Z. Winkelstern, personal communication, 2016), suggesting a measure of transport. However, it is questionable whether the shells could have been moved offshore the many tens of kilometers implied from near the contemporary shoreline (Fig. 2), especially as storm currents (the only plausible agent) have a weaker offshore than alongshore component (Swift et al., 1986).

An alternative explanation for the winter temperatures from Rushmere *Mercenaria* is that growth slowed during that season and sampling was insufficiently close to identify winter extremes of shell $\delta^{18}O$. In modern *Mercenaria* the optimum temperature for growth is about 20 °C, and in most populations growth is significantly reduced at temperatures below the minimum (15.3 °C) recorded by Rushmere forms, although it may continue to 9 °C (Ansell, 1968). The reduction in growth with declining temperature seems a likely explanation for the relatively low winter values of shell $\delta^{18}O$, compared to those predicted, in modern forms from Cedar Key, Florida, where water temperature does not fall below 10 °C (Elliot et al., 2003, figs. 2, 7). The possibility of a combined growth rate/sampling effect on the winter temperatures supplied by Rushmere forms is borne out by a comparison of the data of Winkelstern et al. (2013) from *Mercenaria* of the early Pleistocene Chowan River Formation with that of Krantz (1990) from *Carolinapecten eboreus* of the same unit. The former was obtained, like the data from Rushmere *Mercenaria*, by sampling the hinge plate, while the latter derives from sampling the full shell along the axis of maximum growth, providing better temporal resolution even at the greater sample spacing (~1 mm compared to ~0.1 mm). The mean of maximum summer temperatures from six Chowan River *Mercenaria* (Winkelstern et al., 2013, p. 655; calculated with $\delta^{18}O_{seawater} = 0.0‰$) is 22.0 ± 2.2 °C, similar to the mean of 20.8 ± 1.9 °C derived from eight Chowan River *C. eboreus* calculated using data for minimum shell $\delta^{18}O$ (Krantz et al., 1990, table 4), the same value for $\delta^{18}O_{seawater}$, and Equation 1. By contrast, the equivalent winter temperature from Chowan River *Mercenaria* is 10.5 ± 0.5 °C while that from *C. eboreus* is 7.5 ± 1.6 °C, and it should be noted that the latter figure is probably something of an overestimate because four of the shells used have major growth breaks close to the position of the samples yielding the highest value of $\delta^{18}O$ (Krantz, 1990, figs. 6, 7). Notwithstanding this confirmation of differences between estimates of winter temperature from the full shell of scallops and the hinge plate of *Mercenaria*, it should be noted that at least some of the *Mercenaria* profiles (both from the Rushmere Member and the Chowan River Formation) have well-resolved ('sinusoidal') winter portions, and a profile obtained from the full shell of one *Mercenaria* specimen shows winter (and summer) extreme values very similar to those in an equivalent profile from the hinge plate (Winkelstern et al., 2013, fig. 5). Moreover, while some of the winter sectors of Chowan River *Mercenaria* profiles correspond to dark shell material (usually associated with slow growth), others correspond to light shell material, and this is generally the case for Rushmere shells, arguing for relatively fast growth in winter.

While certain possibilities require further investigation, there is at present nothing to invalidate the temperature data from Rushmere *Mercenaria*. One must therefore accept that these lived at a time when winter temperature was higher than that experienced by the investigated mid-Pliocene scallops, and that this resulted from warm-current influence, probably in addition to the effect of general climatic warming.

SEASONAL TEMPERATURE RANGE FOR THE HIGHER YORKTOWN FORMATION FROM BRYOZOAN ZOOID SIZE

Knowles at al. (2009) measured variation in the size of zooids through the growth (astogeny) of colonies of various cheilostome bryozoan species to obtain estimates of mean annual range in temperature (MART) for the Rushmere (6.39 ± 0.69 °C; ± 1σ), Morgarts Beach (6.54 ± 1.24 °C) and Moore House (6.39 ± 0.96 °C) members in Virginia. These estimates are very much lower than the figures for seasonal temperature range obtained by isotopic analysis of scallops and also lower than the figure obtained by isotopic analysis of *Mercenaria*. The bryozoan MART technique is generally robust (Okamura et al., 2011), and the colonies analysed by Knowles et al. (2009) did not apparently contain any growth breaks which would have led to underestimates of seasonal range, so on the face of it one must take these results as a further indication of the periodic influence of warm currents in the mid-Pliocene of Virginia. However, the maps of average monthly benthic temperature supplied by Atkinson et al. (1983) show that in the shelf region at present influenced by warm currents on the eastern seaboard of the USA only a small area of the outer shelf within the southern SAB experiences a seasonal range in benthic temperature as low as that indicated by bryozoans for the mid-Pliocene in Virginia. While there are some uncertainties about the depth and distance from shore at which the sequences concerned accumulated, it is unlikely that they represent outer shelf settings, so it is questionable whether the figures supplied by the bryozoan MART technique are accurate in this case. One way of testing would be to obtain oxygen isotope profiles across the colonies, although extracting samples which are not time-averaged would be difficult in species exhibiting secondary skeletal wall thickening (Knowles et al., 2010). Until confirmatory isotopic evidence is provided we feel that the figures for seasonal temperature range from zooid-size variation in Yorktown Formation bryozoans cannot be accepted.

CONCLUSIONS AND FURTHER WORK

New oxygen isotope data presented herein from the scallop *Placopecten clintonius* of the early Pliocene Sunken Meadow Member (Yorktown Formation) confirms other indications of a relatively cool marine climate at this time on the US eastern seaboard in the area of Virginia and North Carolina. The uniformity of cool conditions over this region is most simply interpreted as a consequence of greater southward penetration of cold currents in the absence of a feature analogous to Cape Hatteras. This interpretation is supported by evidence of high primary productivity, which could have been promoted by supply of nutrient-rich cold water from the north. Reanalysis of existing oxygen isotope data from other scallop taxa confirms previous reports of a warming of marine climate in the mid-Pliocene but indicates that this was not due to the impingement of warm currents but to a warming of general climate, with cold currents still penetrating the area. Existing oxygen isotope data from examples of the infaunal bivalve *Mercenaria* from the mid-Pliocene Rushmere member of the Yorktown Formation suggests otherwise. The warm-current influence indicated by the high median temperature and low seasonal range derived from these shells probably reflects a temporary increase in the vigour of the Gulf Stream rather than a switch from an ‘off’ to an ‘on’ state. The former state may have existed in marine isotope stage M2 (~3.3 Ma; Lisiecki ad Raymo, 2005), when there is evidence of extensive glaciation in the northern hemisphere, but during the subsequent MPWP the Gulf Stream was probably continuously ‘on’ (De Schepper et al., 2013). While local topographic change might have allowed Gulf Stream water to temporarily displace water derived from the north on the shelf of the eastern USA, the evidence of Pliocene fluctuations in warm-current influence derived from isotopic and other indications of temperature on the other side of the North Atlantic (Fig. 1; Johnson et al., 2009; Knowles et al., 2009; Williams et al., 2009; Valentine et al., 2011) indicates the likelihood of regional oceanographic change, e.g., in the strength and position of the North Atlantic gyre, of which the Gulf Stream is part.

Clearly, there is a need for data from co-occurring, autochthonous bivalves and bryozoans from the mid-Pliocene of the eastern USA to determine whether the divergent estimates of seasonal temperature range from isotopic analysis of the former and zooid-size analysis of the latter are a reflection of inaccuracies in one or other approach, or representative of real differences resulting from temporal variation in warm-current influence. If congruent data are obtained, it will be useful to chart stratigraphic changes in seasonality in greater detail and to investigate whether there is a matching pattern on the other side of the Atlantic, and hence a common cause. Insights into the controls on marine climate in the North Atlantic region are likely to be derived from investigations into primary productivity, since this is dependent on nutrient supply, which is in turn influenced by current patterns. Indications of primary productivity can be obtained from sediment composition, and from the abundance, diversity and type of the fossil biota (see above). However, they can also be obtained from shell profiles of indicative trace elements (Krantz et al., 1988; Haveles et al., 2010; Thébault and Chauvaud, 2013) and from evidence of shell growth rates (Kirby, 2000, 2001; Johnson et al., 2007; Haveles et al., 2010). In the latter respect it is of interest that Goewert and Surge (2008) recorded extensional growth rates up to nearly 70 mm per annum in their specimens of *Chesapecten madisonius* from the Moore House Member. This is similar to the maximum annual growth rate in wild modern scallops (Yamamoto, 1953; Bricelj and Shumway, 1991, fig. 7) and certainly implies an abundant food source, probably phytoplankton. Other bivalve taxa (e.g., *Carolinapecten, Mercenaria*, *Glycymeris*, *Marvacrassatella*, *Dinocardium*) reach a large size in the Yorktown Formation and may likewise have grown rapidly. Sclerochronological techniques provide a means of testing this.

ACKNOWLEDGMENTS

We are deeply grateful to Lauck (‘Buck’) Ward for guiding ALAJ in the field, supplying material from the Virginia Museum of Natural History (Martinsville) and providing information on the precise stratigraphic provenance of specimens. Roger Portell (Florida Museum of Natural History, Gainesville) and Virginie Héros (Muséum National d’Histoire Naturelle, Paris) also kindly supplied material from collections in their care. Isotopic analysis at Mainz was conducted with the assistance of Michael Maus. Isotopic analysis at Keyworth was conducted in the context of agreed collaborative work between the NERC Isotope Geosciences Laboratory and British Geological Survey (BGS) in 2010-2011. AV was in receipt of a PhD studentship award through the BGS University Funding Initiative (BUFI S157). ALAJ received support for travel from the Research-Inspired Curriculum Fund, University of Derby. We thank Harry Dowsett (US Geological Survey), Joel Hudley

(University of North Carolina), Stan Riggs (East Carolina University), Mark Williams (University of Leicester) and numerous students from the University of Derby for contributing useful insights during the course of the work, and the reviewers (especially David Krantz) for constructive comments on the manuscript.

REFERENCES

ANSELL, A.D., 1968, The rate of growth of the hard clam *Mercenaria mercenaria* (L) throughout the geographical range: Journal du Conseil permanent international pour l'Exploration de la Mer, v. 31, p. 364–409.

ARTHUR, M.A., WILLIAMS, D.F., and JONES, D.S., 1983, Seasonal temperature-salinity changes and thermocline development in the mid-Atlantic Bight as recorded by the isotopic composition of bivalves: Geology, v. 11, p. 655-659, doi: 10.1130/0091-7613(1983)11<655:STCATD>2.0.CO;2.

ATKINSON, L.P., 1977, Modes of Gulf Stream intrusions into the South Atlantic Bight shelf waters: Geophysical Research Letters, v. 4, p. 583-586, doi: 10.1029/GL004i012p00583.

ATKINSON, L.P., LEE, T.N., BLANTON, J.O., and CHANDLER, W.S., 1983, Climatology of southeastern United States shelf waters: Journal of Geophysical Research – Oceans and Atmospheres, v. 88, p. 4705-4718, doi:10.1029/JC088iC08p04705.

BARRERA, E., TEVESJ, M.J.S., and CARTER, J.G., 1990, Variations in oxygen and carbon isotopic compositions and microstructure of the shell of *Adamussium colbecki* (Bivalvia): PALAIOS, v. 5, p. 149-159, doi: 10.2307/3514811.

Belda, M., Holtanová, E., Halenka, T., and Kalvová, J., 2014, Climate classification revisited: from K□ppen to Trewartha: Climate Research, v. 59, p. 1-13, doi: 10.3354/cr01204.

Berggren, W.A., 1973, Pliocene time scale: Calibration of planktonic foraminiferal and calcareous nannoplankton zones: Nature, v. 243, p. 391-397, doi: 10.1038/243391a0.

Berggren, W.A., Kent, D.V., and Van Couvering, J.A., 1985, Neogene geochronology and chronostratigraphy, *in* Snelling, N.J., ed., The Chronology of the Geological Record. Memoir 10: The Geological Society, London, UK, p. 211-260.

Bisbal, G.A., 1995, The Southeast South American shelf large marine ecosystem: Evolution and components: Marine Policy, v. 19, p. 21-38, doi: 10.1016/0308-597X(95)92570-W.

Blow, W.H., 1969, Late middle Eocene to Recent planktonic foraminiferal biostratigraphy, *in* Bronniman, P., and Renz, H.H., eds., Proceedings of the First International Conference on Planktonic Microfossils: E.J. Brill, Leiden, The Netherlands, p. 199-422.

Böhm, E., Hopkins, T.S., Pietrafesa, L.J., and Churchill, J.H., 2006, Continental slope sea level and flow variability induced by lateral movements of the Gulf Stream in the Middle Atlantic Bight: Progress in Oceanography, v. 70, p. 196-212, doi: 10.1016/j.pocean.2006.07.005.

Bricelj, V.M., and Shumway, S., 1991, Physiology: energy acquisition and utilization, *in* Shumway, S.E., ed., Scallops: Biology, Ecology and Aquaculture. Developments in Aquaculture and Fisheries Science 21: Elsevier, Amsterdam, The Netherlands and New York, p. 305-346.

Bumpus, D.F., 1973, A description of the circulation on the continental shelf of the east coast of the United States: Progress in Oceanography, v. 6, p. 111-157.

CHUTE, A.S., WAINRIGHT, S.C., and HART, D.R., 2012, Timing of shell ring formation and patterns of shell growth in the sea scallop *Placopecten magellanicus* based on stable oxygen isotopes: Journal of Shellfish Research, v. 31, p. 649-662, doi: 10.2983/035.031.0308.

COLLING, A., ed., 2001. Ocean Circulation, 2nd edition: Butterworth Heinemann, Oxford, UK and New York, 286 p.

CRAIG, H., 1961, Standard for reporting concentrations of deuterium and oxygen-18 in natural waters: Science, v. 33, p. 1833-1834.

CRONIN, T.M., 1988. Evolution of marine climates of the U.S. Atlantic coast during the past four million years: Philosophical Transactions of the Royal Society of London B, v. 318, p. 661-678, doi: 10.1098/rstb.1988.0029.

CRONIN, T.M., 1991, Pliocene shallow water paleoceanography of the North Atlantic Ocean based on marine ostracodes: Quaternary Science Reviews, v. 10, p. 175–188, doi: 10.1016/0277-3791(91)90017-O.

CRONIN, T.M., and DOWSETT, H.J., 1996, Biotic and oceanographic response to the Pliocene closing of the Central American Isthmus, *in* Jackson, J.B.C., Budd, A.F., and Coates, A.G., eds., Evolution and Environment in Tropical America: University of Chicago Press, Chicago, Illinois, p. 76-104.

CSANADY, G.T., and HAMILTON, P., 1988, Circulation of slopewater: Continental Shelf Research, v. 8, p. 565-624, doi: 10.1016/0278-4343(88)90068-4.

DE SCHEPPER, S., HEAD, M.J., and LOUWYE, S., 2009, Pliocene dinoflagellate cyst stratigraphy, palaeoecology and sequence stratigraphy of the Tunnel-Canal dock, Belgium: Geological Magazine, v. 146, p. 92-112, doi: 10.1017/S0016756808005438.

DE SCHEPPER, S., GROENEVELD, J., NAAFS, B.D.A., VAN RENTERGHEM, C., HENNISSEN, J., HEAD, M.J., LOUWYE, S., and FABIAN, K., 2013, Northern hemisphere glaciation during the globally warm early Late Pliocene. PLoS ONE, vol. 8, no. e81508, 52.07 KB, http://journals.plos.org/plosone/article?id=10.1371%2Fjournal.pone.0081508. Checked August 2016.

DICKIE, L. M., 1958, Effects of high temperatures on survival of the giant scallop: Journal of the Fisheries Research Board of Canada. v. 15, p. 1189-1121.

DOWSETT, H.J., and WIGGS, L.B., 1992, Planktonic foraminiferal assemblage of the Yorktown Formation, Virginia: Micropaleontology, v. 38, p. 75-86, doi: 10.2307/1485844.

DOWSETT, H.J., CRONIN, T., POORE, R.Z., THOMPSON, R.S., WHATLEY, R.C., and WOOD, A.M., 1992, Micropaleontological evidence for increased meridional heat transport in the North Atlantic Ocean during the Pliocene: Science, v. 258, p. 1133-1135, doi: 10.1126/science.258.5085.1133.

DOWSETT, H.J., ROBINSON, M.M., HAYWOOD, A.M., HILL, D.J., DOLAN, A.M., STOLL, D.K., CHAN, W.-L., ABE-OUCHI, A., CHANDLER, M.A., ROSENBLOOM, N.A., OTTO-BLIESNER, B.L., BRAGG, F.J, LUNT, D.J. FOLEY, K.M., and RIESSELMAN, C.R., 2012, Assessing confidence in Pliocene sea surface temperatures to evaluate predictive models: Nature Climate Change, v. 2, p. 365-371, doi: 10.1038/NCLIMATE1455.

DOWSETT, H.J., FOLEY, K.M., STOLL, D.K., CHANDLER, M.A., SOHL, L.E., BENTSEN, M., OTTO-BLIESNER, B.L., BRAGG, F.J., CHAN, W.-L., CONTOUX, C., DOLAN, A.M., HAYWOOD, A.M., JONAS, J.A., JOST, A., KAMAE, Y., LOHMANN, G., LUNT, D.J., NISANCIOGLU, K.H., ABE-OUCHI, A., RAMSTEIN, G., RIESSELMAN, C.R., ROBINSON, M.M., ROSENBLOOM, N.A., SALZMANN, U., STEPANEK, C., STROTHER, S.L., UEDA, H., YAN, Q., and ZHANG, Z., 2013, Sea surface temperature of the mid-Piacenzian ocean: a data-model comparison: Scientific Reports, vol. 3, no. 2013, 55.51 KB, http://www.nature.com/articles/srep02013. Checked August 2016.

Dowsett, H., Dolan, A., Rowley, D., Moucha, R., Forte, A.M., Mitrovica, J.X., Pound, M., Salzmann, U., Robinson, M., Chandler, M., Foley, K., and Haywood, A., 2016, The PRISM4 (mid-Piacenzian) palaeoclimate reconstruction: Climate of the Past, v. 12, p. 1519-1538, doi: 10.5194/cp-12-1519-2016.

Elliot, M., deMenocal, P.B., Linsley, B.K., and Howe, S.S., 2003, Environmental controls on the stable isotopic composition of *Mercenaria mercenaria*: Potential application to paleoenvironmental studies: Geochemistry, Geophysics, Geosystems, vol. 4, iss. 7, 247.66 KB, http://onlinelibrary.wiley.com.ezproxy.derby.ac.uk/doi/10.1029/2002GC000425/full. Checked August 2016.

Epstein, S., Buchsbaum, R., Lowenstam, H.A., and Urey, H.C., 1953, Revised carbonate-water isotopic temperature scale: Bulletin of the Geological Society of America, v. 64, p. 1315-1326.

Fedorov, A.V., Brierley, C.M., Lawrence, K.T., Liu, Z., Dekens, P.S., and Ravelo, A.C., 2013, Patterns and mechanisms of early Pliocene warmth: Nature, v. 496, p. 43-49, doi: 10.1038/nature12003.

Fierstine, H.L., 2001, Analysis and new records of billfish (Teleostei: Perciformes: Istiophoridae) from the Yorktown Formation, early Pliocene of eastern North Carolina at Lee Creek Mine, *in* Ray, C.E., and Bohaska, D.J., eds., Geology and Paleontology of the Lee Creek Mine, North Carolina, III: Smithsonian Contributions to Paleobiology, no. 90, p. 21-70.

Goewert, A.E., and Surge, D., 2008, Seasonality and growth patterns using isotope sclerochronology in shells of the Pliocene scallop *Chesapecten madisonius*: GeoMarine Letters, v. 28, p. 327–338, doi: 10.1007/s00367-008-0113-7.

Gonfiantini, R., Stichler, W., and Rozanski, K., 1995, Standards and intercomparison materials distributed by the International Atomic Energy Agency for stable isotope measurements, *in* International Atomic Energy Agency, Reference and Intercomparison Materials for Stable Isotopes of Light Elements: IAEA-TECDOC-825, Vienna, Austria, p. 13-29.

Grossman, E.L., and Ku, T., 1986, Oxygen and carbon isotope fractionation in biogenic aragonite: Temperature effects: Chemical Geology (Isotope Geoscience Section), v. 59, p. 59–74, doi: 10.1016/0009-2541(86)90044-6.

Haveles, A.W., and Ivany, L.C., 2010, Rapid growth explains large size of mollusks in the Eocene Gosport Sand, United States Gulf Coast: PALAIOS, v. 25, p. 550-564, doi: 10.2110/palo.2009.p09-148r.

Hazel, J.E., 1971, Paleoclimatology of the Yorktown Formation (upper Miocene and Lower Pliocene) of Virginia and North Carolina: Bulletin du Centre de Recherches de Pau, SNPA (Société Nationale de Pétroles d'Aquitaine), v. 5 (suppl.), p. 361-375.

Hazel, J.E., 1988, Determining late Neogene and Quaternary palaeoclimates and palaeotemperature regimes using ostracods, *in* De Deckker, P., Colin, J.P., and Peypouquet, J.P., eds., Ostracoda in the Earth Sciences: Elsevier, Amsterdam, The Netherlands, p. 89–101.

Hickson, J.A., Johnson, A.L.A., Heaton, T.H.E., and Balson, P.S., 1999, The shell of the Queen Scallop *Aequipecten opercularis* (L.) as a promising tool for palaeoenvironmental reconstruction: evidence and reasons for equilibrium stable-isotope incorporation: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 154, p. 325-337, doi: 10.1016/S0031-0182(99)00120-0.

Huddlestun, P.F. 1988. A revision of the lithostratigraphic units of the coastal plain of Georgia, the Miocene through Holocene: Georgia Geological Survey Bulletin, v. 104, p. 1-162.

Johnson, A.L.A., Liquorish, M.N., and Sha, J., 2007. Variation in growth-rate and form of a Bathonian (Middle Jurassic) oyster in England, and its environmental implications: Palaeontology, v. 50, p. 1155-1173, doi: 10.1111/j.1475-4983.2007.00698.x.

Johnson, A.L.A., Hickson, J.A., Bird, A., Schöne, B.R., Balson, P.S., Heaton, T.H.E., and Williams, M., 2009, Comparative sclerochronology of modern and mid-Pliocene (c. 3.5 Ma) *Aequipecten opercularis* (Mollusca, Bivalvia): an insight into past and future climate change in the north-east Atlantic region: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 284, p. 164-179, doi: 10.1016/j.palaeo.2009.09.022.

Kirby, M.X., 2000, Paleoecological differences between Tertiary and Quaternary *Crassostrea* oysters as revealed by stable isotope sclerochronology: PALAIOS, v. 15, p. 132-141, doi: 10.2307/3515499.

Kirby, M.X., 2001, Differences in growth rate and environment between Tertiary and Quaternary *Crassostrea* oysters: Paleobiology, v. 27, p. 84-103, doi: 10.1666/0094-8373(2001)027<0084:DIGRAE>2.0.CO;2.

Knowles, T., Taylor, P.D., Williams, M., Haywood, A.M., and Okamura, B., 2009, Pliocene seasonality across the North Atlantic inferred from cheilostome bryozoans. Palaeogeography, Palaeoclimatology, Palaeoecology, v. 77, p. 226-235, doi: 10.1016/j.palaeo.2009.04.006.

Knowles, T., Leng, M.J., Williams, M., Taylor, P.D., Sloane, H.J., and Okamura, B., 2010, Interpreting seawater temperature range using oxygen isotopes and zooid size variation in *Pentapora foliacea* (Bryozoa): Marine Biology, v. 157, p. 1171-1180, doi: 10.1007/s00227-010-1397-5.

Krantz, D.E., 1990, Mollusk-isotope records of Plio-Pleistocene marine paleoclimate, U.S. Middle Atlantic Coastal Plain: PALAIOS, v. 5, p. 317–335, doi: 10.2307/3514888.

Krantz, D.E., 1991, A chronology of Pliocene sea-level fluctuations: The U.S. Middle Atlantic Coastal Plain record: Quaternary Science Reviews, v. 10, p. 163-174, doi: 10.1016/0277-3791(91)90016-N.

Krantz, D.E., Jones, D.S., and Williams, D.F., 1984, Growth rates of the sea scallop, *Placopecten magellanicus*, determined from the $^{18}O/^{16}O$ record in shell calcite: Biological Bulletin, v. 167, p. 186-199, doi: 10.2307/1541347.

Krantz, D.E., Kronick, A.T., and Williams, D.F., 1988, A model for interpreting continental-shelf hydrographic processes from the stable isotope and cadmium:calcium profiles of scallop shells: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 64, 123-140, doi: 10.1016/0031-0182(88)90002-8.

Krantz, D.E., Williams, D.F., and Jones, D.S., 1987, Ecological and paleoenvironmental information using stable isotope profiles from living and fossil molluscs: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 58, p. 249-266, doi: 10.1016/0031-0182(87)90064-2.

Lisiecki, L.E., and Raymo, M.E., 2005, A Pliocene-Pleistocene stack of 57 globally distributed benthic $\delta^{18}O$ records: Paleoceanography, vol. 20, no. PA1003, 322.11 KB, http://onlinelibrary.wiley.com.ezproxy.derby.ac.uk/doi/10.1029/2004PA001071/full. Checked December 2016.

Mallinson, D., Riggs, S., Thieler, E.R., Culver, S., Farrell, K., Foster, D.S., Corbett, D.R., Horton, B., and Wehmiller, J.F., 2005, Late Neogene and

Quaternary evolution of the northern Albemarle Embayment (mid-Atlantic continental margin, USA): Marine Geology, v. 217, p. 97-117, doi: 10.1016/j.margeo.2005.02.030.

Martínez-Boti, M.A., Foster, G.L., Chalk, T.B., Rohling, E.J., Sexton, P.F., Lunt, D.J., Pancost, R.D., Badger, M.P.S., and Schmidt, D.N., 2015, Plio-Pleistocene climate sensitivity evaluated using high-resolution CO_2 records: Nature, v. 518, p. 49-54, doi: 10.1038/nature14954.

Masson-Delmotte, V., Schulz, M., Abe-Ouchi, A., Beer, J., Ganopolski, A., González Rouco, J.F., Jansen, E., Lambeck, K., Luterbacher, J., Naish, T., Osborn, T., Otto-Bliesner, B., Quinn, T., Ramesh, R., Rojas, M., Shao, X., and Timmermann, A., 2013, Information from paleoclimate archives, *in* Stocker, T.F., Qin, D., Plattner, G.-K., Tignor, M., Allen, S.K., Boschung, J., Nauels, A., Xia, Y., Bex, V., and Midgley, P.M., eds., Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change: Cambridge University Press, Cambridge, UK and New York, p. 383-464.

McKinney, C.R., McCrea, J.M., Epstein, S., Allen, H.A., and Urey, H.C., 1950, Improvements in mass spectrometers for the measurement of small differences in isotope abundance ratios: Review of Scientific Instruments, v. 21, p. 724-730.

NDBC (National Data Buoy Center, US Department of Commerce, National Oceanic and Atmospheric Administration), undated, Southeast USA (including Chesapeake Bay) Historical Marine Data, http://www.ndbc.noaa.gov/maps/southeast_hist.shtml. Checked August 2016.

Okamura, B., O'Dea, A., and Knowles, T., 2011, Bryozoan growth and environmental reconstruction by zooid size variation: Marine Ecology Progress Series, v. 430, p. 133-146, doi: 10.3354/meps08965.

Olson, S.L., and Rasmussen, P.C., 2001, Miocene and Pliocene birds from the Lee Creek Mine, North Carolina, *in* Ray, C.E., and Bohaska, D.J., eds., Geology and Paleontology of the Lee Creek Mine, North Carolina, III: Smithsonian Contributions to Paleobiology, no. 90, p. 233-365.

Owen, R., Kennedy, H., and Richardson, C., 2002a, Isotopic partitioning between scallop shell calcite and seawater: Effect of shell growth rate: Geochimica et Cosmochimica Acta v. 66, p. 1727-1737, doi: 10.1016/S0016-7037(01)00882-1.

Owen, R., Kennedy, H., and Richardson, C., 2002b, Experimental investigation into partitioning of stable isotopes between scallop (*Pecten maximus*) shell calcite and sea water: Palaeogeography, Palaeclimatology, Palaeoecology, v. 185, p. 163-174, doi: 10.1016/S0031-0182(02)00297-3.

Pazzaglia, F.J., 1993, Stratigraphy, petrography, and correlation of late Cenozoic middle Atlantic Coastal Plain deposits: Implications for late-stage passive-margin geologic evolution: Geological Society of America Bulletin, v. 105, p. 1617-1634, doi: 10.1130/0016-7606(1993)105<1617:SPACOL>2.3.CO;2.

Pietrafesa, L.J., Morrison, J.M., McCann, M.P., Churchill, J., Böhm, E., and Houghton, R.W., 1994, Water mass linkages between the Middle and South Atlantic Bights: Deep Sea Research, Part II-Topical Studies in Oceanography, v. 41, p. 365–389, doi: 10.1016/0967-0645(94)90028-0.

Pinet, P.R., and Popenoe, P., 1985, A scenario of Mesozoic-Cenozoic ocean circulation over the Blake Plateau and its environs: Geological Society of America Bulletin, v. 96, p. 618-626, doi: 10.1130/0016-7606(1985)96<618:ASOMOC>2.0.CO;2.

Purdy, R.W., Schneider, V.P., Applegate, S.P., McLellan, J.H., Meyer, R.L., and Slaughter, B.H., 2001, The Neogene sharks, rays, and bony fishes from the Lee Creek Mine, Aurora, North Carolina, *in* Ray, C.E., and Bohaska, D.J., eds., Geology and Paleontology of the Lee Creek Mine, North Carolina, III: Smithsonian Contributions to Paleobiology, no. 90, p. 71-202.

Riggs, S.R., and Belknap, D.F., 1988, Upper Cenozoic processes and environments of continental margin sedimentation: eastern United States, *in* Sheridan, R.E. and Grow, J.A., eds., The Geology of North America, v. I-2, The Atlantic Continental Margin, U.S.: The Geological Society of America, Boulder, Colorado, p. 131-176.

Riggs, S., Snyder, S., Ames, D., and Stille, P., 2000, Chronostratigraphy of Upper Cenozoic phosphorites on the North Carolina continental margin and the oceanographic implications for phosphogenesis, *in* Glenn, C.R, Prév□t-Lucas, L., and Lucas, J., eds., Marine Authigenesis: From Global to Microbial: Society of Economic Paleontologists and Mineralogists, Special Publication, No. 66, p. 369-385.

Rowley, D.B., Forte, A.M., Moucha, R., Mitrovica, J.X., Simmons, N.A., and Grand, S.P., 2013, Dynamic topography change of the eastern United States since three million years ago: Science, v. 340, p. 1560-1563, doi: 10.1126/science.1229180.

Schmidt, D.N., 2007. The closure history of the Central American Seaway: evidence from isotopes and fossils to models and molecules, *in* Williams, M., Haywood, A.M., Gregory, F.J., and Schmidt, D.N., eds., Deep Time Perspectives on Climate Change: Marrying the Signal from Computer Models and Biological Proxies. The Micropalaeontological Society, Special Publications: The Geological Society, London, UK, p. 427-442.

Schöne, B.R., and Gillikin, D.P., 2013, Unraveling environmental histories from skeletal diaries – Advances in sclerochronology: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 373, p. 1-5, doi: 10.1016/j.palaeo.2012.11.026.

Snyder, S.W., Mauger, L.L., and Ames, D., 2001. Benthic foraminifera and paleoecology of the Pliocene Yorktown and Chowan River Formations, Lee Creek Mine, North Carolina, USA: Journal of Foraminiferal Research, v. 31, p. 244-274, doi: 10.2113/31.3.244.

Swift, D.J.P., Han, G., and Vincent, C.E., 1986, Fluid processes and sea floor response on a modern storm-dominated shelf: middle Atlantic shelf of North America. Part I, the storm current regime, *in* Knight, R.J., and McLean, J.R., eds., Shelf Sands and Sandstone Reservoirs: Canadian Society of Petroleum Geologists, Memoir 11, p. 99-119.

Tan, F.C., Cai, D., and Roddick, D.L., 1988. Oxygen isotope studies on sea scallops, *Placopecten magellanicus*, from Browns Bank, Nova Scotia: Canadian Journal of Fisheries and Aquatic Sciences, v. 45, p. 1378-1386.

Thébault, J., and Chauvaud, L., 2013, Li/Ca enrichments in great scallop shells (*Pecten maximus*) and their relationship with phytoplankton blooms. Palaeogeography, Palaeoclimatology, Palaeoecology, v. 373, p. 108-122, doi: 10.1016/j.palaeo.2011.12.014.

Tucker, M.E., and Wright, V.P., 1990, Carbonate Sedimentology: Blackwell, Oxford, 482 p.

Valentine, A., Johnson, A.L.A., Leng, M.J., Sloane, H.J., and Balson, P.S., 2011, Isotopic evidence of cool winter conditions in the mid-Piacenzian (Pliocene) of the southern North Sea Basin: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 309, p. 9-16, doi: 10.1016/j.palaeo.2011.05.015.

Wang, T., Surge, D., and Lees, J.M., 2015, ClamR: A statistical evaluation of isotopic and temperature records in sclerochronologic studies: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 437, 26-32, doi: 10.1016/j.palaeo.2015.07.008.

Ward, L.W., 1989, Tertiary stratigraphy of the central Virginia Coastal Plain, *in* Harris, W.B., Hurst, V.J., Nystrom, P.G., Jr., and Ward, L.W., Upper Cretaceous and Cenozoic Geology of the Southeastern Atlantic Coastal Plain. Field Trip Guidebook T172: 28th International Geological Congress, Washington, D.C., p. 63-90.

Ward, L.W., and Blackwelder, B.W., 1980, Stratigraphic revision of upper Miocene and lower Pliocene beds of the Chesapeake Group, Middle Atlantic Coastal Plain: U.S. Geological Survey Bulletin, v.1482-D, p. 1-61.

Ward, L.W., Bailey R.H., and Carter, J.G., 1991, Pliocene and early Pleistocene stratigraphy, depositional history, and molluscan paleobiogeography of the Coastal Plain, *in* Horton, J.W., Jr., and Zullo, V.A., eds., The Geology of the Carolinas: University of Tennessee Press, Knoxville, Tennessee, p. 274–289.

Williams, M., Haywood, A.M., Harper, E.M., Johnson, A.L.A., Knowles, T., Leng, M.J., Lunt, D.J., Okamura, B., Taylor, P.D., and Zalaziewicz, J., 2009, Pliocene climate and seasonality in North Atlantic shelf seas: Philosophical Transactions of the Royal Society of London, Series A, v. 367, p. 85–108, doi: 10.1098/rsta.2008.0224.

Winkelstern, I., Surge, D., and Hudley, J.W., 2013, Multiproxy sclerochronological evidence for Plio-Pleistocene regional warmth: United States Mid-Atlantic Coastal Plain: PALAIOS, v. 28, p. 649-660, doi: 10.2110/palo.2013.p13-010r.

Yamamoto, G., 1953, Ecology of the scallop, *Pecten yessoensis* Jay: Science Reports of the Tohoku University, 4th Series (Biology), v. 20, p. 11-32.

ZARRA, L., 1989, Sequence stratigraphy and foraminiferal biostratigraphy for selected wells in the Albemarle embayment, North Carolina: North Carolina Geological Survey, Open-File Report 89-5, p. 1-48.

ZHANG, Z.-S., NISANCIOGLU, K.H., CHANDLER, M.A., HAYWOOD, A.M., OTTO-BLIESNER, B.L., RAMSTEIN, G., STEPANEK, C., ABE-OUCHI, A., CHAN, W.-L., BRAGG, F.J., CONTOUX, C., DOLAN, A.M., HILL, D.J., JOST, A., KAMAE, Y., LOHMANN, G., LUNT, D.J., ROSENBLOOM, N.A., SOHL, L.E., and UEDA, H., 2013, Mid-pliocene Atlantic Meridional Overturning Circulation not unlike modern: Climate of the Past, v. 9, p. 1495-1504, doi: 10.5194/cp-9-1495-2013.

FIGURE CAPTIONS

FIG. 1.—Stratigraphy of the Yorktown Formation and correlative formations in the southern North Sea Basin, based on Krantz (1991), Williams et al. (2009), De Schepper et al. (2009), Winkelstern et al. (2013) and discussion herein (ages in Ma; MPWP = Mid-Pliocene/Piacenzian Warm Period). Note that Winkelstern et al. (2013) placed the Moore House Member after the MPWP (3.29-2.97 Ma) on the basis of Sr isotope dating but Krantz (1991) related it to a phase of warming and global ice-volume reduction identified in the deep-ocean $\delta^{18}O$ record at 3.1-3.0 Ma. Williams et al. (2009) placed the Rushmere Member before the MPWP but since it and the succeeding Morgarts Beach Member are representative of the same major transgression we consider, like Winkelstern et al. (2013), that at least part of the Rushmere Member must fall within the MPWP. Asterisks indicate units for which there are existing isotopic paleotemperature determinations (Krantz, 1990; Goewert

and Surge, 2008; Johnson et al., 2009; Williams et al., 2009; Valentine et al., 2011; Winkelstern et al. 2013).

FIG. 2.—Geography and Pliocene paleogeography of study area, and sample sites. **A**) Surface currents adjacent to the eastern US seaboard at present, principally based on Cronin (1988), Csanady and Hamilton (1988) and Böhm et al. (2006). Thick red arrow = Gulf Stream (strong, warm); thin red arrow = Carolina Coastal Current (weak, warm); thin blue arrows = (left) Virginia Coastal Current (weak, cool) and (right) 'slope current' (weak, cool); □200m contour included to show position of shelf edge. **B**) Enlargement of area indicated by box in A, with positions of Sunken Meadow (brown), Rushmere/Morgarts Beach (mauve) and Moore House (green) shorelines according to Ward et al. (1991). Numbers indicate collection locations referred to herein (coloured to match the relevant shoreline): 1 = Grove Wharf; 2 = Claremont; 3 = Lee Creek Mine, Aurora; 4 = Sunken Meadow Creek; 5 = Kingsmill; 6 = Fort Boykin, Burwell Bay; 7 = Lieutenant Run, Petersburg; 8 = Riddick Pit, Chuckatuck; 9 = Yadkin Pit, Deep Creek. Note that location 2 is actually closer to Sunken Meadow Creek than location 4. Locations 1, 2, 3, 4, 5 and 7 correspond to localities 31, 7, 49, 42, 81 and 43, respectively, of Ward and Blackwelder (1980). Location 6 is midway between localities 55 and 61 of Ward and Blackwelder (1980). Goewert and Surge (2008) and Ward (1989) provide sedimentary logs for locations 8 and 9, respectively.

FIG. 3.—Morphology of *Placopecten clintonius*. **A**) Specimen NC1, showing the positions of major (filled triangles) and moderate (open triangles) growth breaks indicated in Figs. 4E and 6E. **B**) Microgrowth increments in the mid-ventral sector of

specimen VA2 (filled triangle identifies the major growth break at 68 mm shell height indicated in Figs. 4B and 6B). Scale bars = 10 mm.

FIG. 4.—Oxygen isotope (red line), carbon isotope (black line) and microgrowth increment (blue line) data from *Placopecten clintonius* specimens from the Sunken Meadow Member of Virginia (**A-D**) and North Carolina (**E-H**). Isotopic axis reversed so that lower values of $\delta^{18}O$ (representative of higher temperatures) plot towards the top. Thin, dashed blue lines = raw increment data; thicker, continuous blue lines = five-point averages. Filled blue triangles = major growth breaks; open blue triangles = moderate growth breaks (indicated by a less pronounced 'step' in the shell profile). S1, S2… and W1, W2… refer respectively to summers and winters as identified from the $\delta^{18}O$ profiles. Stars = anomalous values which are excluded from subsequent analysis (see text for explanation). Exclusion of the two anomalous values in C has the effect of making the immediately preceding value representative of S2 (see Fig. 6C for clarification).

FIG. 5.—Patterns of variation in microgrowth increment size in *Aequipecten opercularis*. **A**) Three individuals (blue, green and yellow lines) from a seasonally stratified setting (50 m depth) in the Gulf of Tunis, Mediterranean Sea. **B**) Three individuals (blue, green and yellow lines) from the continuously well-mixed waters of the southern North Sea. Plots are five-point averages of the raw data, which has been excluded for purposes of clarity. Red lines = $\delta^{18}O$ profiles from the specimens providing the blue increment profiles in A and B (data from Johnson et al., 2009, figs. 5, 4C, respectively). Profiles of $\delta^{18}O$ for the specimens providing the green and yellow increment profiles in B are given in Johnson et al. (2009, fig. 4B, 4D,

respectively) but are not available for the specimens providing the green and yellow profiles in A. The scales of the axes are the same as those of Figure 4 to facilitate comparison (note, however, that the bounds of the increment height axes differ). Specimens represented in A are: blue increment profile (and $\delta^{18}O$ profile), Muséum National d'Histoire Naturelle, Paris (MNHN) IM-2008-1537; green increment profile, MNHN IM-2008-1539; yellow increment profile, MNHN IM-2008-1538. Specimens represented in B are: blue increment profile (and $\delta^{18}O$ profile), British Geological Survey (BGS), Zt 9955; green increment profile, BGS Zt 9953; yellow increment profile, BGS Zt 9957.

FIG. 6.—Temperature profiles from *Placopecten clintonius* specimens from the Sunken Meadow Member of Virginia (**A**-**D**) and North Carolina (**E**-**H**), calculated using the data in Figure 4 (anomalous points excluded), Equation 1 and values for $\delta^{18}O_{seawater}$ of −0.4‰, +0.7‰ and +1.1‰ (respectively, lower, middle and upper lines in each plot). The preferred profiles ($\delta^{18}O_{seawater}$ = +0.7‰) are indicated by a thicker line. Symbols indicating seasonal assignment (S1, S2…; W1, W2…) and growth breaks (filled and open blue triangles) are explained in Figure 4. Symbols for growth breaks beyond the ventral end of the temperature profile have been excluded from H.

FIG. 7.—False colour satellite images of the Atlantic Ocean off the coast of southeastern Canada and the northeastern US (Gulf of Maine to Cape Hatteras), showing geographic variation in environmental parameters of surface waters. **A**) Temperature. **B**) Concentration of phytoplankton pigments. Data collected on 14 June 1979 by the Coastal Zone Color Scanner on the Nimbus-7 satellite. In A the warmest water (about 25 °C) is shown by orange/red and the coldest (about 6 °C) by dark blue,

with water of intermediate temperature shown by yellow and green; in B the highest concentrations of phytoplankton pigment are shown by dark brown and the lowest by blue, with intermediate concentrations shown by yellow and green (land is light brown and clouds are white or beige). Note the low concentration of phytoplankton pigments (implied low primary productivity) in warm, Gulf Stream water (including a warm-core eddy east of the Delmarva Peninsula) and the higher concentration of phytoplankton pigments typical of the cooler water nearer the shelf (Slope Sea). The very high concentration of phytoplankton pigments close to much of the coastline is a reflection of nutrient input from the land, itself strongly influenced by human activities. Adapted from Colling et al. (2001, fig. 4.31), with permission from Elsevier.

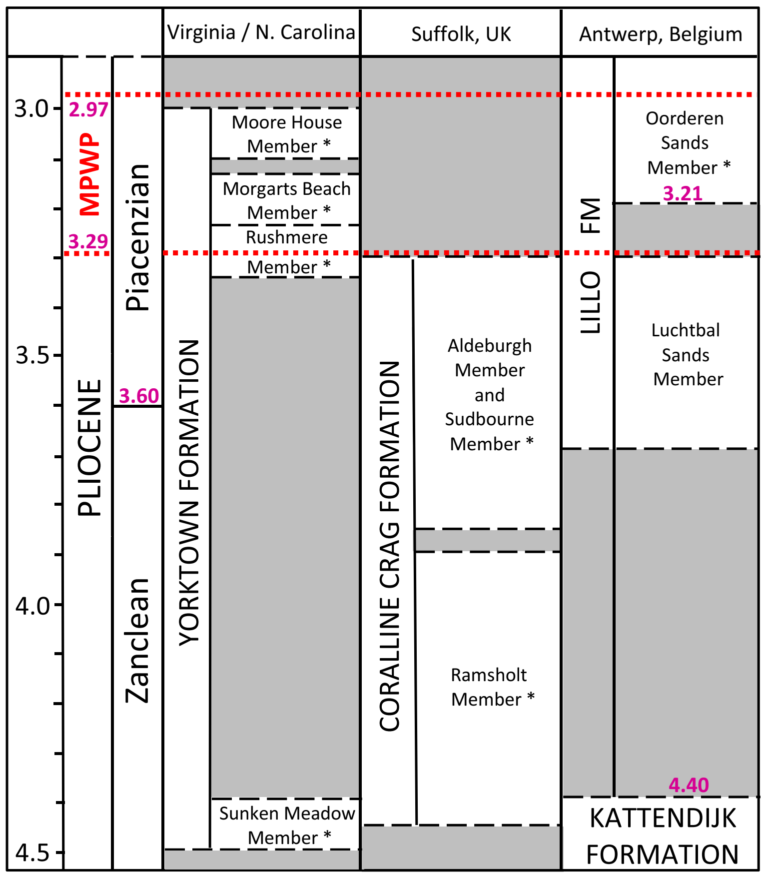
FIG. 8.—Envisaged disposition of currents in the area shown in Figure 2B during deposition of the basal Sunken Meadow Member. Brown line = position of shoreline according to Ward et al. (1991); position of shelf edge is that of present □200 m contour. The greater southward penetration of cool currents (medium thickness blue arrows) on shelf and slope compared to now (Fig. 2A) is explained in the text, as are the intrusions of slope water (thin blue arrows) onto the shelf. Arguably, shallows in the region of the Mid-Carolina Platform High (see text) may have prevented cool-current flow on the shelf as far south as is shown (i.e. into southernmost North Carolina). The point of meeting and eastward deflection of the slope current and Gulf Stream (thick red arrow) may have been farther south than shown. A similar disposition of currents is envisaged during deposition of most of the higher Yorktown Formation, but at times the Gulf Stream may have been more vigorous, penetrating farther north and influencing marine climate on the shelf (see text).

TABLE CAPTIONS

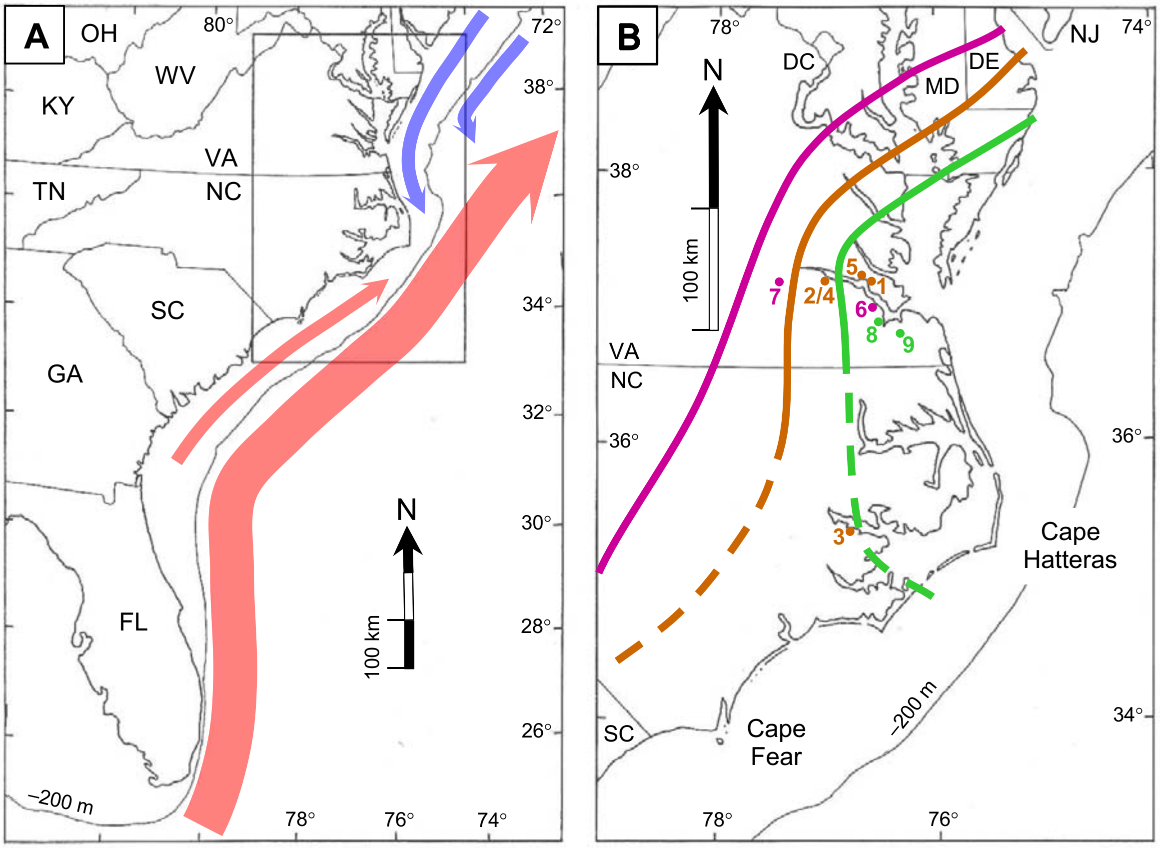
TABLE 1.—Mean winter minimum and mean summer maximum sea-surface temperatures, and seasonal ranges, for selected coastal to outer shelf locations up to approximately three degrees of latitude north and south of Cape Hatteras. Locations listed in order of decreasing latitude. Descriptions of shelf settings refer to relative position between the coast and shelf break, not water depth. Minimum and maximum temperatures were read from graphs of mean monthly temperature supplied by the National Data Buoy Center (NDBC, undated) and are accurate to the nearest whole number.

TABLE 2.—Calculated temperatures (ºC) for maximum (winter) and minimum (summer) values of $\delta^{18}O_{calcite}$ ($\delta^{18}O_c$; ‰) in scallop specimens from the Yorktown Formation, using selected values for $\delta^{18}O_{seawater}$ ($\delta^{18}O_{sw}$; ‰). [a] = specimens of *Placopecten clintonius* for which raw data are presented herein; [b] = specimens of *Chesapecten jeffersonius* (CJ), *C. madisonius* (CM) and *Carolinapecten eboreus* (EB) for which raw data are given by Krantz (1990), and [c] = specimens of *C. madisonius* (CMAD) for which raw data are given by Goewert and Surge (2008). The $\delta^{18}O$ values for *Chesapecten* and *Carolinapecten* are the extreme maxima and minima from the shells concerned as given in Krantz (1990, table 3) and Goewert and Surge (2008, table 2). These values have been assigned to a particular winter or summer by reference to the full $\delta^{18}O$ profiles supplied by these authors. The Summer 1 values from NC2, SM-CJ1 and LTRUN-EB are considered to be from incomplete summer records and have been handled accordingly (see text). The Summer 2 value from

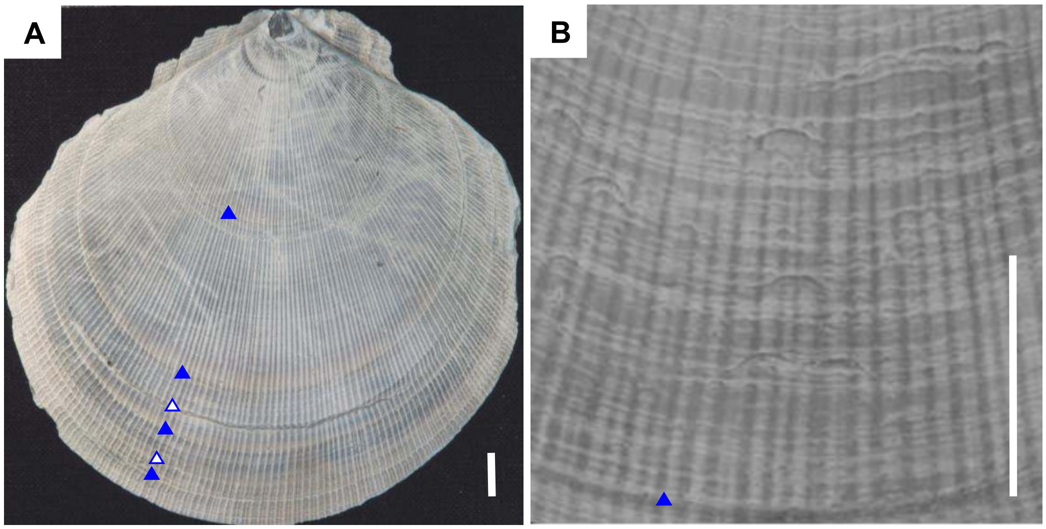
CMAD-4 has been corrected to agree with the corresponding profile (Goewert and Surge, 2008, fig. 3b).

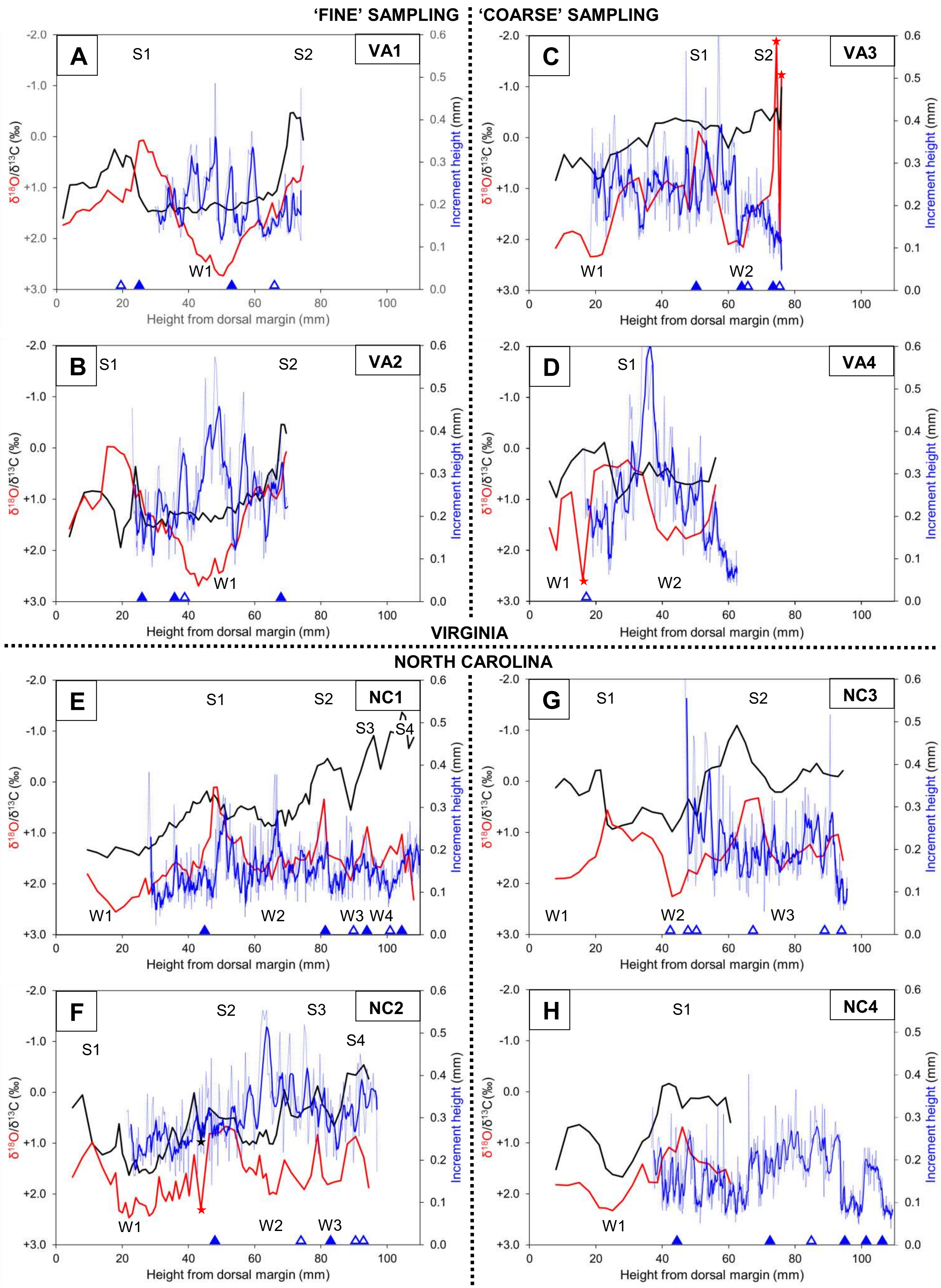


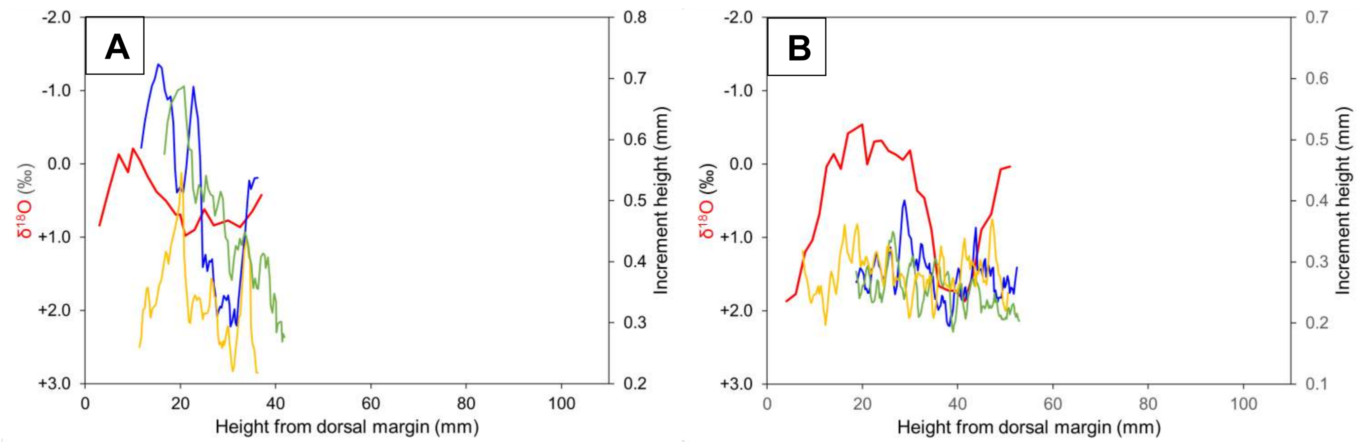
Virginia / N. Carolina
Suffolk, UK
Antwerp, Belgium
3.0
3.5
4.0
4.5
2.97
3.29
3.60
MPWP
PLIOCENE
Piacenzian
Zanclean
YORKTOWN FORMATION
Moore House Member *
Morgarts Beach Member *
Rushmere Member *
Sunken Meadow Member *
CORALLINE CRAG FORMATION
Aldeburgh Member and Sudbourne Member *
Ramsholt Member *
LILLO FM
Oorderen Sands Member *
3.21
Luchtbal Sands Member
4.40
KATTENDIJK FORMATION



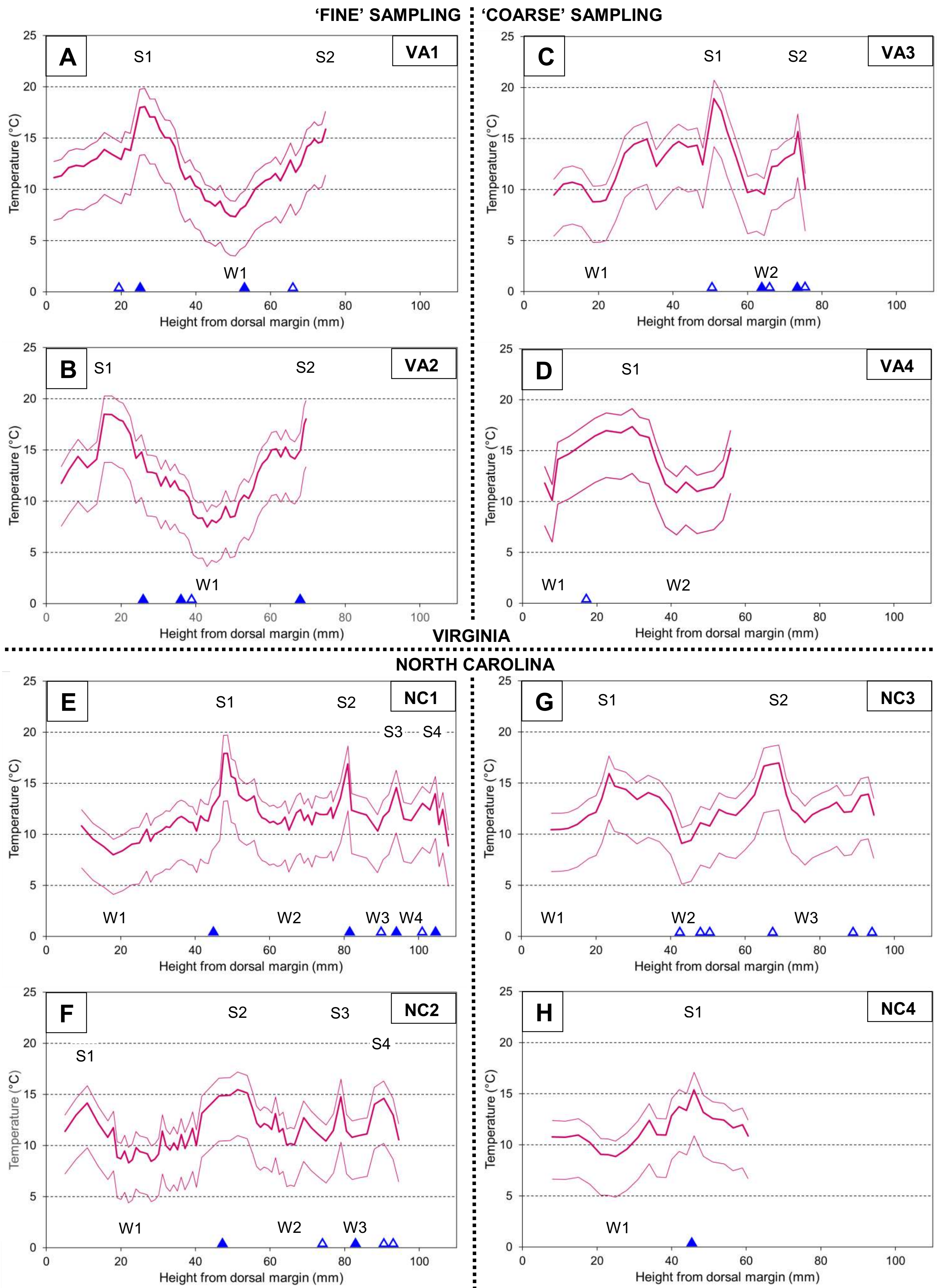
A
OH
80°
72°
WV
KY
38°
VA
NC
TN
SC
34°
GA
32°
30°
FL
28°
100 km
N
26°
–200 m
78°
76°
74°
B
78°
74°
NJ
DC
DE
MD
N
38°
100 km
5
1
7
2/4
6
8
9
VA
NC
36°
36°
3
Cape
Hatteras
–200 m
34°
SC
Cape
Fear
78°
76°


A
B

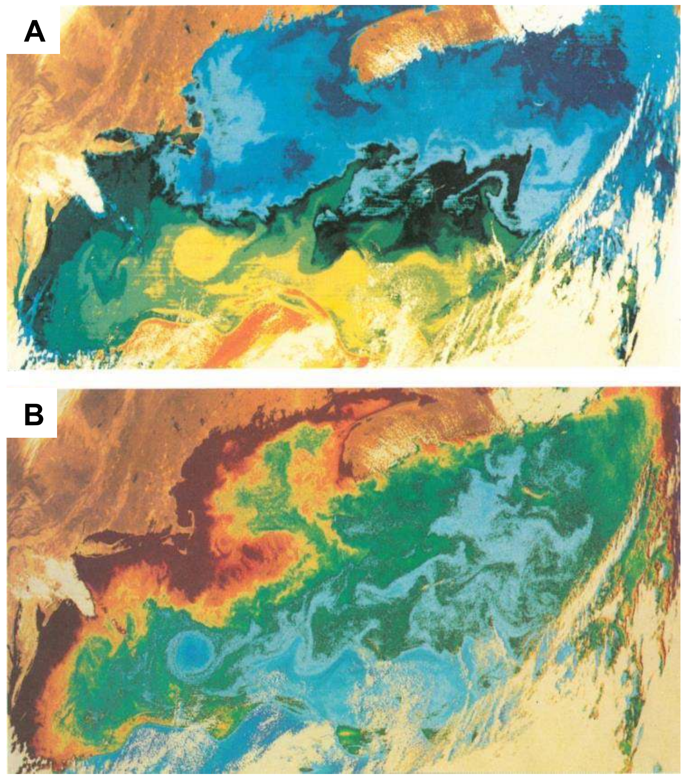

'FINE' SAMPLING
'COARSE' SAMPLING
VIRGINIA
NORTH CAROLINA
A
VA1
B
VA2
C
VA3
D
VA4
E
NC1
F
NC2
G
NC3
H
NC4
S1
S2
S3
S4
W1
W2
W3
W4
δ18O/δ13C (‰)
Increment height (mm)
Height from dorsal margin (mm)

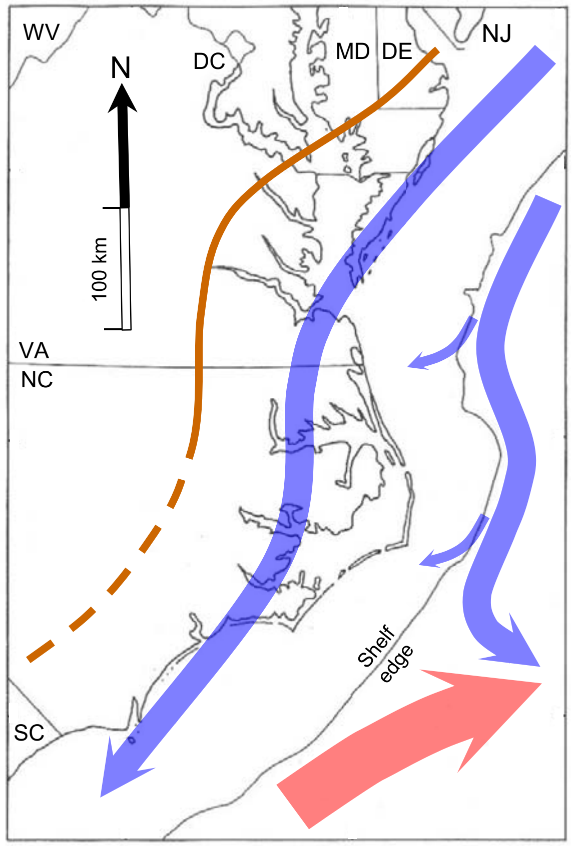


A
B
δ18O (‰)
Increment height (mm)
Height from dorsal margin (mm)



'FINE' SAMPLING
'COARSE' SAMPLING
A
VA1
B
VA2
C
VA3
D
VA4
VIRGINIA
NORTH CAROLINA
E
NC1
F
NC2
G
NC3
H
NC4
S1
S2
S3
S4
W1
W2
W3
W4
Temperature (°C)
Height from dorsal margin (mm)


A
B


WV
DC
MD
DE
NJ
N
100 km
VA
NC
SC
Shelf
edge

Station code	Setting	Latitude and longitude	Period of records	Minimum temp. (°C)	Maximum temp. (°C)	Range in temp. (°C)
44009	Inner shelf	38.5° N 74.7° W	1984-2008	5	23	18
OCIM2	Shore	38.3° N 75.1° W	2008-2012	5	24	19
KPTV2	Shore	37.2° N 76.0° W	2005-2012	5	26	21
CBBV2	Estuary	37.0° N 76.1° W	2005-2012	6	26	20
CHLV2	Inner shelf	36.9° N 75.7° W	1984-2005	6	25	19
44014	Outer shelf	36.6° N 74.8° W	1990-2008	9	25	16
44006	Mid-shelf	36.3° N 75.4° W	1980-1995	6	25	19
DUCN7	Shore	36.2° N 75.7° W	1997-2008	7	24	17
ORIN7	Shore	35.8° N 75.5° W	2005-2012	8	27	19
CAPE HATTERAS		35.3° N 75.5° W				
DSLN7	Mid-shelf	35.2° N 75.3° W	1984-2001	15	27	12
41025	Mid-shelf	35.0° N 75.4° W	2003-2008	16	28	12
JMPN7	Shore	34.2° N 77.8° W	2006-2012	11	28	17
41036	Mid-shelf	34.2° N 76.9° W	2006-2008	15	28	13
MROS1	Shore	33.7° N 78.9° W	2005-2012	10	29	19
FPSN7	Mid-shelf	33.5° N 77.6° W	1984-1996	17	28	11
41013	Mid-shelf	33.4° N 77.7° W	2003-2008	17	28	11
SCIS1	Shore	32.9° N 79.7° W	2006-2008	13	29	16
41004	Inner shelf	32.5° N 79.1° W	1978-2008	18	28	10
FRPS1	Shore	32.3° N 80.5° W	2006-2008	12	29	17

Member and shell ID	Winter 1				Winter 2				Summer 1				Summer 2			
	$\delta^{18}O_c$	Temperature			$\delta^{18}O_c$	Temperature			$\delta^{18}O_c$	Temperature			$\delta^{18}O_c$	Temperature		
		$\delta^{18}O_{sw}$ −0.4	$\delta^{18}O_{sw}$ +0.7	$\delta^{18}O_{sw}$ +1.1		$\delta^{18}O_{sw}$ −0.4	$\delta^{18}O_{sw}$ +0.7	$\delta^{18}O_{sw}$ +1.1		$\delta^{18}O_{sw}$ −0.4	$\delta^{18}O_{sw}$ +0.7	$\delta^{18}O_{sw}$ +1.1		$\delta^{18}O_{sw}$ −0.4	$\delta^{18}O_{sw}$ +0.7	$\delta^{18}O_{sw}$ +1.1
Sunken Meadow																
VA1[a]	+2.73	3.5	7.3	8.8	-	-	-	-	+0.07	13.4	18.1	19.9	+0.57	11.4	15.9	17.6
VA2[a]	+2.69	3.6	7.5	9.0	-	-	-	-	−0.03	13.8	18.5	20.3	+0.08	13.4	18.0	19.8
VA3[a]	+2.34	4.8	8.8	10.3	+2.15	5.5	9.5	11.1	−0.13	14.2	18.9	20.7	+0.62	11.2	15.7	17.4
VA4[a]	+2.00	6.0	10.1	11.7	+1.80	6.7	10.9	12.5	+0.23	12.8	17.4	19.1	-	-	-	-
NC1[a]	+2.55	4.1	8.0	9.5	+1.91	6.3	10.4	12.0	+0.10	13.3	17.9	19.7	+0.34	12.3	16.9	18.7
NC2[a]	+2.47	4.4	8.3	9.8	+2.01	6.0	10.1	11.7	+0.98	9.8	14.2	15.9	+0.67	11.0	15.5	17.2
NC3[a]	+1.90	6.4	10.5	12.0	+2.25	5.1	9.1	10.7	+0.56	11.4	15.9	17.7	+0.32	12.4	17.0	18.7
NC4[a]	+2.33	4.9	8.9	10.4	-	-	-	-	+0.69	10.9	15.4	17.1	-	-	-	-
KING-CJ[b]	-	-	-	-	+1.5	7.8	12.1	13.7	−1.2	18.8	23.9	25.8	-	-	-	-
SM-CJ1[b]	+1.9	6.4	10.5	12.1	-	-	-	-	−0.3	14.9	19.7	21.5	-	-	-	-
SM-CJ2[b]	-	-	-	-	+1.6	7.5	11.7	13.3	−0.8	17.1	22.0	23.9	-	-	-	-

Rushmere																
BB-CM[b]	+1.6	7.5	11.7	13.3	-	-	-	-	-	-	-	-	□1.9	22.0	27.3	29.3
Morgarts Beach																
LTRUN-EB[b]	+1.8	6.7	10.9	12.5	-	-	-	-	+0.3	12.5	17.1	18.8	-	-	-	-
Moore House																
YAD-CM1[b]	+1.6	7.5	11.7	13.3	-	-	-	-	□1.9	22.0	27.3	29.3	-	-	-	-
YAD-CM2[b]	+1.3	8.6	12.9	14.5	-	-	-	-	□2.0	22.5	27.8	29.8	-	-	-	-
YAD-EB1[b]	+1.3	8.6	12.9	14.5	-	-	-	-	□1.4	19.7	24.8	26.8	-	-	-	-
CMAD-2[c]	+2.7	3.6	7.5	8.9	-	-	-	-	□1.4	19.7	24.8	26.8	-	-	-	-
CMAD-4[c]	+2.5	4.3	8.2	9.7	-	-	-	-	-	-	-	-	□1.5	20.2	25.3	27.3
CMAD-5[c]	+2.2	5.3	9.3	10.9	-	-	-	-	-	-	-	-	□1.1	18.4	23.4	25.3