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Late Holocene climate change for the eastern interior United States: evidence from high-resolution $\delta^{18}O$ values of sagittal otoliths

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Abstract

Stable oxygen isotope values were determined for freshwater drum (*Aplodinotus grunniens*) sagittal otoliths recovered from the Eastman rockshelter archaeological site in northeast Tennessee to evaluate climate change for the eastern interior United States from 5500 calendar years ago to the present. Micromilled samples representing less than six days of otolith growth were extracted to acquire high-resolution intra-otolith variation in $\delta^{18}O$ values. Freshwater drum sagittal otoliths form annual bands due to thermally induced growth cessation below 10°C. $\delta^{18}O_{(H_2O)}$ values can be calculated once a high-resolution carbonate sample representing the beginning of the growing season is isolated. Maximum summer temperature can be calculated using the seasonal minimum $\delta^{18}O_{(CaCO_3)}$ value and the $\delta^{18}O_{(H_2O)}$ value.

Maximum summer temperatures calculated from the freshwater drum sagittae suggest that summer temperatures generally decrease from 29°C at ~5.5 ka to 22°C at ~0.3 ka. However, warmer climates at 2.9, 1.7–1.6, and 1.2–1.0 ka punctuate this trend. A more complete picture of the climate is reconstructed, because $\delta^{18}O_{(H_2O)}$ values, which are a function of the ratio of summer to winter precipitation, are also calculated. A relatively low average $\delta^{18}O_{(H_2O)}$ value of −8.1‰ VSMOW was calculated at 1.0 ka, suggesting cold winters, dry summers, and/or wet winters may have prevailed during part of the Medieval Warm Period in Tennessee. Contrary to studies suggesting that the Holocene was extremely stable and the Hypsithermal was invariably warm and dry, additional evidence suggesting both significant climate variability and evidence for a wetter mid-Holocene is presented. © 2001 Elsevier Science B.V. All rights reserved.

Keywords: Hypsithermal; Microsampling; Otoliths; Seasonality; Stable isotopes

1. Introduction

The Holocene has been characterized as an 'extremely stable' climate compared to glacial periods (Dansgaard et al., 1993). Greenland ice core $\delta^{18}O$ values manifest 4–5‰ millennial oscillations during glacial periods compared to relatively minor Holocene variation (Johnsen et al., 1992; Jouzel et al., 1993). Although the Holocene may be described as comparatively stable, polar ice cores may not record significant regional climate variation suggested by alternative climate proxies such as fossil pollen (Larsen et al., 1995). Additionally, there is a growing body of evidence that indicates climate has varied significantly and rapidly over the last 10 000 yr (e.g. Karlén and Kuylenstierna, 1996; Wolfe et al., 1996; Anderson et al., 1997; Xia et al., 1997).

These studies suggest that climate variation during this period was neither of similar magnitude, nor

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regionally synchronous (Sandweiss et al., 1999). For example, contrary to the traditional view that the Holocene Hypsithermal (9–4 kra) was invariably warm and dry, the Finger Lakes region (Dwyer et al., 1996; Mullins, 1998), eastern Canada (Duthie et al., 1996; Edwards et al., 1996), southeast Georgia (Leigh and Feeney, 1995), the tropical Sahara (Ganopolski et al., 1998) and the western Pacific were warm and wet (Gagan et al., 1998). Peak dryness in the mid-western United States was time-transgressive (Wright, 1992; Yu et al., 1997; Baker et al., 1998), and some evidence indicates that climate during the Hypsithermal may also have been variable (Valero-Garces et al., 1996; Sandweiss et al., 1999).

There is a strong link between Holocene climate variability and human cultural evolution (e.g. Wendland and Bryson, 1974; Lamb, 1995). Large-scale human migration often correlates with significant climatic shifts (e.g. Binford et al., 1997). Profound human cultural transitions that coincide with regional climate change are noted between 8 and 3 ka (Sandweiss et al., 1999), lending support that Holocene climate was not ‘extremely stable’. More recently, the Roman Warm Period (~300 BC–400 AD), Medieval Warm Period (~900–1300 AD), and Little Ice Age (~1300–1800 AD) are examples of climate periods that significantly influenced human culture (Lamb, 1995). Studies detailing the impact of climate change on human subsistence patterns, and on economic and social structures, show the importance for prediction of future regional and global climate change (e.g. Houghton et al., 1995).

A number of general circulation models (GCMs) are used to predict future global climate (Houghton et al., 1995; Scorer, 1997) and to evaluate the mechanisms that force climate change (e.g. COHMAP project members, 1988; Bartlein et al., 1998). However, environmental mathematical models fail to duplicate or predict real world conditions (Hall, 1991; Oreskes et al., 1994). In particular, GCM results only reproduce certain qualitative features of paleoclimate proxy data when used to postdict climate change (Hansen et al., 1997; Prentice et al., 1998). A greater breadth of paleoclimate data is necessary to define initial conditions for model experiments, highlight points of weakness in GCM results, and to understand how the climate system operates through data-model comparisons. Specifically, it is important to quantitatively characterize climate prior to 1850 AD because of the paucity of meteorological data.

In this study, we focus on the prehistoric climate of eastern interior North America. High-resolution paleoclimate data are rare for continental regions, particularly for the eastern interior United States (Royall et al., 1991; Wilkins et al., 1991; Jackson et al., 1997). Although qualitative floral and faunal studies suggest little change after 9.5 kra (Delcourt, 1979; Semken, 1983; Jackson et al., 1997), minor vegetation change to oak suggesting a drier climate than today is noted between 8–5 kra (Delcourt, 1979; Wilkins et al., 1991). However, these floral assemblages may not be sufficiently sensitive to short-term climate variation because ecotone displacement in this region occurs only after a sustained and systematic change in mean annual temperature of about 200 yr (Delcourt and Delcourt, 1987). Therefore, it is necessary to develop and use other proxy methods to quantitatively evaluate higher resolution climate change from the mid-Holocene to the present.

An alternative method involves the analysis of $\delta^{18}O$ values in carbonates. However, temperate continental climate records are difficult to interpret from stable isotope values due to the confounding influences of temperature and $\delta^{18}O_{(H_2O)}$ values (Grootes, 1993; Koch, 1998). These variables can be quantified by evaluating $\delta^{18}O$ values of accretionary carbonates at high-resolution (e.g. Dettman and Lohmann, 1995; Wurster et al., 1999). Fish otoliths are among the carbonate structures commonly found at archaeological sites.

Otoliths are accretionary aragonitic structures that usually grow incrementally within the inner ear of most fish such that they display daily and yearly bands (Panella, 1980). Sagittal otoliths (sagittae) are usually selected for sampling because they are the largest of three pairs of otoliths (Panella, 1980). These structures preserve a record of the fish’s growth rate, environmental parameters (such as temperature, salinity), and metabolic state (Degens et al., 1969; Kalish, 1991a; Patterson et al., 1993; Thorrold et al., 1997; Patterson, 1998). Growth banding occurs because of a periodic cessation of aragonitic deposition and/or an increase in the relative proportion of the protein otolin to aragonite (Degens et al., 1969). Deposition may cease because of daily fluctuation in temperature and nutrition, or by less frequent ecological and physiological stresses such as starvation and/or mating (Panella, 1980). Importantly, $\delta^{18}O_{(CaCO_3)}$ values are deposited in equilibrium with the environment (Kalish, 1991a,b; Patterson et al., 1993; Thorrold et al., 1997). When high-resolution seasonal variation in the $\delta^{18}O_{(CaCO_3)}$ values is acquired, $\delta^{18}O_{(H_2O)}$ values and growing season temperatures can be calculated (Patterson et al., 1993).

The purpose of this paper is to apply a micromilling technique for extracting high-resolution intra-otolith carbonate samples to evaluate climate change in Tennessee from the mid-Holocene (5.5 ka) to the present using $\delta^{18}O$ values of carbonate extracted from freshwater drum sagittal otoliths (sagittae). $\delta^{18}O$ values of freshwater drum sagittae recovered from the Eastman rockshelter yields a preliminary quantitative estimation of maximum summer temperatures and $\delta^{18}O_{(H_2O)}$ values for the eastern interior United States that test the traditional view of late Holocene climate change.

2. Methods

2.1. Study site and materials

The Eastman rockshelter PaleoIndian archaeological site located near the Interior Low Plateau of the Valley and Ridge physiographic division (Schuldenrein, 1996), adjacent to the South Fork Holston River in the northeast corner of Tennessee near Kingsport (Fig. 1). It is likely that the rockshelter was used, in part, because of access to fishing, suggesting that the abundant fish remains were from the nearby river. In particular, excavation of fish remains yielded abundant *Aplodinotus grunniens* (freshwater drum) sagittae. Twentieth-century climate of eastern Tennessee has highest mean monthly temperatures in July (24.2°C) and the lowest in January (3.0°C). Flooding prior to damming near the South Fork Holston River covered the Eastman rockshelter an average of once every 10 yr (Manzano, 1986).

Sixty-eight freshwater drum sagittae were recovered from 14 of 19 well-defined levels. Rare artifacts from the early Archaic period (10–8 ka) originally implied an age of 10 ka for the oldest levels; however, subsequent radiocarbon dates from level 19 suggest that the rare and oldest artifacts were reworked by later inhabitants. Because the earliest fish remains used in this study were found at level 17, a radiocarbon date from level 16 suggests that the oldest faunal remains used in this study are close to 5.5 ka. After level 16, clear transitions in lithic and ceramic artifacts from the late Archaic (6.0–2.9 ka) to the Mississippian (1.0–0.3 ka) indicate that later levels were deposited in stratigraphic continuity with no evidence of disturbance (Manzano, 1986). Changes in the development and use of artifacts (cultural periods) are in the predicted chronological sequence, and well constrained by radiocarbon dates (Fig. 2; Bense, 1994). Assuming linear sedimentation rate within each defined cultural period an age model for the levels at the Eastman rockshelter PaleoIndian archaeological site was developed (Fig. 2).

2.2. Micromilling of Eastman rockshelter sagittae and analysis of stable isotope values

Fourteen *A. grunniens* sagittae were micromilled to collect and analyze high-resolution intra-annually deposited carbonate powder for $\delta^{18}O$ and $\delta^{13}C$ values using the methodology described in Wurster et al. (1999). Specimens were polished to reveal growth annuli (Fig. 3). The otoliths were then attached to a stage beneath a fixed dental drill, and viewed on a monitor via a color digital camera. Annuli were digitized in real-time as a series of three-dimensional coordinates and 'smoothed' intermediate coordinates were interpolated using a cubic spline. Intermediate sampling paths were calculated between digitized curves. Sampling path arrays serve to guide three high-precision actuators, which position the sample stage relative to the fixed dental drill. Morphological features (annuli) were therefore used in conjunction with resulting $\delta^{18}O_{(CaCO_3)}$ values to determine representative years. In total, 3249 samples were micromilled (an average of 232 samples/specimen) in order to collect subweekly (~4–6 days) deposited carbonate. Smallest sample size was estimated to be 20–30 μg. Because only seasonal minima and maxima $\delta^{18}O_{(CaCO_3)}$ values are necessary for calculation of maximum summer temperatures and $\delta^{18}O_{(H_2O)}$ values, a subset of 1673 microsamples (targeting seasonal maxima and minima) were analyzed (an average of 120 samples/specimen). Mineralogy of

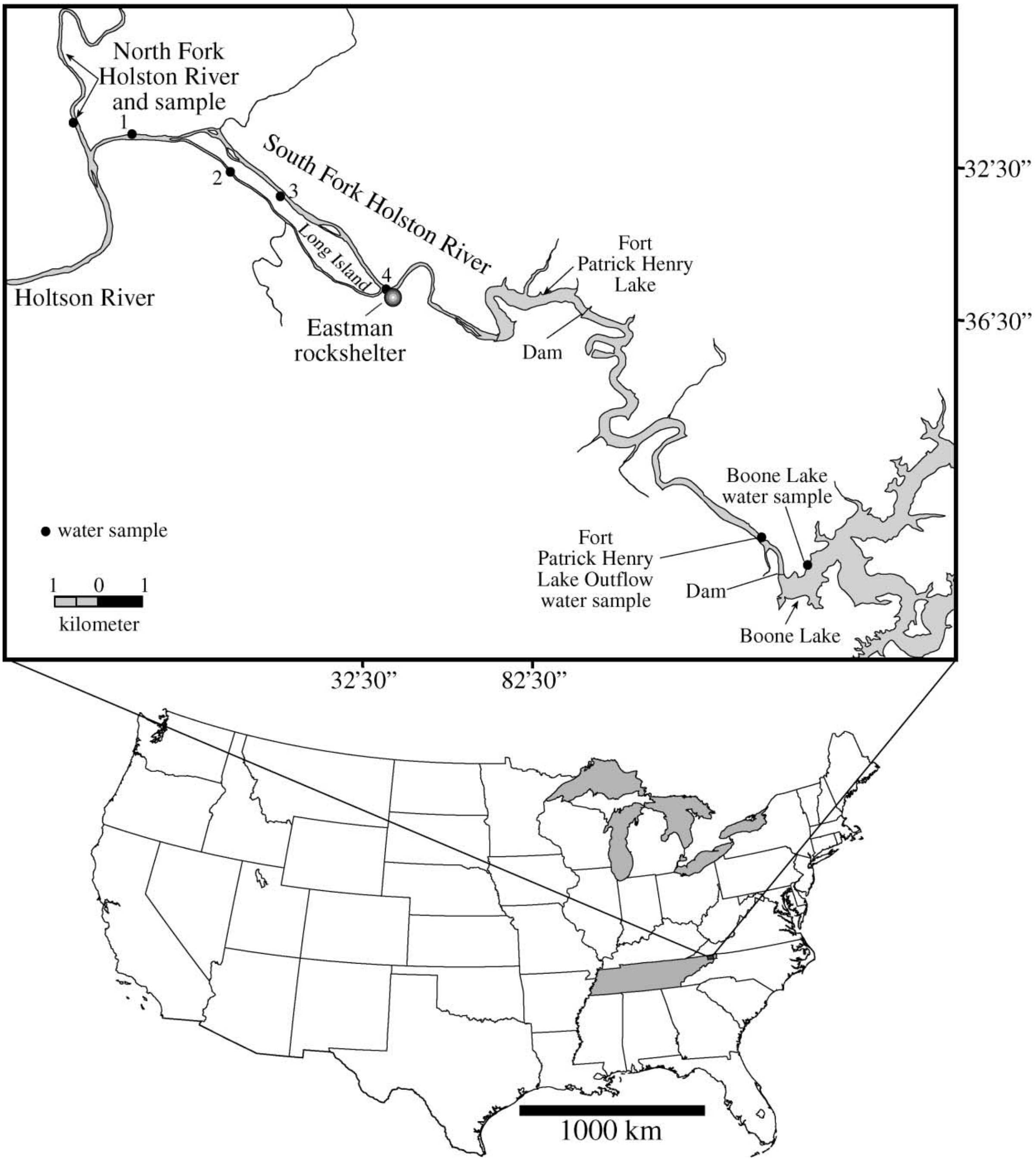

Fig. 1. Map showing the location of the Eastman rockshelter, its proximity to the South Fork Holston River, and the location of water samples collected 3/24/99.

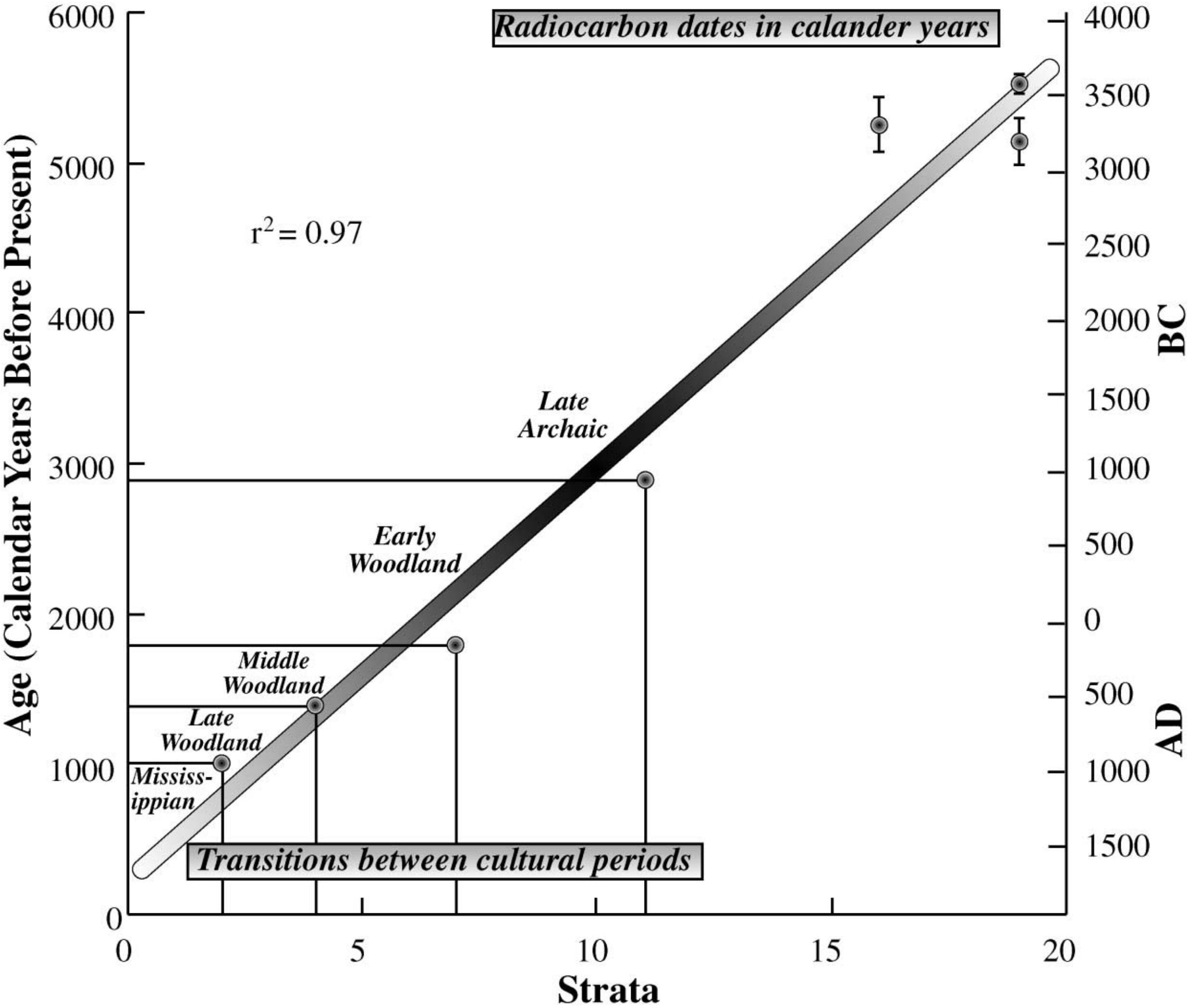

Fig. 2. Age model for Eastman rockshelter levels. Radiocarbon dates for level 16 and 19 and transitions between cultural periods provide constraints. Cultural Period ages are also marked.

sampled otoliths was determined by X-ray diffraction to ensure that only pristine otolith aragonite was analyzed.

Once samples have been milled, they are roasted in vacuo for 1 h at 200°C to remove volatiles that may interfere with $\delta^{18}O$ or $\delta^{13}C$ values. Samples were reacted in a Kiel III automated carbonate preparation device directly coupled to a Finnigan MAT 252 stable isotope ratio mass spectrometer at Syracuse University. Carbon dioxide was generated by reaction of carbonate with three drops of anhydrous phosphoric acid in individual reaction vessels at 70°C. Individual samples are run using a micro-inlet which reduces sample 'memory' and permits analysis of ~20 μg of carbonate. Individual samples of carbonate were analyzed with standard precision of ±0.08‰ (1σ). $\delta^{18}O_{(H_2O)}$ values were determined using a Finnigan HDO-II water equilibration device directly coupled to a Finnigan MAT 252 gas ratio mass spectrometer. Standard CO_2 gas is equilibrated with water samples for 6 h at 25°C and then sequentially analyzed. Values are reported to ±0.1‰. Replicate analyses of water samples were within ±0.1‰.

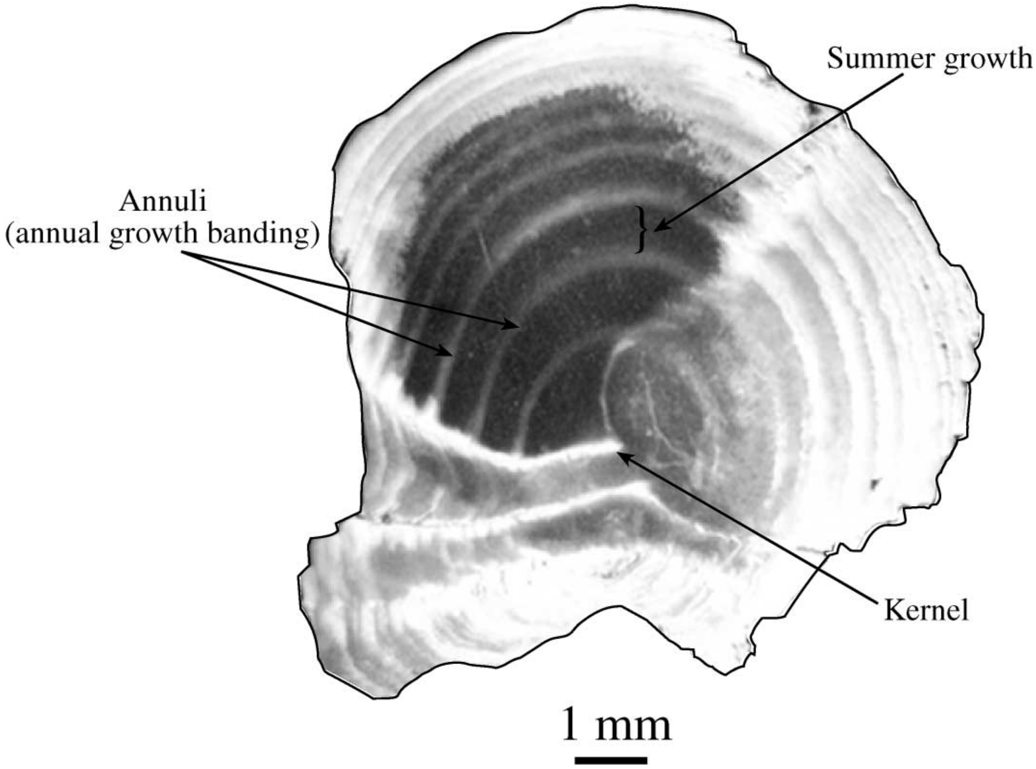

Fig. 3. Digital image of specimen sampled from 3.5 ka illustrates a typical prepared freshwater drum sagittal otolith.

2.3. Calculation of $\delta^{18}O_{(H_2O)}$ value and yearly maximum temperature

Freshwater drum (*Aplodinotus grunniens*) sagittae are particularly well suited for δ^{18}O paleothermometry. First, they are comparatively large, permitting greater sampling resolution. Second, the aragonite structures record water temperature, which closely tracks air temperature (McCombie, 1959; Edinger et al., 1968; Dingman, 1972; Webb, 1974; Livingstone and Lotter, 1998). Third, freshwater drum sagittae have a known growth cessation below 10°C (Patterson et al., 1993; McInerny and Held, 1995; Patterson, 1998). Therefore, maximum seasonal $\delta^{18}O_{(CaCO_3)}$ values occur at a temperature of 10°C making it possible to calculate a $\delta^{18}O_{(H_2O)}$ value. Once the $\delta^{18}O_{(H_2O)}$ value is known, maximum temperature for that particular year can be calculated using the $\delta^{18}O_{(H_2O)}$ value.

$\delta^{18}O_{(H_2O)}$ values and maximum summer temperatures are calculated as follows. (1) Using a seasonal maximum $\delta^{18}O_{(CaCO_3)}$ value and 10°C (the temperature at which the sagitta begins annually accreting), a $\delta^{18}O_{(H_2O)}$ value is calculated using the temperature-fractionation equation developed specifically for freshwater fish sagittal otoliths (Patterson et al., 1993). Patterson et al. (1993) used deep water obligate benthic species from the hypolimnion of the Laurentian Great Lakes and Lake Baikal, Siberia, to provide cold water end member values for aragonitic δ^{18}O values. Warm water values were obtained from naturally grown fish and from fish grown in aquaria under controlled conditions. The resultant equation is indistinguishable from the hypothesized inorganic aragonite δ^{18}O temperature-fractionation relationship (Campana, 1999). This calculated $\delta^{18}O_{(H_2O)}$ value is assumed to represent the beginning of the growing season. (2) We attempt to account for a seasonal change in $\delta^{18}O_{(H_2O)}$ value from the beginning of the growing season to the mid-summer by adjusting the calculated $\delta^{18}O_{(H_2O)}$ value by $+1 \pm 0.5‰$ (see Section 4.1). (3) The calculated $\delta^{18}O_{(H_2O)}$ value

$+1 \pm 0.5‰$ is used in conjunction with the subsequent seasonal minimum $\delta^{18}O_{(CaCO_3)}$ value to calculate maximum summer temperatures. In the case where a seasonal minimum $\delta^{18}O_{(CaCO_3)}$ value is not preceded by a seasonal maximum $\delta^{18}O_{(CaCO_3)}$ value (as would be the case of a fish born in the early summer) the next seasonal maximum $\delta^{18}O_{(CaCO_3)}$ value would be used in step 1. Therefore, the seasonal maximum $\delta^{18}O_{(CaCO_3)}$ value is directly related to the calculated $\delta^{18}O_{(H_2O)}$ value. The range between seasonal maximum and minimum $\delta^{18}O_{(CaCO_3)}$ values ($\Delta\delta^{18}O_{(Seasonal\ CaCO_3)}$) is directly related to maximum summer temperature.

3. Results

In order to interpret climate for the eastern interior United States, both maximum temperature and $\delta^{18}O_{(H_2O)}$ values were calculated from sagittal otolith $\delta^{18}O_{(CaCO_3)}$ values (Table 1). A comparison between average $\delta^{18}O_{(CaCO_3)}$ values of micromilled otoliths and $\delta^{18}O_{(CaCO_3)}$ values of whole otoliths from the same level suggest that representative specimens from each level were selected ($r^2 = 0.86$). However, an exception is the otolith micromilled at 1.2 ka, which may represent a transitional period between level 2 and 3 (Fig. 4). 1σ for $\delta^{18}O_{(H_2O)}$ value is calculated as 1 standard deviation of year-to-year variability rather than analytical error. The maximum summer temperature is estimated by assuming a $+1 \pm 0.5‰$ seasonal change in $\delta^{18}O_{(H_2O)}$ value from those calculated at the beginning of the growing season. 1σ is calculated as 1 standard deviation of the average maximum summer temperature; this also represents year-to-year variability rather than analytical error.

Three sagittal otoliths were sampled from ~5.5 ka (Fig. 5). Because a radiocarbon date for level 16 (5.3 ka) is intermediate between 2 radiocarbon dates at 19 (5.1 and 5.5 ka), sagittal otoliths sampled from levels 16 and 17 were assumed to be <5.5 ka. Maximum summer temperature and $\delta^{18}O_{(H_2O)}$ values were calculated from 3.5, 2.9 and 2.7 ka, using intra-otolith $\delta^{18}O_{(CaCO_3)}$ values (Fig. 6). Some of the highest $\delta^{18}O_{(H_2O)}$ values were calculated from seasonal maximum $\delta^{18}O_{(CaCO_3)}$ values from 2.9 ka. Intra-otolith $\delta^{18}O$ values for specimens sampled from 2.3, 1.8, 1.7, and 1.6 ka are illustrated in Fig. 7. High year-to-year variation is displayed for 2.3 and 1.7 ka while the specimen sampled from 1.6 ka is less variable. Variation in the fourth summer of specimen 1.8 ka does not show the typical seasonal pattern. Finally, 1.4, 1.2, 1.0, and 0.3–1.0 ka intra-otolith $\delta^{18}O$ values are plotted in Fig. 8.

Modern water samples from locations along the South Fork Holston River, North Fork Holston River, Boone Lake, and Fort Patrick outflow on 3/24/99 were collected and analyzed for $\delta^{18}O_{(H_2O)}$ values (Table 2). Damming the South Fork Holston River created both Boone and Fort Patrick Lakes. $\delta^{18}O_{(H_2O)}$ values from the South Fork Holston River averaged $-7.3 \pm 0.2‰$ VSMOW, and were 0.2‰ more positive than water analyzed from Boone dam and Fort Patrick. $\delta^{18}O_{(H_2O)}$ values from the South Fork Holston River were found to be approximately 0.5–1‰ lower than Tennessee rivers to the south (Coplen and Kendall, 2000).

4. Discussion

4.1. Interpretation of climate from $\delta^{18}O_{(CaCO_3)}$ values

$\delta^{18}O_{(H_2O)}$ values at the beginning of the growing season and estimated maximum summer temperatures were derived from the Eastman rockshelter freshwater drum sagittae (Fig. 9). The eastern United States receives the bulk of summer precipitation from air masses that originate in the Gulf of Mexico, while dominant winter storms generally arrive from the Pacific Ocean or Arctic Canada (Harmon, 1979; Gat et al., 1994). Air masses originating from the Pacific Ocean, Arctic Canada, and the Gulf of Mexico begin with average meteoric $\delta^{18}O_{(H_2O)}$ values near $-9‰$, $-19‰$, and $-3‰$ VSMOW, respectively (Rozanski et al., 1993). Therefore, variation in $\delta^{18}O_{(H_2O)}$ values calculated from sagittal otoliths may reflect a significant change in the ratio of winter precipitation to summer precipitation, as a result of changing atmospheric circulation patterns.

In order to calculate maximum summer temperatures, we assume minor seasonal variability in $\delta^{18}O_{(H_2O)}$ values of the South Fork Holston River based on published values for similar sized unimpounded rivers. The relatively recent impounding of this river does not permit a calibration of the

Table 1
Information retrieved from sagittae recovered from the Eastman rockshelter. Seasonal maximum $\delta^{18}O_{(CaCO_3)}$ value was used to calculate $\delta^{18}O_{(H_2O)}$ value at the beginning of the growing season using the aragonite temperature fractionation equation developed by Patterson et al. (1993) and using a thermal growth tolerance of 10°C. The $\delta^{18}O_{(H_2O)}$ value was then adjusted by 0.5 and 1.5‰ and used in conjunction with the subsequent seasonal minimum $\delta^{18}O_{(CaCO_3)}$ value to estimate a low and high maximum summer temperature, respectively. If a seasonal minimum $\delta^{18}O_{(CaCO_3)}$ value was not preceded by a seasonal maximum $\delta^{18}O_{(CaCO_3)}$ value, the subsequent calculated $\delta^{18}O_{(H_2O)}$ value was used (indicated by '′') or an average of $\delta^{18}O_{(H_2O)}$ values was used (indicated by '*')

Calendar age (ka)	Sagittal age	# of samples extracted/analyzed	Seasonal maximum $\delta^{18}O_{(CaCO_3)}$ value ‰$_{(VPDB)}$	Calculated $\delta^{18}O_{(H_2O)}$ value ‰$_{(VSMOW)}$	Seasonal minimum $\delta^{18}O_{(CaCO_3)}$ value ‰$_{(VPDB)}$	Range in summer maximum temperature (°C)	Mean summer maximum temperature (°C)	Mean $\delta^{18}O_{(H_2O)}$ value ‰$_{(VSMOW)}$
5.5		205/90	− 5.2	− 6.8	− 8.3	26.6–31.6		
Specimen 17-1			− 4.6	− 6.2	− 8.4	30.1–35.1		
			− 5.3	− 6.9	− 8.5	27.1–32.1		
	17		− 6.1	− 7.7	− 8.6	23.7–28.6	29.4 ± 2.8	− 6.9 ± 0.6
5.5		195/85		− 6.8'	− 7.6	23.2–28.1		
Specimen 16-1			− 5.2	− 6.8	− 8.3	26.6–31.6		
			− 4.3	− 5.9	− 8.0	29.6–34.6		
			− 4.5	− 6.1	− 7.6	26.6–31.5		
	9		− 3.4	− 5.0			28.8 ± 2.6	− 6.0 ± 0.7
5.5		325/315		− 7.9'	− 7.7	18.5–23.2		
Specimen 16-2			− 6.3	− 7.9	− 7.7	18.5–23.2		
			− 5.6	− 7.2	− 7.9	22.8–27.6		
	24		− 6.4	− 8.0	− 8.2	20.4–25.1	22.3 ± 2.2	− 7.7 ± 0.4
3.5		273/161		− 7.9'	− 8.3	21.3–26.1		
			− 6.3	− 7.9	− 9.4	26.6–31.6		
			− 6.5	− 8.1	− 8.8	22.8–27.6		
			− 5.6	− 7.2	− 8.7	26.6–31.6		
			− 5.7	− 7.3	− 8.3	24.2–29.1		
	12		− 6.2	− 7.8	− 8.5	22.8–27.6	26.5 ± 2.2	− 7.7 ± 0.4
2.9		192/78	− 3.8	− 5.4	− 6.9	26.6–31.5		
			− 4.5	− 6.1	− 7.9	28.1–33.1		
	18		− 4.3	− 5.9			29.8 ± 1.3	− 5.8 ± 0.4
2.7		245/115	− 6.8	− 8.4	− 9.8	26.1–31.1		
			− 6.6	− 8.2	− 9.0	23.2–28.1		
			− 6.7	− 8.3	− 8.2	19.0–23.7		
	12		− 6.5	− 8.1			25.1 ± 3.7	− 8.3 ± 0.1
2.3		235/115	− 7.0	− 8.6	− 8.3	18.1–22.8		
			− 7.3	− 8.9	− 8.6	18.1–22.8		
			− 5.8	− 7.4	− 7.8	21.3–26.1		
			− 5.6	− 7.2	− 8.4	25.2–30.1		
			− 6.6	− 8.2	− 8.4	20.4–25.2		
			− 5.8	− 7.4	− 7.7	20.9–25.6		
	14		− 5.8	− 7.4	− 8.7	25.6–30.6	23.7 ± 3.2	− 7.9 ± 0.7
1.8		215/113		− 7.9'	− 8.5	22.3–27.1		

			− 6.3	− 7.9	− 8.4	21.8–26.6		
			− 6.4	− 8.0	− 8.1	19.9–24.7		
			− 6.5	− 8.1	–	–		
	8			− 8.1*	− 8.3	20.4–25.2	23.6 ± 1.2	− 8.0 ± 0.1
1.7		172/88		− 6.6’	− 7.8	25.1–30.0		
			− 5.0	− 6.6	− 8.0	26.1–31.0		
			− 6.7	− 8.3	− 8.3	19.5–24.2		
			− 5.3	− 6.9	− 8.7	28.1–33.1		
	8		− 7.0	− 8.5			27.2 ± 3.8	− 7.6 ± 1.0
1.6		193/78		− 6.9’	− 8.4	26.6–31.6		
			− 5.3	− 6.9	− 8.2	25.6–30.6		
			− 6.2	− 7.8	− 8.4	22.3–27.1		
	21		− 6.4	− 8.0	− 8.4	21.3–26.1	26.4 ± 2.8	− 7.6 ± 0.6
1.4		240/89	− 7.3	− 8.9	− 8.8	19.0–23.7		
			− 7.0	− 8.6	− 8.6	19.5–24.2		
			− 6.7	− 8.3	–	–		
	18		− 6.2	− 7.8	− 8.6	23.2–28.1	22.9 ± 2.3	− 8.4 ± 0.5
1.2		165/82		− 7.4’	− 8.5	24.7–29.6		
			− 5.8	− 7.4	− 7.8	21.3–26.1		
			− 5.2	− 6.8	− 7.6	23.2–28.1		
			− 6.9	− 8.5	–	–		
	8		− 6.2	− 7.8	− 8.8	24.2–29.1	25.8 ± 1.6	− 7.8 ± 0.7
1.0		340/187		− 8.3’	− 9.4	24.7–29.6		
			− 6.7	− 8.3	− 10.2	28.6–33.6		
			− 6.4	− 8.0	− 9.1	24.7–29.6		
			− 6.5	− 8.1	− 8.6	21.8–26.6		
	21		− 6.3	− 7.9	− 8.4	21.8–26.6	26.8 ± 2.8	− 8.1 ± 0.2
0.3–1.0		254/82	− 6.4	− 8.0	− 8.7	22.8–27.6		
			− 6.5	− 8.2	− 8.5	20.9–25.6		
			− 6.8	− 8.4	− 8.2	18.5–23.2		
	7		− 7.1	− 8.7	− 8.4	18.1–22.8	22.4 ± 2.1	− 8.3 ± 0.3

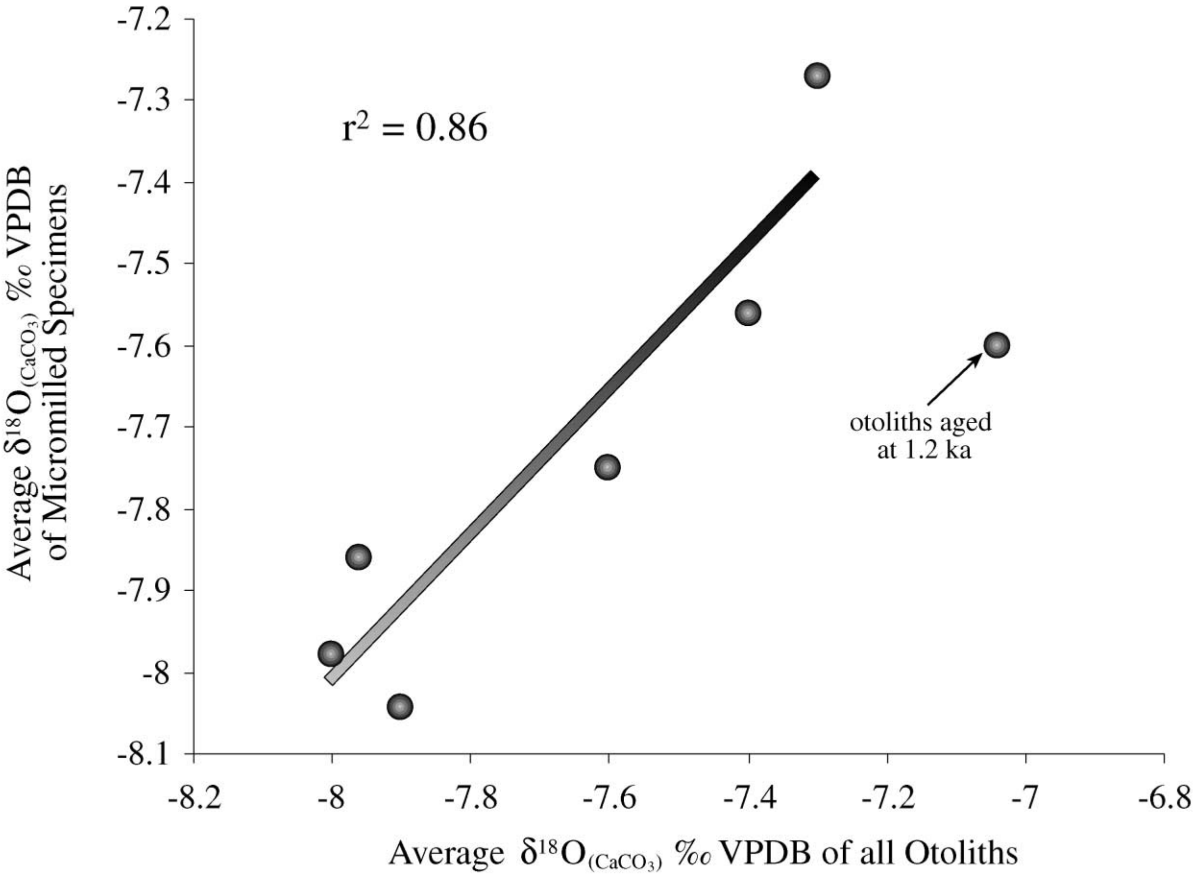

Fig. 4. Comparison of average micromilled $\delta^{18}O_{(CaCO_3)}$ values to values of whole otoliths recovered from identical strata. Most specimens selected for micromilling are representative samples. The specimen sampled from 1.2 ka is an exception, and the average $\delta^{18}O_{(CaCO_3)}$ value is significantly more positive than $\delta^{18}O_{(CaCO_3)}$ values of other specimens recovered from the same strata. Only those strata where at least four otoliths were recovered were used in the regression.

seasonal change in $\delta^{18}O_{(H_2O)}$ values for application to the mid-to-late Holocene. Today, many rivers from Tennessee display relatively constant seasonal $\delta^{18}O_{(H_2O)}$ values, or are controlled by releasing of dams (Coplen and Kendall, 2000). Therefore, examination of modern otoliths from the South Fork Holston River would be of little value because the damming of the river has substantially moderated seasonal variability in $\delta^{18}O_{(H_2}O)$ values.

River water is generally derived from surface runoff and groundwater (Fritz, 1981). The ratio of these two complimentary sources is dependant on the physical setting of the drainage basin and climatic parameters (Fritz, 1981). Although evaporation may significantly alter $\delta^{18}O_{(H_2O)}$ values in arid regions, it is insignificant in humid environments such as Tennessee (Stewart, 1975; Fritz, 1981; Gat, 1981; Ingraham et al., 1990; Ingraham and Criss, 1998). Seasonally, groundwater $\delta^{18}O$ values often display a moderation in the seasonal variation in $\delta^{18}O$ values of precipitation (Clark and Fritz, 1997). Below a critical depth there is no seasonal variation, and groundwater $\delta^{18}O$ values reflect average regional $\delta^{18}O$ values of precipitation (Clark and Fritz, 1997). Examples from the Eastern United States include springs from Missouri that displayed approximately a 1‰ variation in $\delta^{18}O_{(H_2O)}$ values over a year (Frederickson and Criss, 1999), while Harmon (1979) concluded that only storms of large magnitude and/or distinct $\delta^{18}O$ values maintain their identity in

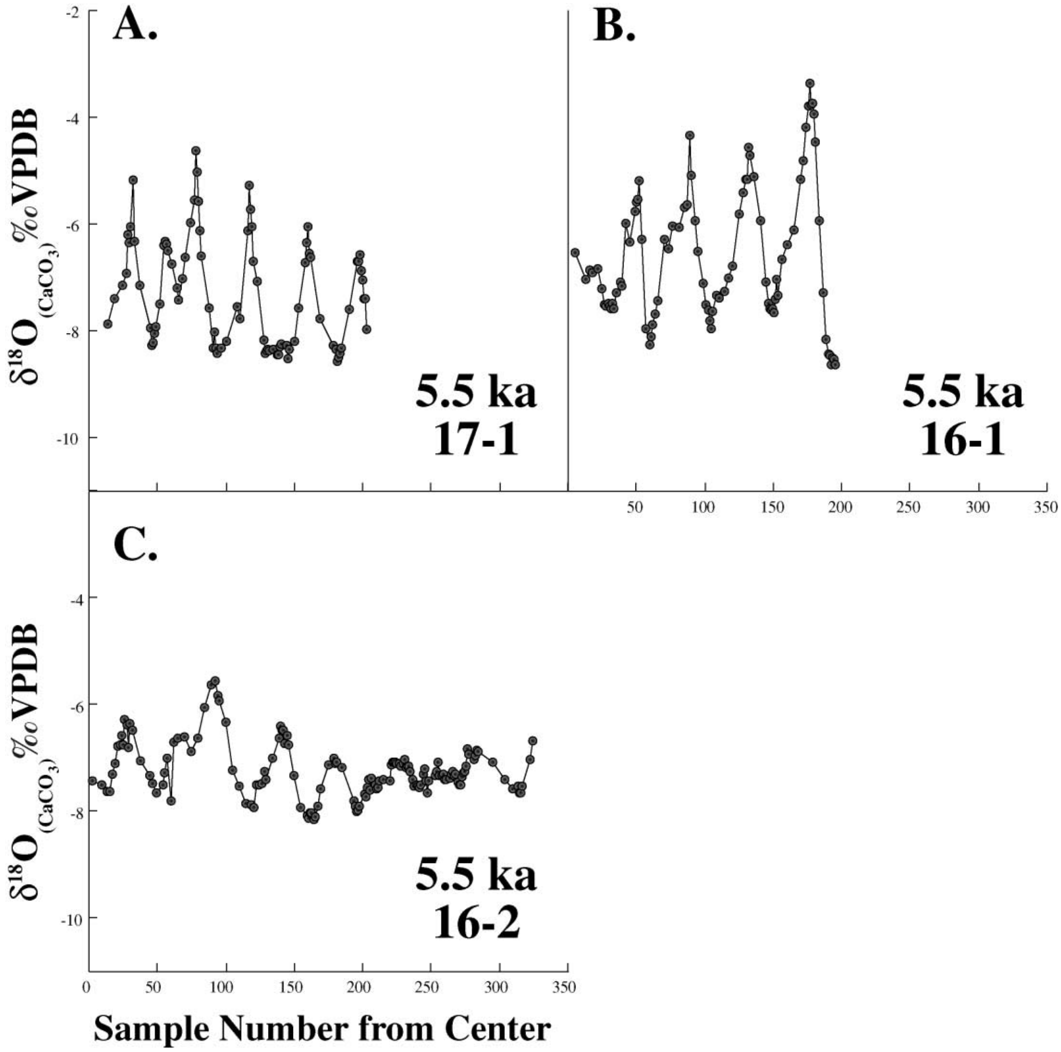

Fig. 5. Intra-otolith variation in $\delta^{18}O$ values for sagittae sampled from ~5.5 ka. Each plot represents one otolith. Each sample represents an individual carbonate aliquot removed by micromilling. Sample number from center identifies a consecutively numbered carbonate aliquot recovered sequentially from the origin of the otolith. For example, sample number 50 is the 50th consecutive aliquot from the kernel. Only a subset of micromilled samples was analyzed for $\delta^{18}O_{(CaCO_3)}$ value, therefore less samples are plotted on the figure then were recovered. (A) Sagitta 17-1 recovered from level 17 dated ~5.5 ka. (B) Sagitta 16-1 recovered from level 16 dated ~5.5 ka. (C) Sagitta 16-2 recovered from level 16 dated ~5.5 ka.

Kentucky groundwater. $\delta^{18}O_{(H_2O)}$ values of unimpounded temperate rivers that are similar in volume and drainage to the South Fork Holston River, increase approximately by $+1.0 \pm 0.5$‰ during the summer contemporaneous with maximum summer temperature (e.g. Clark and Fritz, 1997; Frederickson and Criss, 1999; Coplen and Kendall, 2000).

Although discrete $\delta^{18}O_{(H_2O)}$ values are calculated at the beginning of each growing season from seasonal maximum $\delta^{18}O_{(CaCO_3)}$ values of otoliths, we have to estimate the seasonal increase in $\delta^{18}O_{(H_2O)}$ value during the summer in order to calculate maximum summer temperature. Storm events may influence $\delta^{18}O_{(H_2O)}$ values during a given year; however, these storms should be readily recognized as a short deviation in the seasonal pattern of $\delta^{18}O_{(CaCO_3)}$ values.

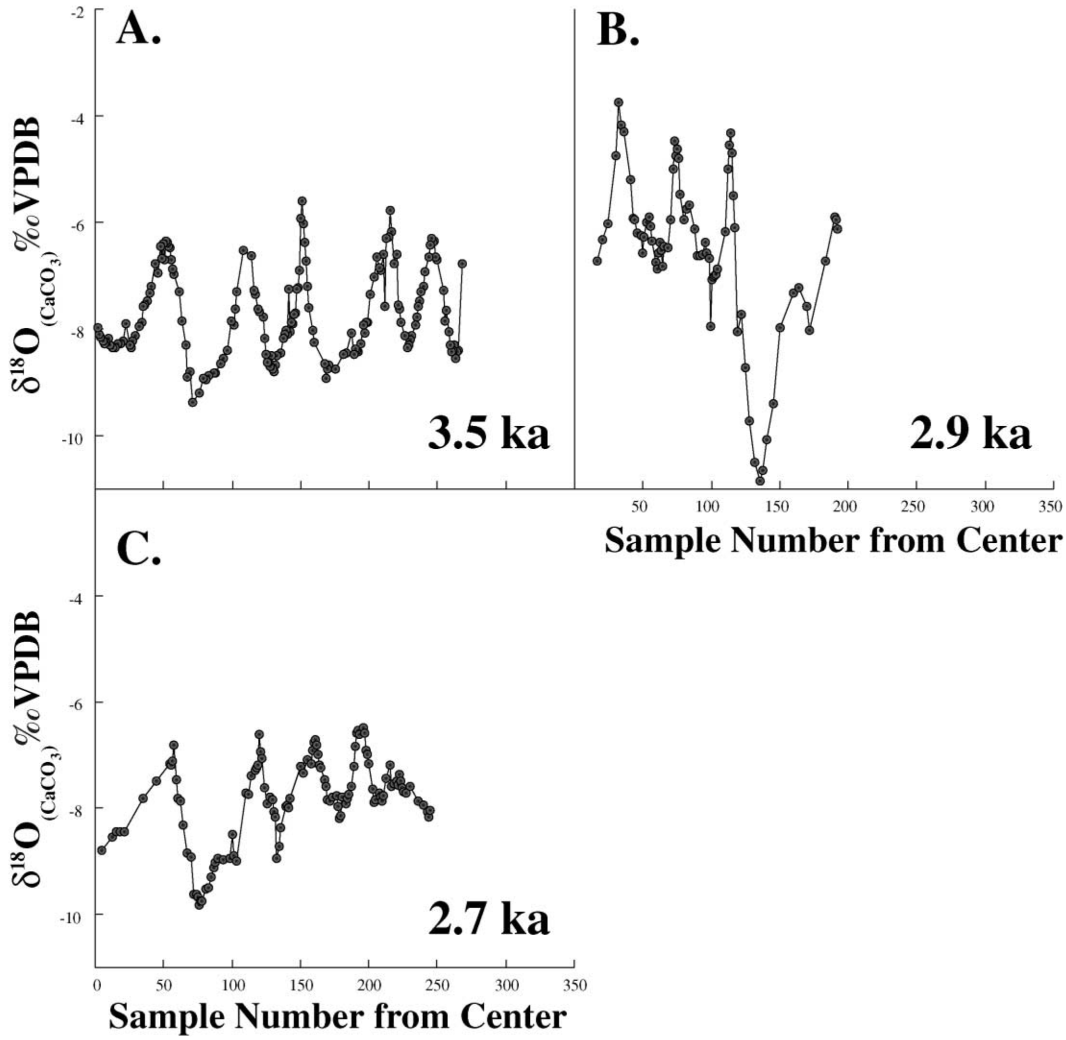

Fig. 6. Intra-otolith variation in $\delta^{18}O$ values for sagittae sampled from 3.5–2.7 ka. (A) Sagitta recovered from level 12 dated ~3.5 ka. (B) Sagitta recovered from level 11 dated ~2.9 ka. (C) Sagitta recovered from level 10 dated ~2.7 ka.

Anomalous events with distinctive $\delta^{18}O_{(H_2O)}$ values such as flooding or Hurricanes are also readily recognized by a deviant seasonal pattern in $\delta^{18}O_{(CaCO_3)}$ values. Of 63 analyzed $\Delta\delta^{18}O_{(seasonal\ CaCO_3)}$ values, 5 could not be explained solely as a function of temperatures, instead requiring an unusual seasonal pattern in $\delta^{18}O_{(H_2O)}$ values. We assume therefore that $\delta^{18}O_{(H_2O)}$ values at the maximum summer temperature to be $1 \pm 0.5‰$ more positive than those calculated at the beginning of the growing season (e.g. Clark and Fritz, 1997; Frederickson and Criss, 1999; Coplen and Kendall, 2000).

4.2. Trends in late Holocene climate in the eastern interior United States

Maximum summer temperatures and $\delta^{18}O_{(H_2O)}$ values are summarized in Fig. 9. Significant differences in estimated summer temperatures were calculated for three examined otoliths dated ~5.5 ka. Calculated average maximum summer temperatures for specimen 17-1 and 16-1 have a similar range and average approximately 29°C, and are amongst the highest in this study. By contrast, maximum summer temperatures for specimen 16-2 were

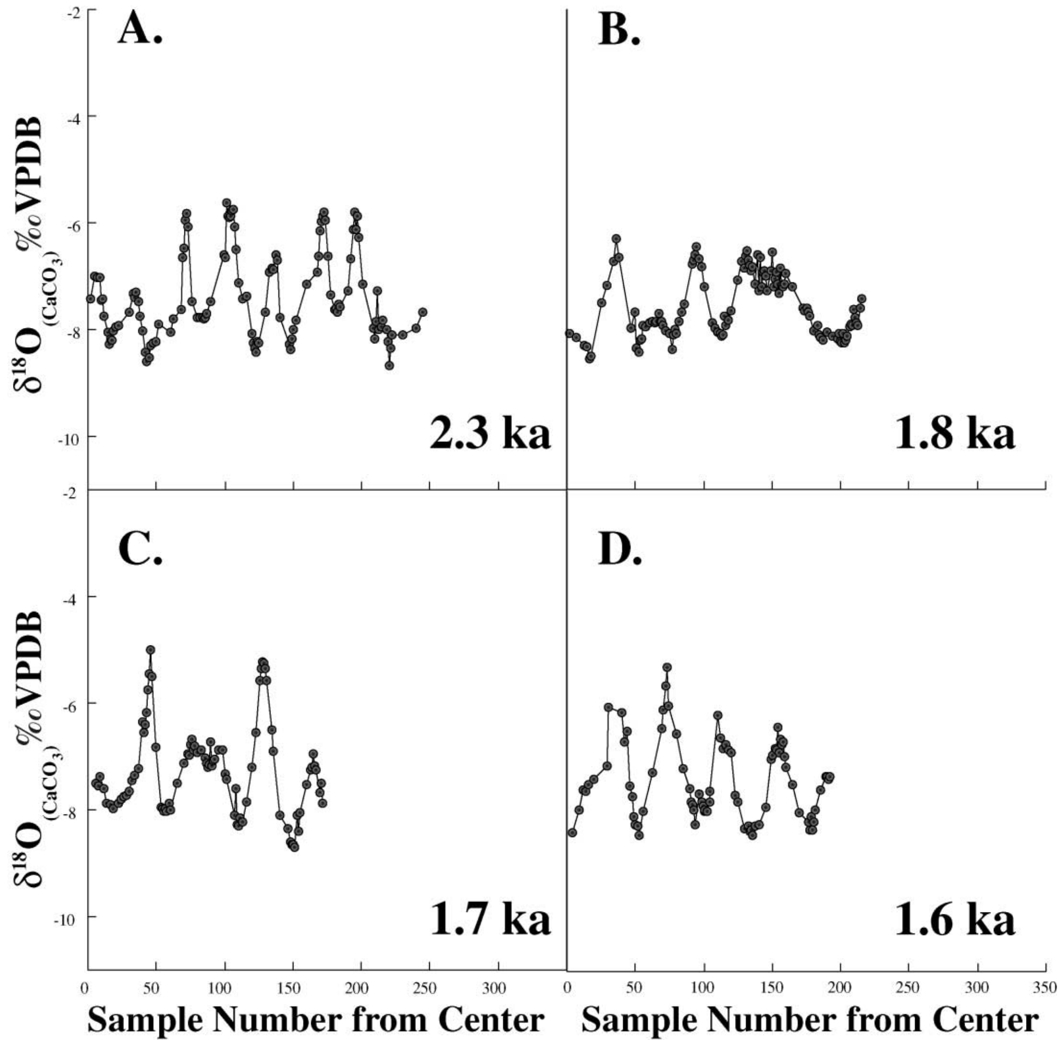

Fig. 7. Intra-otolith variation in δ^{18}O values for sagittae sampled from 2.3–1.6 ka. (A) Sagitta recovered from level 8 dated ~2.3 ka. (B) Sagitta recovered from level 7 dated ~1.8 ka. (C) Sagitta recovered from level 6 dated ~1.7 ka. (D) Sagitta recovered from level 5 dated ~1.6 ka.

approximately 8°C lower, amongst the lowest estimated in this study. Radiocarbon dates indicate that these levels were mixed and otoliths may have differed in age by several centuries (Fig. 2). It is possible that the difference in estimated temperature represents sample heterogeneity. For example, specimen 16-2 might have been captured from a location exhibiting increased seasonal $\delta^{18}O_{(H_2O)}$ variation such as a tributary; therefore, maximum summer temperatures would be significantly higher than those calculated. In order to obtain maximum summer temperatures comparable to the 2 other specimens, tributary $\delta^{18}O_{(H_2O)}$ values must have been ~2.5‰ more positive during a significant portion of each summer and be of sufficient size to support these relatively large fish. While this scenario is possible, it is unlikely that two relatively large rivers draining the same region would have such distinctive $\delta^{18}O_{(H_2O)}$ values. Rather, we suggest that temperatures calculated from mid-Holocene sagittae indicate high variability. Generally, warm conditions at ~5.5 ka agree with other Hypsithermal studies (e.g. COHMAP project members, 1988).

Sagittal otoliths 17-1and 16-1 from 5.5 ka also

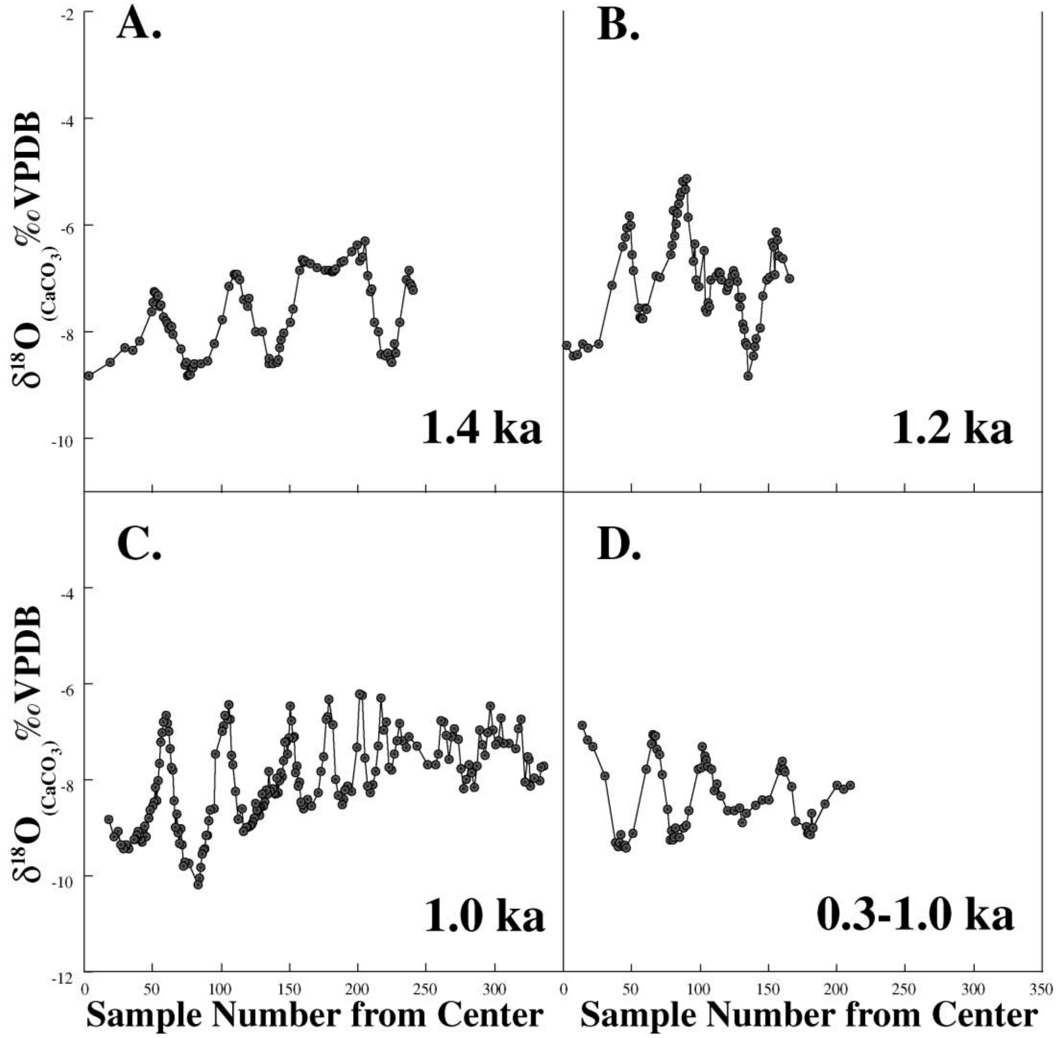

Fig. 8. Intra-otolith variation in $\delta^{18}O$ values for sagittae sampled from 1.4–0.3 ka. (A) Sagitta recovered from level 4 dated ~1.4 ka. (B) Sagitta recovered from level 3 dated ~1.2 ka. (C) Sagitta recovered from level 2 dated ~1.0 ka. (D) Sagitta recovered from level 1 dated 1.0–0.3 ka.

recorded relatively more positive $\delta^{18}O_{(H_2O)}$ values (only exceeded by the specimen sampled from 2.9 ka), which may indicate a greater proportion of moisture derived from the Gulf of Mexico relative to the modern. Evidence north of Tennessee at this time indicates high lake levels presumably resulting from greater precipitation (Dwyer et al., 1996; Edwards et al., 1996; Duthie et al., 1996; Mullins, 1998), which agrees with pollen data that indicate increasing moisture (Delcourt, 1979).

Estimated maximum summer temperatures are estimated to have decreased by 3.5 ka to ~27°C. By 2.9 ka, temperatures are estimated to have been 30 ± 1°C. The similarity of estimated maximum summer temperature and calculated $\delta^{18}O_{(H_2O)}$ values for this specimen to those for 16-1 and 17-1 suggest a return to Hypsithermal-like conditions. A return to such conditions is not generally described at ~2.9 ka, although some European historical records and peat bog cores (Lamb, 1995), northern hemisphere *Picea mariana* tree-line migration (Wolfe et al., 1996), and $\delta^{18}O$ values of ostracodes (Xia et al., 1997) suggest a century-long return to Holocene climate optimum conditions at ~1000 BC.

Table 2
$\delta^{18}O$ values from water samples collected from sites located close to the Eastman rockshelter. Water samples were collected on 3/24/99; the location of each sample is shown on Fig. 1. Damming the South Fork Holston River created both Boone and Fort Patrick Lakes.

Location	$\delta^{18}O_{(H_2O)}$ value (‰ VSMOW)
South Fork Holston River sample 1.	− 7.7
South Fork Holston River sample 2.	− 7.1
South Fork Holston River sample 3.	− 7.1
South Fork Holston River sample 4.	− 7.7
South Fork Holston River average	**− 7.3 ± 0.2**
North Fork Holston River sample	− 7.4
Fort Patrick Lake outflow	− 7.4
Boone Lake	− 7.6

After 2.7 ka, calculated $\delta^{18}O_{(H_2O)}$ values remain within 0.5‰, suggesting that summer/winter precipitation may not have significantly fluctuated. Temperatures are comparatively low at 2.3 and 1.8 ka, with maximum summer temperatures averaging nearly 2°C lower than the 20th century. The 2.3 ka sagittal otolith records temperatures for the first two years and year 5 approximately 4–6°C lower than years 3 and 4, and calculated $\delta^{18}O_{(H_2O)}$ values shift more than 1‰ a total of three times suggesting a variable interannual climate. Seasonal maximum $\delta^{18}O_{(CaCO_3)}$ values are more consistent at 1.8 ka than at 2.3 ka (Fig. 8).

A return to warmer conditions is supported by an increase in maximum summer temperature and $\delta^{18}O_{(H_2O)}$ values at 1.7 and 1.6 ka corresponding with the Roman Warm Period (~400 AD) evident in European records (Lamb, 1995; Hass, 1996). Temperature estimates for 1.7 ka suggest large inter-annual fluctuation, but generally warm conditions. It appears that by 1.6 ka, inter-annual maximum summer temperatures were less variable. Because $\delta^{18}O_{(H_2O)}$ values are consistent with relatively warm summer temperatures, this may imply more advected moisture during the summer relative to winter. Of note, the sagittal otolith at 1.7 ka undergoes alternating years with high and low $\delta^{18}O_{(H_2O)}$ values. This is interpreted to be reflective of a variable climate, and suggests that years relatively high in summer moisture may have been followed by years with a greater proportion of winter precipitation.

Maximum summer temperatures estimated at 1.4 ka are relatively cool and associated with relatively low $\delta^{18}O_{(H_2O)}$ values. However the maximum summer temperatures is estimated to have increased by 1.2 and 1.0 ka. The warm maximum summer temperature at 1.0 ka is coincident with a relatively low average $\delta^{18}O_{(H_2O)}$ value (−8.1‰ VSMOW). It is interpreted that a greater amount of snowfall and/or winter rainfall occurred at 1.0 ka. Higher temperature could potentially lead to greater advection of moisture and consequently greater precipitation during winter months (Rind et al., 1991). An alternate explanation is that colder winters and/or drier summers could have maintained low $\delta^{18}O$ values of the South Fork Holston River.

Following 1.0 ka, average maximum summer temperature is estimated to have decreased from 27 to 22°C, and the average $\delta^{18}O_{(H_2O)}$ value decreased to −8.3‰ VSMOW, suggesting that summers are colder and/or winters are longer. Although, the geologic age of the youngest otolith cannot be constrained further than 1.0–0.3 ka, it demonstrates that warm conditions did not continue through the Mississippian Cultural Period. This is amongst the lowest maximum summer temperature estimates in this study, averaging nearly 2°C lower than the 20th century average, perhaps indicating Little Ice Age cooling in Tennessee.

4.3. Anomalous $\Delta\delta^{18}O_{(seasonal\ CaCO_3)}$

Five $\Delta\delta^{18}O_{(seasonal\ CaCO_3)}$ values from 5 specimens indicated anomalous seasonal $\delta^{18}O_{(H_2O)}$ value patterns. The specimen sampled at 1.4 ka yields an unusually low maximum summer temperature (14°C) for the fourth summer (Fig. 8A). An increase in summer $\delta^{18}O_{(H_2O)}$ values in excess of the estimated 1‰ as a result of increased summer precipitation or migration to a tributary, combined with lower summer temperatures may explain the low calculated maximum summer temperature. Additionally fourth summer variability in $\delta^{18}O_{(CaCO_3)}$ values from the 2.7 ka specimen probably indicates fluctuating $\delta^{18}O_{(H_2O)}$ values throughout the summer. It is possible that this year was cool and wet with high variation in $\delta^{18}O_{(H_2O)}$ values.

Particular attention is given to three anomalous $\Delta\delta^{18}O_{(seasonal\ CaCO_3)}$ events, where $\delta^{18}O_{(CaCO_3)}$ values are significantly lower during the summer. Because the calculated maximum summer temperature of

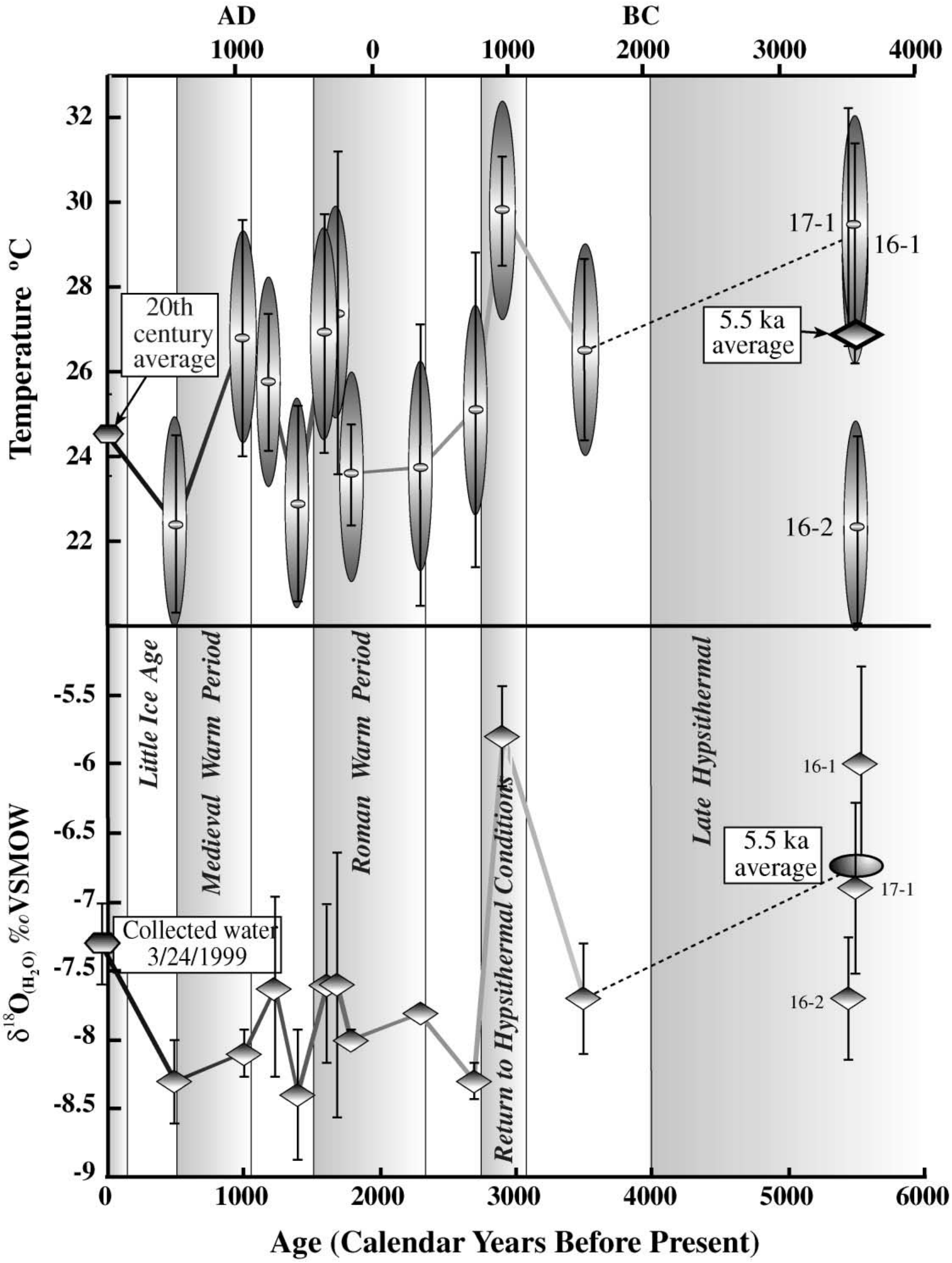

Fig. 9. Estimated range in average summer maximum temperature trends from 5.5 to 0.3–1.0 ka derived from intra-otolith $\delta^{18}O_{(CaCO_3)}$ values of 14 sagittae recovered from the Eastman rockshelter (see text and Table 1). Standard deviation represents variation in estimated temperatures using calculated $\delta^{18}O_{(H_2O)}$ $+1$‰ and is not representative of error. Twentieth-century average maximum temperature is also plotted for comparison. Also plotted is the average $\delta^{18}O_{(H_2O)}$ values from 5.5 to 0.3–1.0 ka derived from intra-otolith $\delta^{18}O_{(CaCO_3)}$ values of 14 sagittae. Standard deviation represents variation in estimated temperatures and is not representative of error. Modern $\delta^{18}O_{(H_2O)}$ value (collected 3/24/99, the approximate start of fish growth) is also plotted for comparison.

40°C exceeds the lethal thermal tolerance of 32.8°C for freshwater drum (Eaton et al., 1995), $\delta^{18}O_{(CaCO_3)}$ values during the 5th summer for 16-1 that decrease approximately 5.5‰ cannot be reflective solely of seasonal temperature change (Fig. 5B). If maximum summer temperature remained approximately 28°C, the average estimated maximum summer temperature from the remaining years, a $\delta^{18}O_{(H_2O)}$ value of −6‰ VSMOW is calculated from the seasonal minimum $\delta^{18}O_{(CaCO_3)}$ value. Because $\delta^{18}O_{(H_2O)}$ value at the beginning of the growing season was calculated at −5‰ VSMOW, this indicates that $\delta^{18}O_{(H_2O)}$ value decreased during the summer. A similar pattern is noted for the specimen sampled from 2.9 ka. During the third summer $\delta^{18}O_{(CaCO_3)}$ values drop 7‰, yielding a calculated maximum summer temperature of 54.5°C (Fig. 6B). If maximum summer temperatures are assumed to be 30°C, the average summer maximum temperature estimated for the other summers from this specimen, $\delta^{18}O_{(H_2O)}$ values decreased from −5.9‰ to approximately −8.5‰ VSMOW during this summer. Finally, the specimen sampled from 1.2 ka yields lower $\delta^{18}O_{(CaCO_3)}$ values at the end of the 3rd summer and at the beginning of the fourth summer (Fig. 8B). These patterns all suggest lower $\delta^{18}O_{(H_2O)}$ values during the mid-to-late summer than at the beginning of the growing season. Several mechanisms can produce the apparent shift in $\delta^{18}O_{(H_2O)}$ values, including delayed snowmelt, migration, and incursion of highly distilled tropical moisture. The seasonal timing of this decrease is critical for proper interpretation.

Snowmelt with characteristically low $\delta^{18}O_{(H_2O)}$ values would begin before temperatures reached 10°C, resulting in lower $\delta^{18}O_{(CaCO_3)}$ values at the beginning of the growing season. This would also yield a relatively low calculated maximum summer temperature and low $\delta^{18}O_{(H_2O)}$ value. This scenario is not supported by our data from which relatively high temperature is calculated. Furthermore, it is counterintuitive that snow-pack should be deeper during the relative warmth of the last 5.5 ka at this latitude.

Another possibility is that the each fish migrated during the summer to water with relatively lower $\delta^{18}O_{(H_2O)}$ values. This would result in a relatively low seasonal minimum $\delta^{18}O_{(CaCO_3)}$ value, which would result in a high summer temperature calculation. However, tributaries must have had a significantly more negative $\delta^{18}O_{(H_2O)}$ values and be of a size that can support these relatively large, non-migratory fish.

We favor a mechanism whereby an increased incursion of highly distilled ^{18}O-depleted tropical moisture decreases $\delta^{18}O_{(H_2O)}$ values. Hurricanes are noted to produce rainfall with characteristically low $\delta^{18}O$ values that could alter surface waters for a significant portion of a season (Lawrence and Gedzelman, 1996; Lawrence, 1998). Therefore, significantly lower seasonal minimum $\delta^{18}O_{(CaCO_3)}$ values may be the result of hurricane-derived precipitation that decreases river water values during the late summer months. In addition, decreased $\delta^{18}O_{(H_2O)}$ values have been reported in Missouri during an El Niño event (Frederickson and Criss, 1999), which would also result in a pattern of $\delta^{18}O_{(CaCO_3)}$ values similar to this anomaly. Therefore, tropical storms or El Niño events are possibly being recorded.

Significantly, these anomalies occur during the warmest periods estimated in this study. Tropical storms, which dissipate low-latitude heat excess, are more likely to occur during particularly warm summers due to an increased energy flux to the tropical Atlantic Ocean. Thus, the climate data we derive for the mid-to-late Holocene suggests an increase in tropical storm incursions during warm periods in agreement with models that estimate a 40–50% increase in hurricanes due to projected future global warming (Emanuel, 1987; Gutowski et al., 1994).

5. Conclusion

We present data on secular variation in estimated maximum summer temperatures and $\delta^{18}O_{(H_2O)}$ values over the past 5.5 ka for the eastern interior United States. Uncertainties include precise dating, limited number of specimens, and quantification of the summer seasonal change in $\delta^{18}O_{(H_2O)}$ values. However, $\delta^{18}O_{(CaCO_3)}$ values from sagittal otoliths support a general decline in maximum summer temperatures from approximately 5.5 to 1.0–0.3 ka, punctuated by warmer climates at 2.9, 1.7–1.6, and 1.2–1.0 ka for the central interior United States. Contemporaneous with the Roman and Medieval Warm Periods of Europe, we find an increase of ~2°C. At 5.5 ka maximum summer temperatures calculated from 2 otoliths were 8°C warmer than maximum summer temperatures calculated from a third. Additionally, there is evidence for a return to mid-Holocene conditions at 2.9 ka. Although we cannot discount these results as an artifact of low population size or sample heterogeneity, it is tentatively suggested as evidence for Holocene climate variation.

Finally, there are three examples where temperatures calculated exceed the lethal thermal tolerance of freshwater drum, requiring that $\delta^{18}O_{(H_2O)}$ values must have decreased during the late summer. Several alternate ideas are presented; however the favored hypothesis is that low $\delta^{18}O_{(H_2O)}$ values during the summer are derived from tropical storms, and that these events correlate with relatively high summer temperatures. These late summer excursions are in agreement with increased mid-latitude storminess during warmer climate periods predicted by some global climate models.

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