



Climatic and human controls on Holocene floodplain vegetation changes in eastern Pennsylvania based on the isotopic composition of soil organic matter

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ARTICLE INFO

Article history:

Received 13 May 2012

Available online 11 April 2013

Keywords:

$\delta^{13}\text{C}$ of soil organic matter

Alluvial deposit

Buried soil

Holocene paleoclimate

Eastern North America

ABSTRACT

A paleoenvironmental time-series spanning the Holocene was constructed using 29 radiocarbon ages and 149 standardized $\delta^{13}\text{C}_{\text{som}}$ values from alluvial terrace profiles along the middle Delaware River valley. There is good agreement between increasing $\delta^{13}\text{C}_{\text{som}}$ and Panicoideae phytolith concentrations, suggesting that variations in C_4 biomass are a major contributor to changes in the soil $\delta^{13}\text{C}$. A measurement error deconvolution curve over time reveals two isotope stages (II-I), with nine sub-stages exhibiting variations in average $\delta^{13}\text{C}_{\text{som}}$ (average $\%C_4$). Stage II, ~10.7–4.3 ka, shows above-average $\delta^{13}\text{C}_{\text{som}}$ (increase $\%C_4$) values with evidence of an early Holocene warming and dry interval (sub-stage IIb, 9.8–8.3 ka) that coincides with rapid warming and cool-dry abrupt climate-change events. Sub-stage IIc, 7.0–4.3 ka, is an above average $\delta^{13}\text{C}_{\text{som}}$ (increase $\%C_4$) interval associated with the mid-Holocene warm-dry hypsithermal. The Stage II-I shift at 4.3 ka documents a transition toward below average $\delta^{13}\text{C}_{\text{som}}$ (decrease $\%C_4$) values and coincides with decreasing insolation and hydroclimatic change. Sub-stages Ib and Id (above average $\%C_4$) coincide with the first documented occurrence of maize in the northeastern USA and a substantial increase in human population during the Late Woodland. These associations suggest that people influenced $\delta^{13}\text{C}_{\text{som}}$ during the late Holocene.

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Introduction

The northeastern United States has a rich history of paleovegetation and paleoclimate investigations largely derived from lithologically based lake-level reconstructions, pollen-based vegetation reconstructions, and stable isotope records (Zhao et al., 2010, and references therein). Recent lake-level reconstructions for the Holocene show evidence of century-long intervals of droughts superposed on a long-term trend towards a wetter climate (Newby et al., 2011; Shuman and Plank, 2011). Despite the large number of lacustrine-based paleoclimate and paleovegetation proxies, there are few allostratigraphic and fluvial geomorphic studies that have had success in reconstructing paleoenvironment through inferred changes in the hydrologic budget (e.g., more or less overall flooding through time) (Scully and Arnold, 1981; Vento et al., 1989; Stewart et al., 1991; Thieme, 2001; Schuldenrein, 2003; Stinchcomb et al., 2012). Although these studies link floodplain and alluvial terrace features to known climate episodes (Wendland and Bryson, 1974), they offer little in the way of semi-quantitative paleovegetation reconstruction.

Floodplain-based studies that utilize the carbon isotopic composition of soil organic matter ($\delta^{13}\text{C}_{\text{som}}$) in the Midwest and southeastern USA show evidence of environmental and climate change impacts and

are therefore a valuable paleoenvironmental archive (Nordt et al., 2007; Driese et al., 2008; Nordt et al., 2008; Kocis, 2011). Furthermore, floodplain archives of paleoenvironmental change are more spatially extensive and therefore harbor the potential for creating high-resolution regional reconstructions of changing vegetation and climate for the past ~11,000 years.

In this paper, we present the first Holocene paleoenvironmental time-series for northeastern USA that uses the carbon isotopic composition of organic matter ($\delta^{13}\text{C}_{\text{som}}$) from buried alluvial soils. The results are discussed in terms of regional climatic, geomorphic, and human controls on the variation in the fractional abundance of C_3 and C_4 biomass, and the accompanying environmental and biological effects on plants.

Background: $\delta^{13}\text{C}$ of soil organic matter

The $\delta^{13}\text{C}$ of soil organic matter ($\delta^{13}\text{C}_{\text{som}}$) reflects below-ground biomass production, which is related to the isotopic composition of the above-ground vegetation. This observation has led to application of $\delta^{13}\text{C}_{\text{som}}$ as a paleoenvironmental and paleoclimate proxy in numerous studies (Cerling et al., 1993; Kelly et al., 1993; Nordt et al., 1994; Boutton et al., 1998; Baker et al., 2000; Huang et al., 2001; Nordt et al., 2002; Runge, 2002; Sedov et al., 2003; Webb et al., 2004; Driese et al., 2005, 2008; Nordt et al., 2008; Beach et al., 2009; Wright et al., 2009; Springer et al., 2010; Beach et al., 2011; Cordova, et al., 2011; Cyr et

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al., 2011; Stinchcomb et al., 2011). The basis for using $\delta^{13}\text{C}_{\text{som}}$ as a paleoenvironmental and paleoclimate proxy is largely established by the (1) disparity between average $\delta^{13}\text{C}$ value for plants with different photosynthetic pathways, and (2) environmental and biological isotope effects acting on or within the plant. These are briefly discussed, in turn, and then followed by a note on the caveats of using the $\delta^{13}\text{C}_{\text{som}}$ method for buried soils.

The logic of using $\delta^{13}\text{C}_{\text{som}}$ as a paleoenvironmental and paleoclimate proxy is largely based on the observation that C_3 plants discriminate against ^{13}C more than C_4 plants due to stomatal constraints imposed during CO_2 diffusion and fixation (Farquhar et al., 1989; Cerling et al., 1993; Boutton, 1996; Ehleringer et al., 2000). This discrimination during photosynthesis leads to unique ranges of $\delta^{13}\text{C}$ values for the two plant types: C_3 between -21% and -35% and C_4 between -10% and -16% (Ehleringer et al., 2000). With their unique range of $\delta^{13}\text{C}$ values, C_3 and C_4 plants are also partly reflected below-ground in the $\delta^{13}\text{C}_{\text{som}}$ values, with soil organic matter turnover processes also playing a role (discussed below). As a result, the $\delta^{13}\text{C}$ value of plants and underlying topsoil organic matter, and the fractional abundance of C_3 to C_4 plants have been correlated with temperature and aridity in a number of studies (Teeri and Stowe, 1976; Ehleringer et al., 1997; Tieszen et al., 1997; Huang et al., 2001; Sage, 2004; Nordt et al., 2007; von Fischer et al., 2008). In addition to the fractional abundance, a number of studies have investigated the $\delta^{13}\text{C}$ variation specific to C_3 and C_4 plants caused by environmental effects (Farquhar et al., 1982; Stewart et al., 1995; Schulze et al., 1996). Various factors, including light, temperature, salinity and moisture availability, can influence C_3 and C_4 photosynthesis and the plant $\delta^{13}\text{C}$ value (Farquhar et al., 1982, 1989; Ehleringer et al., 1993; Stewart et al., 1995; Schulze et al., 1996; Williams and Ehleringer, 1996; Peuke et al., 2006).

Despite that $\delta^{13}\text{C}$ variations are partly driven by climate and environmental changes, a growing body of research shows that some carbon isotopic fractionation occurs during soil organic matter turnover and the resulting $\delta^{13}\text{C}_{\text{som}}$ value may not accurately reflect the overlying vegetation (Nadelhoffer and Fry, 1988; Melillo et al., 1989; Balesdent et al., 1993; Von Fischer and Tieszen, 1995; Garten et al., 2000; Dijkstra et al., 2006; Millard et al., 2010). The process of organic matter turnover, especially respiration through soil microbes, can result in a less negative $\delta^{13}\text{C}_{\text{som}}$ value relative to litter and root inputs, which limits the ability of using $\delta^{13}\text{C}_{\text{som}}$ as a paleoclimate proxy. Although soil respiration can result in increasing $\delta^{13}\text{C}_{\text{som}}$ with depth, studies have also shown that the disequilibrium $\delta^{13}\text{C}_{\text{som}}$ at the surface and with depth may be related to recent changes in the relative abundance of C_3 and C_4 vegetation (Dzurec et al., 1985; Boutton et al., 1998). The inability to deconvolve factors influencing $\delta^{13}\text{C}_{\text{som}}$ values suggests that an alternative means of C_4 plant identification is required. We use a phytolith analysis of alluvial soil horizons in this study to independently identify the presence of C_4 plants.

Setting

The Delaware River basin, located in the eastern United States, drains 33,041 km² into the Atlantic Ocean (Fig. 1). The Delaware River forms at the confluence of the West and East branch at Hancock, NY. From this point, the river traverses 300 km across four physiographic provinces to the head of tide located in Trenton, NJ. Upstream from the Delaware Water Gap, the river drains 9971 km² of partly confined and dissected terrain within the Appalachian Ridge and Valley and the Plateau provinces. The main channel within the study area flows through the Ridge and Valley province. This region consists of sinuous, alternating ridges and valleys resulting from differential erosion of the underlying folded and faulted succession of Paleozoic sedimentary rocks (Way, 1999).

The channel itself is classified as a low sinuosity planform-controlled with discontinuous floodplain (Stinchcomb et al., 2012). The bedrock and late Pleistocene periglacial deposits act as confining features for

the channel. Landforms within the study area range from periglacial landforms, alluvial fill terraces, floodplains, islands, and gravel bars. Although many alluvial terraces document overbank flooding that spans the Holocene, there is evidence that documents floodplain and terrace reworking for the past 6000 years (Stinchcomb et al., 2012).

The modern climate for the study area is classified as humid, continental with warm summers (Trewartha, 1957) and a mean annual temperature of $\sim 9.7^\circ\text{C}$ and mean annual precipitation range of 1290 mm y^{-1} (NOAA, 2011). Winter and spring are peak flooding seasons for the study area, with flooding often caused by heavy rainfall resulting from extratropical cyclones combined with snowmelt, land-falling tropical cyclones, or warm-season convective systems (Hirschboeck, 1988; Smith et al., 2010).

Upland and slope vegetation consists of mixed oak-hickory (*Quercus-Carya*) (formerly chestnut, *Castanea*) forests (Braun, 1950; Kuchler, 1964). Steep ravines that flank the Delaware River valley contain old-growth hemlock (*Tsuga*) forests. River valley communities vary from closed canopy, open canopy, and non-forested regions. Closed canopy constitutes two-thirds of a portion of the middle Delaware River valley, i.e., the Delaware Water Gap Recreation Area (Perles et al., 2007). Closed-canopy communities consist mostly of silver maple (*Acer saccharinum*) floodplain forests. Sycamore forests (*Platanus occidentalis*), mixed with maples (*Acer* spp.), ashes (*Fraxinus* spp.), and river birch (*Betula nigra*), are also common. Herbaceous and grassland plant communities are also common along the Delaware River valley floodplain, occurring along islands, shorelines, gravel bars, and within riverbeds (Perles et al., 2007). A big bluestem (*Andropogon gerardii*) and indiagrass (*Sorghastrum nutans*) riverine grassland association has been documented throughout the study area (Perles et al., 2007). This vegetation community also includes switchgrass (*Panicum virgatum*) and little bluestem (*Schizachyrium scoparium*). Unlike the C_3 forested river valley corridors, this grassland community predominantly utilizes the C_4 photosynthetic pathway.

Methods

Soil description and sample collection

Riverbank exposures and previous geoarchaeological excavations along alluvial terraces ($n = 6$), combined with 29 radiocarbon ages, were used to describe and characterize the surface and buried soils that formed during the Holocene along the middle Delaware River Valley. The soil morphology of bank exposure profiles and excavations was described using U.S. Department of Agriculture (USDA) soil nomenclature (Schoeneberger et al., 2002) with modifications (see Holliday, 2004, Appendix 1). Buried soils ("b" horizon designation) were designated based on the presence of material overlying a former soil, regardless of the deposit thickness (Schaetzl and Anderson, 2005). Bulk soil and sediment samples were collected stratigraphically from trenches at 2-, 5-, or 10-cm intervals. The appropriate sampling interval was chosen based on the thickness of soil horizons or alluvial strata so that transitions were adequately sampled and potential vegetation changes recorded in cumelic soils were captured. Oriented samples were collected for soil micromorphological analysis to help determine the degree of bioturbation within the buried soils. Samples were commercially prepared and thin-sections were described using terminology following Stoops (2003).

Carbon isotope and phytolith analysis of bulk soil

Stable isotope analysis of the soil organic matter and phytolith analysis were conducted to first determine the relative proportions of C_3 and C_4 biomass contributions to buried soils. The carbon isotopic composition of soil organic matter ($\delta^{13}\text{C}_{\text{som}}$) was measured for 214 bulk soil and sediment samples collected from Delaware River valley

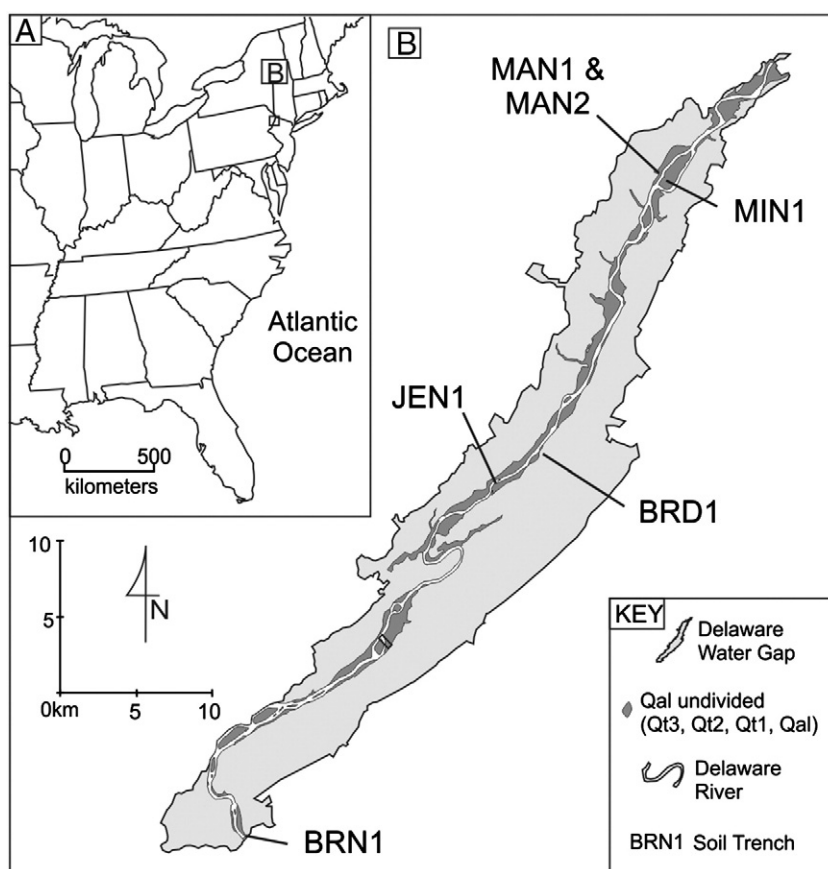


Figure 1. Study area location. (A) Map of eastern North America showing the location of (B) middle Delaware River valley study area.

floodplain profiles in northeastern USA (Fig. 1). Samples were collected in the field with extra care to avoid any cultural or natural disturbance feature (e.g., hearth, pit, large roots); however few of these features were encountered. Samples were air-dried in the laboratory and visible plant remains were hand-picked and removed (e.g., roots, leaf litter fragments) to ensure the organic matter within samples was primarily composed of humified, organomineral-complexed organic matter. Samples were then powdered in mortar-and-pestle and dried in an oven at 60°C, overnight. No carbonate minerals were detected in the field, and therefore, the samples were not treated with HCl or H₂SO₃ prior to combustion. Despite this, additional steps were taken to ensure the absence of carbonate minerals. Two aliquots from each powdered sample were treated with 1 N HCl and observed under binocular microscope to detect the potential presence of trace carbonate minerals. No chemical reaction was observed in any sample and no carbonate minerals were detected in any soil or sediment thin-sections (Fig. 2). Nine paired samples were treated and untreated with H₂SO₃, respectively to further determine if any carbonate minerals were present. The results of this analysis show a near 1:1 correlation between treated and untreated samples, where the regression slope equals 0.9973, R² = 0.96 (Supplemental Fig. 1). Organic matter in the samples was combusted in a Costech elemental analyzer and the resulting CO₂ was analyzed by continuous flow using a Finnigan Thermo-Electron gas-source Mass Spectrometer at Baylor University's stable isotope laboratory (standard error is ≤ 0.04‰). Values are reported here in ‰, with reference to the Pee Dee Belemnite (VPDB) standard, using the following equation:

$$\delta^{13}\text{C}_{\text{som}}(\text{‰}) = \left\{ \left[\left(\frac{^{13}\text{C}_{\text{sample}}}{^{12}\text{C}_{\text{sample}}} \right) / \left(\frac{^{13}\text{C}_{\text{VPDB}}}{^{12}\text{C}_{\text{VPDB}}} \right) \right] - 1 \right\} \times 1000.$$

With these $\delta^{13}\text{C}_{\text{som}}$ values, we estimated the relative contributions of C₃ and C₄ biomass to the soil organic matter pool using a two-end member mixing model:

$$\delta^{13}\text{C}_{\text{som}} = \left[\delta^{13}\text{C}_{\text{C}_4}(x) + \delta^{13}\text{C}_{\text{C}_3}(x-1) \right] * 100,$$

where $\delta^{13}\text{C}_{\text{C}_4}$ is the average carbon isotope value for pure C₄ plant sources, −13‰, $\delta^{13}\text{C}_{\text{C}_3}$ is the average carbon isotope value for a pure C₃ plant sources, −27‰, and x is the relative contribution of C₄ biomass (Boutton et al., 1998). The contribution of C₄ biomass should be regarded as an estimate in this study because of uncertainties in $\delta^{13}\text{C}_{\text{C}_4}$ and $\delta^{13}\text{C}_{\text{C}_3}$ end-member values, variation in the $\delta^{13}\text{C}$ of atmospheric CO₂, and the differential decomposition of organic matter and the resulting enrichment of $\delta^{13}\text{C}$ (Nordt et al., 2002).

Ten soil samples from the JEN1, BRD1, and MAN2 trenches were subjected to a phytolith analysis to determine the presence of C₄ vegetation (Fig. 1). Short-cell grass phytoliths (i.e., Chloridoideae and Panicoideae subfamilies) were analyzed following the methods of Piperno (2006). Soils were first deflocculated and wet sieved through 250- and 53- μm sieves to separate sands from coarse and fine silts. Clays were discarded via gravity sedimentation. Five grams of fine silt was then placed in a test tube and potential carbonates were removed using 10% hydrochloric acid solution (HCl). A 10% solution of potassium hydroxide (KOH) was added for 10 min to breakdown humic compounds before adding concentrated nitric acid (HNO₃) to remove organics. Tubes were heated in a boiling bath to expedite this process. Once complete, phytoliths were separated out using a heavy liquid solution (specific gravity = 2.3) composed of cadmium iodide and potassium iodide (CdI₂/KI). Samples were then rinsed and dried with acetone prior to mounting. Microscope slides were prepared using permount mounting medium. Scanning was carried

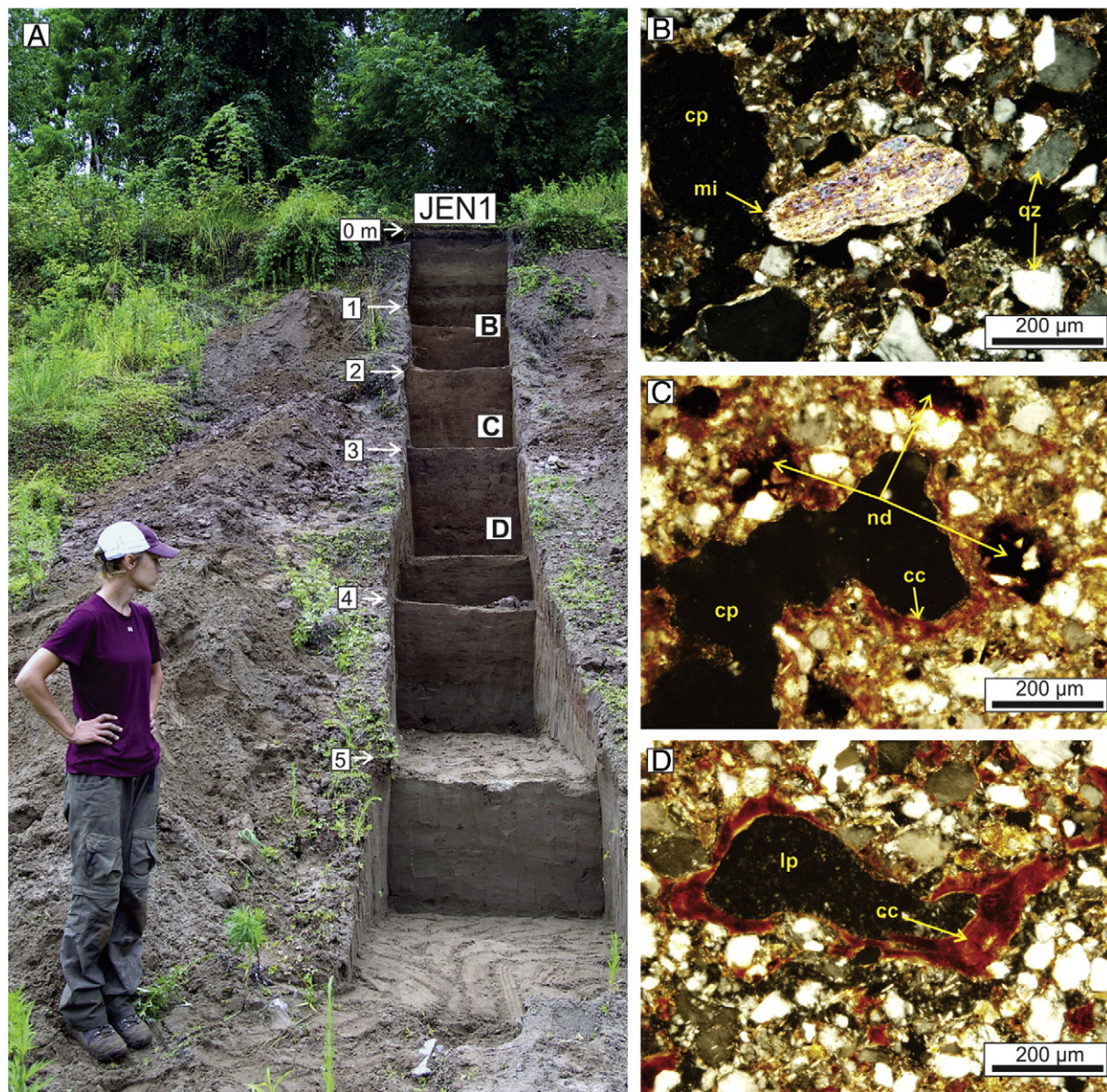


Figure 2. Example of buried soils along the middle Delaware River valley. (A) Photograph of the JEN1 soil trench along alluvial terrace showing multiple buried soils (see Fig. 3) and location of photomicrographs. (B) Photomicrograph of late Holocene buried topsoil (Ab4; 0.76 m below surface) showing hydrated mica grain (mi), weak chitonic microfabric, intrapedal channel pore (cp), and quartz (qz) skeletal fragments and undifferentiated b-fabric. (C) Photomicrograph of middle Holocene buried subsoil (Bwb9, 2.76 m below surface) showing a intrapedal channel pore with reddish brown clay coatings (cc) lining pore walls and Fe-Mn nodules (nd) in a grano- and poro-striated b-fabric. (D) Photomicrograph of early Holocene buried subsoil (Bt1b10, 3.56 m below surface) showing a lenticular pore (lp) with relatively continuous reddish brown clay coatings in a grano- and poro-striated b-fabric.

out along transects at 200 \times prior to permount setting. All short-cell phytoliths were studied in 3D at 400 \times . The results are presented as percent subfamily based on a population of 300 phytoliths counted on each slide. Short cell grass phytolith identifications utilized established protocols (Twiss et al., 1969; Fredlund and Tieszen, 1997; Lu and Liu, 2003).

Construction of carbon isotope and C_4 biomass Holocene time-series

A Holocene time-series of $\delta^{13}C_{\text{som}}$ for the Delaware River valley study area was constructed using (i) age-depth modeling, (ii) data standardization, and (iii) a deconvolution kernel estimation algorithm which adjusts for measurement error in age. Age-depth models were calculated using reported radiocarbon ages and the Classic non-Bayesian Age-Depth Modeling (CLAM) program (Blaauw, 2010) in the statistical software, R. Using the ages and their associated depths, CLAM calibrates radiocarbon ages using the IntCal09 calibration curve

(Reimer et al., 2009) and then calculates the weighted mean age for every 1-cm increment derived from 1000 Monte Carlo age-depth iterations. For this study, linear interpolation between ages was used, and no hiatuses were introduced into the models because of the limited number of radiocarbon ages available from each trench.

Changes in landscape position relative to the channel (i.e., pedofacies, *sensu* Bown and Kraus, 1987; Kraus and Bown, 1988) can have an influence on the type and frequency of vegetation, which can influence the $\delta^{13}C_{\text{som}}$ value. Because the profiles examined occur in different geomorphic settings (Stinchcomb et al., 2012), the $\delta^{13}C_{\text{som}}$ values from buried soils for each trench were standardized to rescale and center the data, making the trenches comparable with each other. For each trench the mean and standard deviation of $\delta^{13}C_{\text{som}}$ were calculated. The standardized value is computed by first taking the difference of the original value and the trench mean. This difference is then divided by the trench standard deviation. This allows us to treat all trenches as homogenous populations, so that we can investigate deviations for $\delta^{13}C_{\text{som}}$ over

time for all trenches on the same scale. Similar calculations were performed on %C₄. The resulting standardized values are reported as $\Delta\delta^{13}\text{C}_{\text{som}}$ and $\Delta\%C_4$. Variations in $\Delta\delta^{13}\text{C}_{\text{som}}$ are inferred here to reflect above- and below-average C₃:C₄ changes that occurred independent of lateral variation in geomorphic setting. The standardization approach is better suited for examining coherent changes in $\delta^{13}\text{C}_{\text{som}}$ because it minimizes the effect lateral variations in closed and open canopy due to geomorphic setting can have on the data. However, this approach does not minimize the effects of widespread geomorphic change through time on the $\Delta\delta^{13}\text{C}_{\text{som}}$ time series (e.g., incision), and thus limits the paleoclimate interpretation.

The soil $\Delta\delta^{13}\text{C}_{\text{som}}$ and $\Delta\%C_4$ from all trenches were aggregated into a single group to produce a time series of estimated average C₃:C₄ biomass contribution. A number of factors can influence the duration of soil formation, which affects the outcome of the modeled age, introducing uncertainty into the estimated age of individual $\Delta\delta^{13}\text{C}_{\text{som}}$ and $\Delta\%C_4$ values. By definition, soils are time-averaged weathering features that mix and homogenize material that formed at different intervals (Holliday, 2004). Alluvial soils that weather for extended periods can result in a condensed version of $\Delta\delta^{13}\text{C}_{\text{som}}$ and $\Delta\%C_4$ through time. Conversely, alluvial soils can accumulate sediment at rates that keep pace with pedogenesis (i.e., cumulation). This style of developmental up-building would potentially capture changes in $\Delta\delta^{13}\text{C}_{\text{som}}$ and $\Delta\%C_4$ values through time as the soil accumulates new sediment and slowly buries older soil that is no longer receiving organic matter contributions. Finally, flooding that contributes sufficient sediment can bury a soil abruptly, shutting off soil formation instantaneously. The potential for floodplain erosion would introduce time gaps in the age-modeled $\Delta\delta^{13}\text{C}_{\text{som}}$ and $\Delta\%C_4$ time series.

Estimating error in the variable time is not a straightforward process. The continuum of alluvial soil formation and burial (prolonged soil formation – cumulation – abrupt burial) introduces an element of uncertainty in the modeled age for buried soils, and therefore the $\delta^{13}\text{C}_{\text{som}}$ and %C₄ for a given time during the Holocene. The measurement error inherent in the regressor, time, was addressed using deconvolution kernel estimation to better estimate the age of the $\delta^{13}\text{C}_{\text{som}}$ and %C₄ values (Carroll et al., 2006; Wang and Wang, 2011). The deconvolution statistical method involves isolating a single variate (i.e., signal) of a mixture model (i.e., signal, plus noise). Since the true age (signal) of our samples is unknown, we rely on a surrogate measure, depth. From this surrogate we obtain an estimated age of the $\delta^{13}\text{C}_{\text{som}}$ sample using CLAM. We also used the age error output from CLAM in our deconvolution model. To see how sensitive our results are to this error specification, we then looked at a constant error of 200 years (200 being slightly larger than the average CLAM error). Both measurement error models were graphically similar. The measurement error model using the error values computed from CLAM was chosen for this study to avoid choosing an arbitrary error structure. Isotope stages and sub-stages were identified based on visual inspection of isotope discontinuities (Nordt et al., 2008).

Results and discussion

The six trenches along alluvial terrace landforms show a predominance of coarse-loamy multistory buried soils weathered in overbank deposits composed of mixed alluvium with quartz as the primary component (Fig. 2) (Stinchcomb et al., 2012). The number of buried soils observed in each profile suggests that the site areas are highly aggradational with occasional episodes of floodplain surface scouring (see Stinchcomb et al., 2012). Calibrated radiocarbon ages (median probability age reported in kilo-annums, ka) and age-depth modeling from this study and previous research show that trenches span various intervals of the Holocene (Figs. 3 and 4). Buried soils range in degree of development and are mostly inceptisolic-like with thin A horizons (where present) and Bw and/or Bt subsoil horizons that show varying degrees of clay illuviation (Fig. 2). Macro- and micromorphological

investigations of surface and buried soils showed no evidence of primary or pedogenic calcium carbonate. Additionally, no major pedoturbation features were detected in thin section. The soil micromorphology results provide support that the soil organic matter and $\delta^{13}\text{C}_{\text{som}}$ reflect vegetation at successive land surfaces with minimal physical mixing.

The results from the isotope analysis of the six trenches show a mean (\pm SD) $\delta^{13}\text{C}_{\text{som}}$ value of -23.8% (± 1.6) for the entire sample (Fig. 3, Table 2). The isotope values range from -26.7% to -17.7% . A Wilcoxon rank-sum test comparing A and B horizons show no significant difference between the median $\delta^{13}\text{C}_{\text{som}}$ for the two groups ($p = 0.55$). The results of the phytolith analysis show that Panicoideae and Chloridoideae (C₄ and C₃) phytoliths are present in the soils (Table 3 and Fig. 5). After removing an outlier associated with a cultural rock-pit feature, the % Panicoideae phytoliths correlate significantly with their corresponding $\delta^{13}\text{C}_{\text{som}}$ values (Fig. 6; $p = 0.0003$). The data were subsequently cross-validated to determine the significance of phytolith data had on the overall regression model. A y-intercept value of -23.9% and model slope of 0.71 was observed when the three Panicoideae values of 0% were removed from the plot. The above y-intercept and slope were not significantly different from the regression equation that includes the three 0% observations. It is important to note that the regression slope is primarily driven by the two higher percentage phytolith observations and thus more data are needed to further evaluate the model.

The presence of C₄ phytoliths and their relationship with $\delta^{13}\text{C}_{\text{som}}$ data suggest that (1) C₄ grasses were present along the Delaware River valley in the past and, (2) there is a positive correlation between Panicoideae phytolith concentrations and $\delta^{13}\text{C}_{\text{som}}$ values. Furthermore, $\delta^{13}\text{C}_{\text{som}}$ values greater than -23.9% suggest that C₄ and C₃ grasses in the Panicoideae subfamily are beginning to influence the isotope signal. Because the Panicoideae subfamily also contains C₃ grasses, it is unclear whether the strong linear correlation is strictly a direct relationship between C₄ plants and $\delta^{13}\text{C}_{\text{som}}$ values or combined increase in C₄ and water-stressed C₃ grasses. Despite the possibility of environmental and biological effects on the $\delta^{13}\text{C}_{\text{som}}$ values, these data provide supporting evidence that $\delta^{13}\text{C}_{\text{som}}$ values for our study area reflect the fractional abundance of C₄ and C₃ grasses.

Inter-profile comparisons show that the mean and standard deviation of $\delta^{13}\text{C}_{\text{som}}$ values can vary significantly by trench even during similar time periods (Table 2; Fig. 7). These data suggest that position on the alluvial landscape (e.g., levee, floodbasin, paleochannel fill) can have a notable effect on the $\delta^{13}\text{C}_{\text{som}}$ value and the riparian plant community. Previous research has shown that variations in modern riparian plant communities vary spatially as a function of fluvial geomorphic processes (Hupp and Osterkamp, 1996). This fluvial geomorphic control on plant communities has been observed along the modern middle Delaware River valley (Perles et al., 2007), and can be inferred for much of the Holocene using a Delaware River floodplain and alluvial terrace development model (Stinchcomb et al., 2012). The time-series of dissimilar $\delta^{13}\text{C}_{\text{som}}$ values suggests that temporally equivalent plant communities can vary in the fractional abundance of C₃ and C₄ plants, which is likely due to position on the landscape and the influence of varying fluvial geomorphic processes (Fig. 7; upper graph).

The $\delta^{13}\text{C}_{\text{som}}$ values, when standardized, show combined plant community variations as they deviate from their respective averages (Fig. 7; lower graph). Furthermore, coherent above- and below-average changes in $\delta^{13}\text{C}_{\text{som}}$ (i.e., $\Delta\delta^{13}\text{C}_{\text{som}}$ values) are observed during intervals that also show heterogeneous plant communities and $\delta^{13}\text{C}_{\text{som}}$ values. These coherent changes provide evidence that the standardization process negates the effects of spatial variation on plant communities as a function of fluvial geomorphic processes. The variations in $\Delta\delta^{13}\text{C}_{\text{som}}$ reflect potential environmental and biological effects on C₃ and C₄ plant communities (e.g., moisture stress). Therefore, the results are discussed in terms of both paleotemperature and paleo-moisture variations.

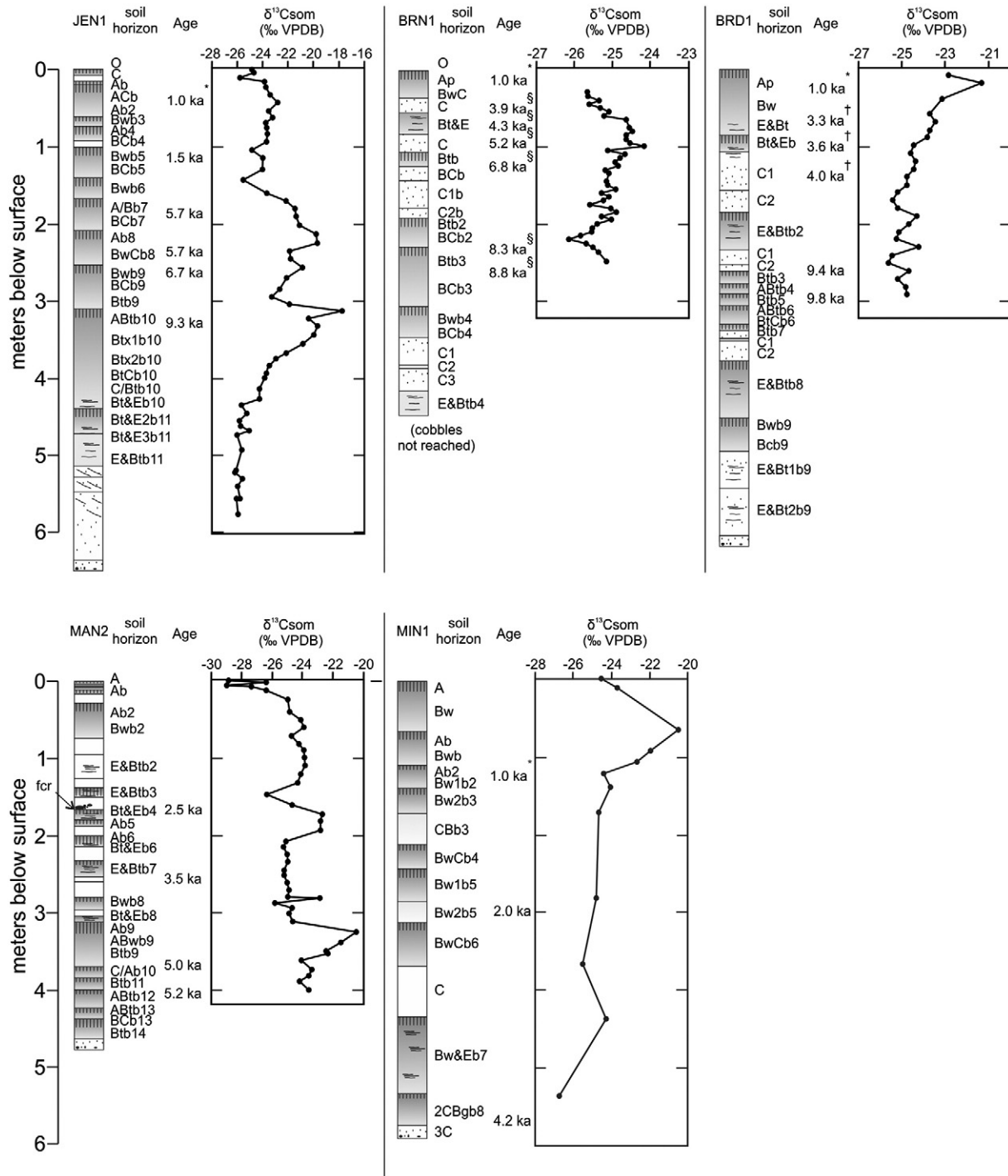


Figure 3. Alluvial terrace soil profiles, calibrated radiocarbon ages (reported in kilo-annums, cal ka BP), and $\delta^{13}\text{C}_{\text{som}}$ data. See Table 1 for radiocarbon age details. Note: Man1 trench (Stinchcomb et al., 2011) not shown here.

Isotope stages and environmental implications

The deconvolution measurement error model results of standardized values ($\Delta\delta^{13}\text{C}_{\text{som}}$ and $\Delta\%C_4$) from the six combined trenches show two major isotope stages and nine sub-stages encompassing the past 10,700 years (Fig. 7). The 2 isotope stages and 9 sub-stages are intervals of time marked by notable deviations from the average $\delta^{13}\text{C}_{\text{som}}$ and $\%C_4$ values and suggest noteworthy changes in plant communities throughout the study area. Because the evolution of riparian plant communities on time scales ranging from 10^2 to 10^3 years

varies as a function of climate change (Richards et al., 2002), we compare these stages to previously published paleoclimate records.

The stage II-I shift in values is apparent in the raw (un-standardized) and standardized $\delta^{13}\text{C}_{\text{som}}$ values (Fig. 8). A one-sided *t*-test shows that the stage II mean is significantly greater than the stage I mean at the 5% significance level ($p = 0.027$). Compared to Stage II (10.7–4.3 cal ka BP), Stage I (4.3–0 cal ka BP) has primarily positive $\Delta\delta^{13}\text{C}_{\text{som}}$ ($\Delta\%C_4$) values that indicate above-average $\delta^{13}\text{C}_{\text{som}}$ ($\%C_4$). The significantly positive values from Stage II suggest a time of overall greater $\%C_4$ and therefore higher growing season temperatures and potentially less

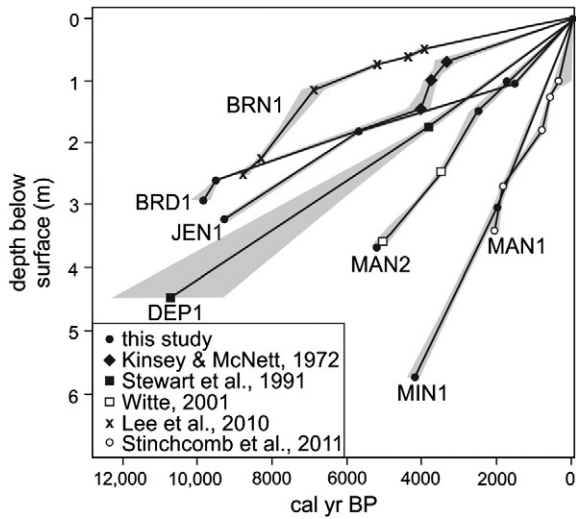


Figure 4. Age-depth model results for alluvial terrace profiles used in this study.

available moisture (depending on the fraction of water-stressed C₃ plants present). The 4.3 ka shift to overall negative Δδ¹³C_{som} (Δ%C₄) values during Stage I indicate overall less %C₄ and/or moisture-stressed C₃ grasses, suggesting a change towards decreasing growing-season temperatures and/or increased available moisture. Midwest δ¹³C_{som} data from buried A horizons also show a modeled decrease in the mean July (growing season) temperature (Nordt et al., 2008), which coincides with the Delaware isotope stage II–I shift. Like Nordt et al. (2008), we partly attribute this isotope shift as a vegetation response to decreasing Northern Hemisphere insolation (Kutzbach, 1981) during the Neoglacial period (Denton and Karlen, 1973; Wendland and Bryson, 1974). As a result, the riparian canopy may

Table 2
Descriptive statistics for δ¹³C_{som} data.

Sample id	No. of samples	Mean (‰)	Std. dev. (‰)	Min (‰)	Max (‰)
MIN1	10	−23.8	± 1.7	−26.7	−20.5
MAN2	30	−23.9	± 1.3	−26.3	−20.4
MAN1 ^a	33	−23.8	± 1.3	−25.4	−20.3
BRD1	17	−24.1	± 1.0	−25.3	−21.3
JEN1	38	−22.5	± 1.8	−25.8	−17.7
BRN1	21	−25.3	± 0.4	−26.1	−24.6
All C horizons	45	−24.8	± 0.6	−26.3	−22.8
Total population	194	−23.8	± 1.6	−26.7	−17.7

^a Data from Stinchcomb et al. (2011).

have become more closed, i.e., increase in C₃ plants, thus decreasing C₄ plant populations. A comparison with regional and continental-scale paleoclimate records shows that the 4.3 cal ka BP shift in the Δδ¹³C_{som} (Δ%C₄) values following the middle Holocene climatic optimum (Viau et al., 2002), as documented by widespread changes in pollen assemblages (Wendland and Bryson, 1974; Bryant and Holloway, 1985) and a North Atlantic ice-rafting debris event, Bond event-3 (Bond et al., 1997; Bond et al., 2001). Other pollen-based temperature reconstructions from northeastern USA and Northern Quebec show decreasing temperatures much later around 3.2 ka (Shuman et al., 2004; Viau et al., 2006; Shuman et al., 2009).

The 4.3 cal ka BP Stage II–I isotope shift also reflects a decrease in water-stressed C₃ plants and agrees with a decrease in moisture deficit documented in ¹³C_{som} alluvium in West Virginia and southeastern Tennessee (Driese et al., 2008; Springer et al., 2010; Kocis, 2011) (Fig. 8). Oxygen isotope records from calcite in northeastern lakes and West Virginia speleothems show that the decreasing insolation resulted in hydroclimatic change and a transition from summer dominated to seasonally distributed precipitation (Shuman and

Table 1
Radiocarbon data from previous research used in this study.

Radiocarbon lab code	Technique	Location ^a	Sample description	Conventional 14C age ^b	Calibrated median age (cal ka BP) ^c	Source
Beta-227477	AMS	MAN1	Maize kernel	270 ± 40	337 (0.3)	Stinchcomb et al. (2011)
Beta-227480	AMS	MAN1	Maize kernel	390 ± 40	448 (0.4)	Stinchcomb et al. (2011)
Beta-227479	AMS	MAN1	Maize kernel	550 ± 40	562 (0.6)	Stinchcomb et al. (2011)
Beta-227482	Radiometric	MAN1	Charcoal from cultural feature	840 ± 70	768 (0.8)	Stinchcomb et al. (2011)
Beta-258885	AMS	MAN1	Bulk organic fraction	1900 ± 40	1845 (1.8)	Stinchcomb et al. (2011)
Beta-280873	AMS	MIN1	Charcoal from matrix	2020 ± 40	1972 (2.0)	Stinchcomb et al. (2012)
Beta-257433	AMS	MAN1	Charcoal from matrix	2070 ± 40	2040 (2.0)	Stinchcomb et al. (2011)
Beta-280271	AMS	MAN2	Charcoal from cultural feature	2420 ± 40	2459 (2.5)	Stinchcomb et al. (2012)
Y-2339	Radiometric	BRD1	Charcoal from cultural feature	3120 ± 120	3322 (3.3)	Kinsey and McNett (1972)
GX-28163	AMS	MAN2	Charcoal from matrix	3230 ± 40	3448 (3.4)	Witte (2001)
Y-2340	Radiometric	BRD1	Charcoal from cultural feature	3390 ± 100	3643 (3.6)	Kinsey and McNett (1972)
Y-2341	Radiometric	BRD1	Charcoal from cultural feature	3570 ± 100	3871 (3.9)	Kinsey and McNett (1972)
Beta-262909	AMS	BRN1	Charcoal from cultural feature	3600 ± 40	3908 (3.9)	Lee et al. (2010)
Y-2342	Radiometric	BRD1	Charcoal from cultural feature	3660 ± 120	3998 (4.0)	Kinsey and McNett (1972)
Beta-280874	AMS	MIN1	Charred plant material	3810 ± 40	4203 (4.2)	Stinchcomb et al. (2012)
Beta-266915	AMS	BRN1	Charcoal from matrix	3900 ± 40	4335 (4.3)	Lee et al. (2010)
Beta-280872	AMS	MAN2	Charcoal from matrix	4410 ± 40	4988 (5.0)	Stinchcomb et al. (2012)
GX-28162	AMS	MAN2	Charcoal from matrix	4500 ± 40	5163 (5.2)	Witte (2001)
Beta-266910	AMS	BRN1	Charcoal from cultural feature	4510 ± 40	5143 (5.1)	Lee et al. (2010)
Beta-289229	AMS	JEN1	Bulk organic fraction	4930 ± 40	5653 (5.7)	Stinchcomb et al. (2012)
Beta-280270	AMS	JEN1	Charcoal from matrix	4940 ± 40	5664 (5.7)	Stinchcomb et al. (2012)
Beta-296721	AMS	JEN1	Bulk organic fraction	5980 ± 40	6819 (6.8)	Stinchcomb et al. (2012)
Beta-266914	Radiometric	BRN1	Charcoal from matrix	5980 ± 110	6828 (6.8)	Lee et al. (2010)
Beta-266913	AMS	BRN1	Charcoal from matrix	7420 ± 50	8257 (8.3)	Lee et al. (2010)
Beta-266912	AMS	BRN1	Charcoal from matrix	7920 ± 50	8760 (8.8)	Lee et al. (2010)
Beta-282837	AMS	JEN1	Bulk organic fraction	8280 ± 40	9287 (9.3)	Stinchcomb et al. (2012)
Beta-280267	AMS	BRD1	Charcoal from matrix	8410 ± 40	9449 (9.4)	Stinchcomb et al. (2012)
Beta-280268	AMS	BRD1	Charcoal from matrix	8800 ± 50	9834 (9.8)	Stinchcomb et al. (2012)
Uga-5488	–	36MR45	Charcoal from cultural feature	9330 ± 545	10629 (10.6)	Stewart et al. (1991)

Ages in parentheses are rounded to nearest century and expressed as kilo-annums (ka).

^a See Figures 1 and 3 for location.

^b Conventional radiocarbon age with standard error. The age is expressed in years before present (BP), 1950.

^c Median probability age calculated in CALIB 6.0.1, using IntCal09 calibration curve (Reimer et al., 2009).

Table 3
Phytolith and $\delta^{13}\text{C}_{\text{som}}$ data from select soil samples.

Sample id	Horizon	Panicoideae (% of total)	Chloridoideae (% of total)	Particulate charcoal (% of total)	$\delta^{13}\text{C}_{\text{som}}$
BRD1 – 70	E&Bt	1.65	0	3.2	–23.4
BRD1 – 90	Bt&Eb	0	0	0	–23.8
BRD1 – 310	BtC	0	0	0	–25.4
MAN2 – 173	Bt&Eb	11	0	45	–22.6
MAN2 – 338	Ab9	2	4	20	–21.4
JEN1 – 55	BCb2	0.53	0.53	3.6	–23.5
JEN1 – 130	BCb5	0.58	0	0.57	–24.0
JEN1 – 207	Ab8	5	0.6	66	–19.8
JEN1 – 285	BCb9	0	0.75	3.6	–22.6
JEN1 – 313	ABtb11	9.3	1	20.5	–17.7

Donnelly, 2006; Hardt et al., 2010; Zhao et al., 2010). The decline in temperature and seasonal contrast may have resulted in a moisture surplus and less water-induced stress on C_3 and drought-sensitive C_4 nicotinamide adenine dinucleotide phosphate – malic enzyme (NADP-ME) panicoid grasses (Ripley et al., 2010). An increasing frequency of large flood events in the NE USA may also indicate increased moisture, especially early spring thaws and snowmelts (Parris et al., 2010) (Fig. 9).

The notably different populations evident in the raw $\delta^{13}\text{C}_{\text{som}}$ data from Stage II suggest that $\delta^{13}\text{C}_{\text{som}}$ values reflect two unique floodplain vegetation environments: open or semi-open canopy where C_4 grasses are better adapted (–22‰ to –20‰) and closed-canopy riparian forest (–26‰ to –24‰) consisting primarily of C_3 plants. This disparity in vegetation associations may be due to variations in fluvial geomorphology. If this is true, then the shift in vegetation at 4.3 cal ka BP may, in fact, be a biotic response to middle Holocene incision (Stinchcomb et al., 2012). Specifically, channel degradation from approximately 6 to 5 cal ka BP resulted in alluvial terrace formation and a decreased incidence of flooding. The decreased frequency of floods may have led to stable alluvial landscapes that could support closed-canopy C_3 forests from 4.3 cal ka BP to modern.

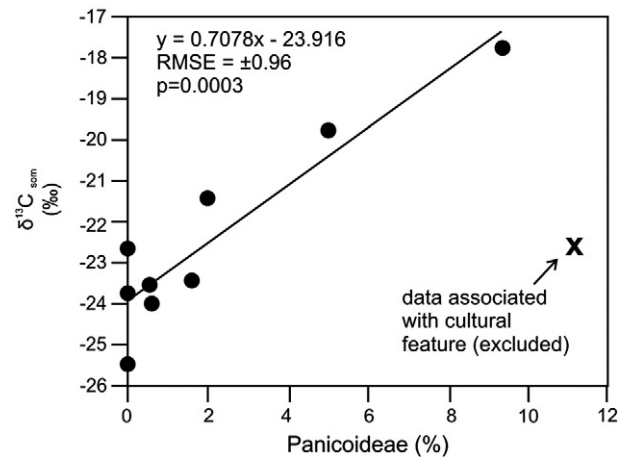


Figure 6. Results of linear regression of $\delta^{13}\text{C}_{\text{som}}$ by Panicoideae (% of total count). RMSE is root mean square error. Panicoideae concentration (% of total) determined from 300 total counts. Although the observed correlation is strong, the regression line is based on few data and the reported function should not be used for predictive purposes.

Isotope sub-stages and environmental implications

Stage II (10.7–4.3 cal ka BP) was divided into 4 sub-stages (a–d) that span the early and middle Holocene (Figs. 7–9). Sub-stage IIa (~10.6–9.9 cal ka BP) is poorly constrained, having 5 $\delta^{13}\text{C}_{\text{som}}$ observations from a single trench (JEN1). Given this, the measurement error model yields negative $\Delta\delta^{13}\text{C}_{\text{som}}$ ($\Delta\% \text{C}_4$) values, indicating below-average $\delta^{13}\text{C}_{\text{som}}$ ($\% \text{C}_4$) values during this time period. The sub-stage IIa interval of below-average $\delta^{13}\text{C}_{\text{som}}$ ($\% \text{C}_4$) suggests lower growing season temperatures and/or a moisture surplus. This isotope-inferred interval of cooler summer temperatures coincides with pollen-based temperature reconstructions for both Northern Quebec and northeastern USA that also

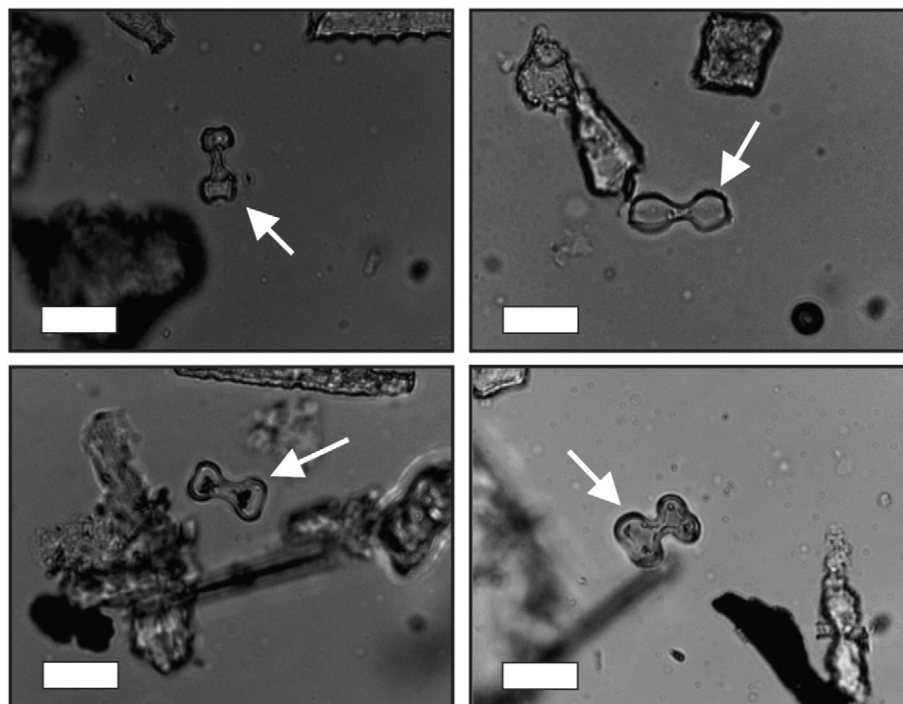


Figure 5. Photomicrographs of short cell grass phytoliths (noted with arrow) of the subfamily Panicoideae isolated from early Holocene soil (ABtb11 horizon, 3.13 m below the surface). White bar length = 20 μm .

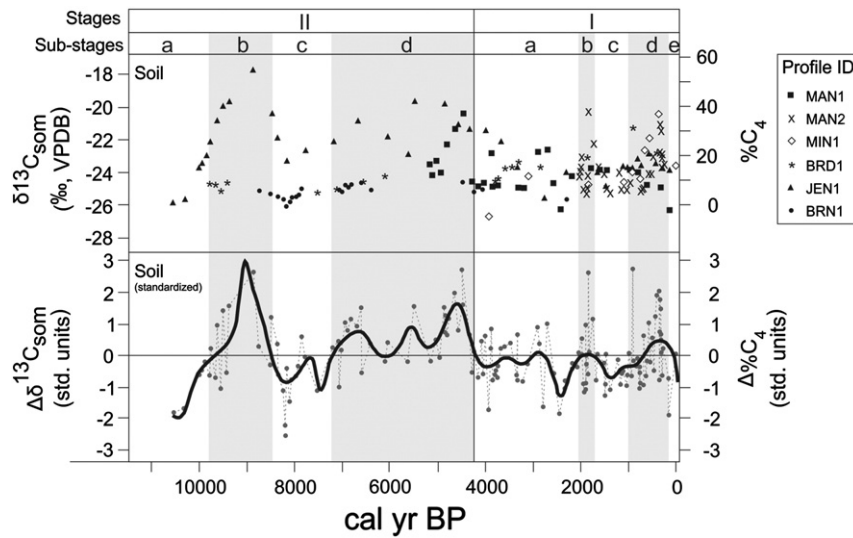


Figure 7. Carbon isotope time-series in unstandardized and standardized format. Carbon isotope stages and sub-stages defined based on raw $\delta^{13}\text{C}_{\text{som}}$ data, standardized $\delta^{13}\text{C}_{\text{som}}$ data, and deconvolution curve output (solid black curve). Isotope stages and sub-stages were determined using notable changes, or discontinuities, in both the raw $\delta^{13}\text{C}_{\text{som}}$ and deconvolution curve.

indicate low mean July temperatures (Viau et al., 2002; Shuman et al., 2004, 2009) (Fig. 9). Sub-stage IIa also coincides with increased moisture levels following a lake lowstand (13.4–10.9 cal ka BP) in the New England region (Newby et al., 2011). However, the Rocky Pond and New Long Pond lake records also document lowstands during this interval (Newby et al., 2009). The beginning of isotope sub-stage IIa is poorly constrained in age and $\delta^{13}\text{C}_{\text{som}}$ and additional radiocarbon data are required to properly evaluate its paleoclimate context.

Sub-stage IIb (9.9–8.3 cal ka BP) shows positive $\Delta\delta^{13}\text{C}_{\text{som}}$ ($\Delta\%C_4$) values using the measurement error model. These values indicate above-average $\delta^{13}\text{C}_{\text{som}}$ and a period of time with above-average $\%C_4$ plants. Phytoliths from this interval show 9.3% *Panicoideae* phytoliths and independently confirms the presence of C_4 grasses (Table 3, Figs. 5 and 6). Additional deconvolution measurement error model runs were performed with the 9.3 ka-equivalent 2.64 $\Delta\delta^{13}\text{C}_{\text{som}}$ value ($\delta^{13}\text{C}_{\text{som}} = -17.7\text{‰}$) removed and the results still show a

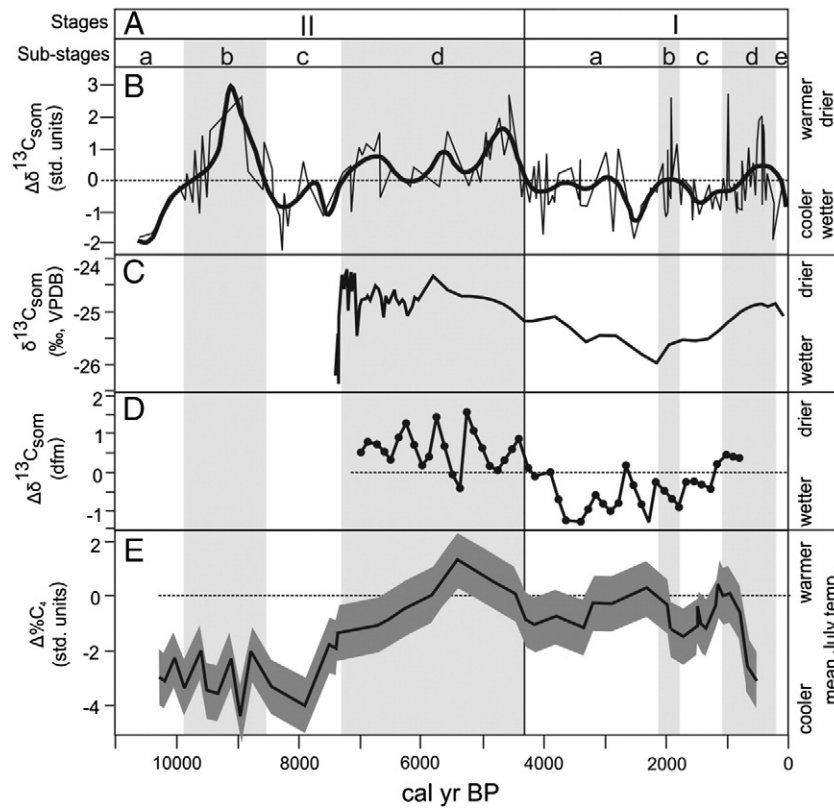


Figure 8. Comparison of carbon isotope stage and sub-stage (A) and deconvolution curve data (B) from this study with previous $\delta^{13}\text{C}$ studies in alluvial and loessal settings (C–E). (C) $\delta^{13}\text{C}$ from alluvial soil organic matter, West Virginia (Springer, 2010). (D) $\delta^{13}\text{C}$ from alluvial soil organic matter from southeastern Tennessee (Driese et al., 2008). Data were resampled and plotted as deviation from mean (dfm) (after Kocis, 2011). (E) Topsoil $\delta^{13}\text{C}$ from alluvial and loessal sites throughout the Midwest, USA (after Nordt et al., 2008). Data were standardized to the relative difference from the modern $\delta^{13}\text{C}$ latitudinal equivalent in the Great Plains.

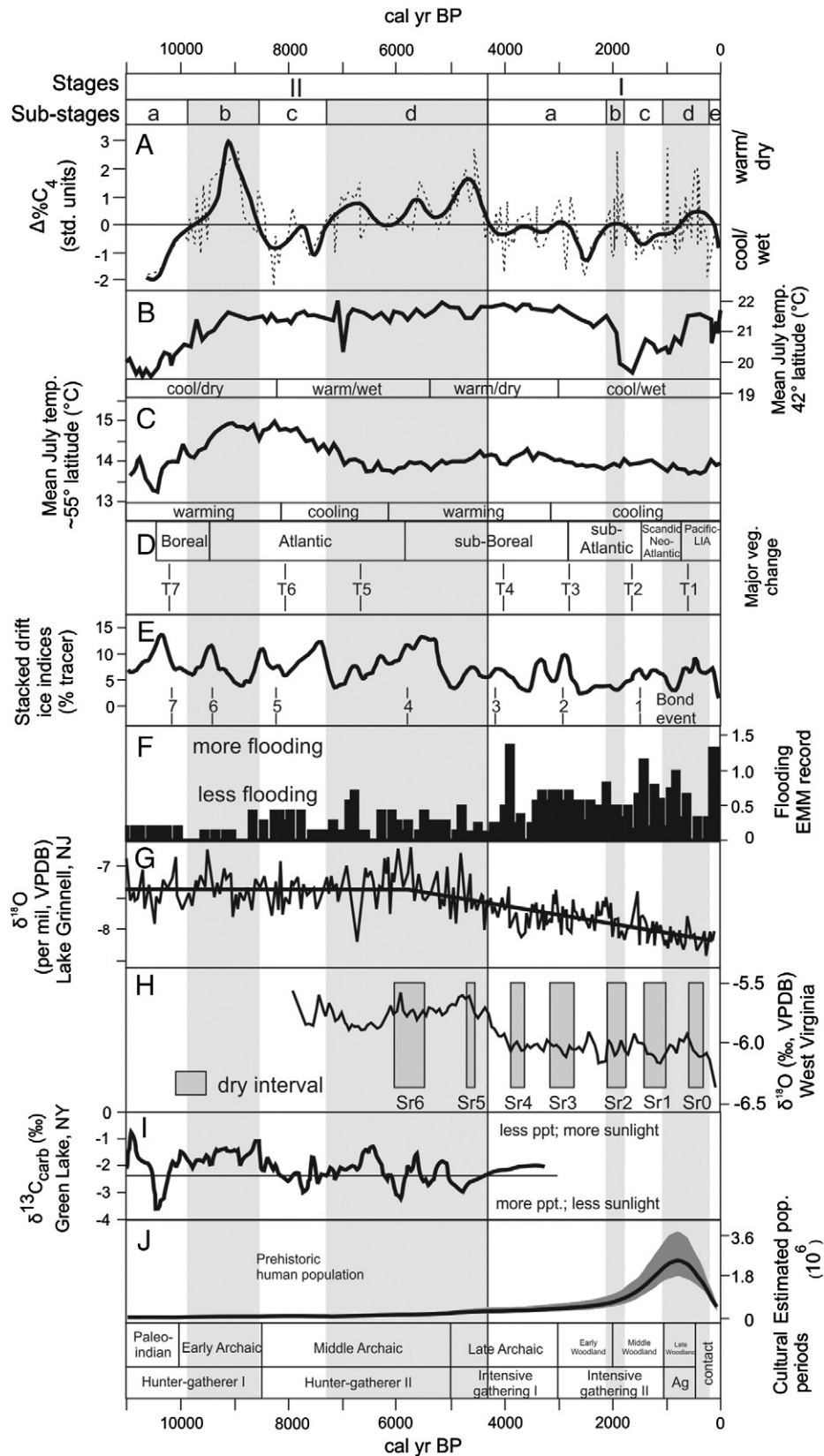


Figure 9. Comparison of carbon isotope stage, sub-stage and deconvolution curve data (A) from this study and regional paleoclimate and archaeological records (B–J). (B) Pollen-derived paleotemperature from Crooked Pond, MA (after Shuman et al., 2009) with northeastern USA temperature/moisture intervals noted (Shuman et al., 2004). (C) Pollen-based mean July paleotemperature for northern Quebec with North American temperature intervals noted (Viau et al., 2006). (D) Major climate episodes, Boreal – Little Ice Age (LIA), (Wendland and Bryson, 1974) along with widespread vegetation transitions, e.g., T6 for North America (Viau et al., 2002). (E) Stacked drift ice index that includes various petrologic traces from North Atlantic cores (Bond et al., 2001) and numbered Bond events (Bond et al., 1997). (F) Lacustrine-based flood events from New England region (Parris et al., 2010). (G) Oxygen isotope data from White Lake, NJ (Zhao et al., 2010). (H) Oxygen isotope data from West Virginia speleothem (Springer et al., 2008). (I) Lacustrine $\delta^{13}\text{C}_{\text{carbonate}}$ data from Green Lake, NY (Kirby et al., 2002). (J) Charcoal-based prehistoric population estimate for North America (Peros et al., 2010) and major cultural periods with subsistence strategy for eastern Pennsylvania (Custer, 1996). Ag = agriculture.

time period marked by above-average $\delta^{13}\text{C}_{\text{som}}$ ($\%C_4$). The increase in C_4 grasses and potential for open or semi-open canopy during sub-stage IIb agrees with previous documentation of early Holocene open forests recorded in the New England region (Faison et al., 2006). Additionally, sub-stage IIb immediately follows major vegetation change documented between 10.2 and 10.5 ka (Wendland and Bryson, 1974; Viau et al., 2002) and an ice-rafting debris event (Bond et al., 1997).

$\Delta\delta^{13}\text{C}_{\text{som}}$ ($\Delta\%C_4$) based paleoclimate interpretations for sub-stage IIb agree with previously published paleotemperature and paleohydrology for the northeastern USA region (Fig. 9). Northern Hemisphere insolation reached a maximum ~10.0 ka (Kutzbach, 1981). Using pollen from 181 Northern Quebec sites, Viau et al. (2006) modeled an early Holocene warming and so sub-stage IIa to IIb could simply reflect an increase in summer growing season temperatures and the expansion of C_4 plants in the floodplain setting. This warming was also documented using pollen in northeastern USA (Shuman et al., 2009). However, pollen, lacustrine sediment, and $\delta^{13}\text{C}$ lacustrine-carbonate data from New York and the New England region suggest cool/dry conditions dominated the early Holocene (Kirby et al., 2002; Shuman et al., 2004; Newby et al., 2009; Mullins et al., 2011; Newby et al., 2011) and Hou et al. (2012) document abrupt cooling events centered at 9.9, 9.2, 8.8, and 8.4 cal ka BP during the sub-stage IIb interval. The effects of the Laurentide Ice Sheet on the regional atmospheric circulation, which lasted until ~8.0 ka, enhanced the strength of the westerlies, creating an overall drier climate. The maximum $\Delta\delta^{13}\text{C}_{\text{som}}$ centered at 9.2 cal ka BP coincides with the abrupt cooling and drying event registered in the New England region (Newby et al., 2009; Hou et al., 2012). And thus, above-average $\delta^{13}\text{C}_{\text{som}}$ ($\%C_4$) values during sub-stage IIb may also reflect $\delta^{13}\text{C}_{\text{som}}$ enrichment due to water-stress effects on C_3 and C_4 plants.

Sub-stage IIc (8.3–7.0 cal ka BP) shows negative $\Delta\delta^{13}\text{C}_{\text{som}}$ ($\Delta\%C_4$) values using the measurement error model. These values indicate a period of below-average $\delta^{13}\text{C}_{\text{som}}$ and a period of time with decreased $\%C_4$ plants, suggesting a possible decrease in growing season temperature, increase in moisture availability, or both. Although the suggested decrease in the growing season temperature inferred from below-average $\delta^{13}\text{C}_{\text{som}}$ ($\%C_4$) coincides with an onset of cooling in Northern Quebec, there is an overall warming inferred in lake data from the New England region (Shuman et al., 2004; Viau et al., 2006). The onset of below-average $\delta^{13}\text{C}_{\text{som}}$ ($\%C_4$) during Sub-stage IIc coincides with the final 8.2 ka collapse of the Laurentide Ice Sheet (Barber et al., 1999). Sub-stage IIc also coincides with ice-rafting Bond event 5 (Bond et al., 1997), and notable change in vegetation, T6 (Wendland and Bryson, 1974; Viau et al., 2002; Shuman et al., 2009).

A number of researchers have hypothesized increase in moisture for northeastern USA during the sub-stage IIc interval. Lacustrine $\delta^{13}\text{C}$ carbonate data from New York indicates a period of increased precipitation (Kirby et al., 2002) and warm/wet conditions were documented from 8.2 to 5.4 cal ka BP as inferred from pollen (Shuman et al., 2004). A cluster of flood events centered at 8.2 cal ka BP are also recorded in lakes from the New Hampshire and Maine region (Parris et al., 2010). Although $\Delta\%C_4$ growing season temperature inferences are less straightforward during this interval, the below-average $\Delta\delta^{13}\text{C}_{\text{som}}$ and inferred decrease in potential moisture-stress agree with paleoprecipitation reconstructions throughout the northeastern USA.

Sub-stage IIId (7.0–4.3 cal ka BP) shows positive $\Delta\delta^{13}\text{C}_{\text{som}}$ ($\Delta\%C_4$) values in the measurement error model. These values indicate a period of above-average $\delta^{13}\text{C}_{\text{som}}$ and a period of time with increased $\%C_4$ plants. Furthermore, the measurement error model shows three distinct modes of elevated $\Delta\delta^{13}\text{C}_{\text{som}}$ ($\Delta\%C_4$) values, which suggests millennial to centennial-scale variations in the $\delta^{13}\text{C}_{\text{som}}$ and $\%C_4$ plants. Middle Holocene ragweed (*Ambrosia* spp.), grasses (*Poaceae*), and sedges (*Cyperaceae*) were also documented along lower Delaware River valley floodplains (Southgate, 2010). Above-average $\delta^{13}\text{C}_{\text{som}}$ and $\%C_4$ plants along the middle Delaware River valley coincide with increased $\%C_4$ throughout the Great Plains (Nordt et al., 2008) and moisture-stressed C_3 plants documented along Savannah Creek, Tennessee (Driese et al.,

2008; Kocis, 2011) (Fig. 8). The coherence between all three $\delta^{13}\text{C}_{\text{som}}$ records lends support for widespread and multiple, centennial-scale, episodes of middle Holocene hypsithermal warming and drying.

The warm/dry interval coeval with sub-stage IIId has been documented throughout much of the continent and northeastern USA (Wendland and Bryson, 1974; Viau et al., 2002). Furthermore, lacustrine $\delta^{13}\text{C}$ carbonate data from New York also suggests centennial-scale episodes of decreased precipitation or more sunlight (Kirby et al., 2002) similar to the multiple modes observed in southeastern Tennessee and along the Delaware River. Lacustrine sediment and pollen data from the New England region also show evidence of a warm/dry interval (Shuman et al., 2004; Zhao et al., 2010; Newby et al., 2011). The combined middle Holocene warming and drying likely influenced the $\delta^{13}\text{C}_{\text{som}}$ through expansion of C_4 plants and moisture stress throughout the study area.

Stage I (4.3–0 cal ka BP) was divided into 5 sub-stages (a–e) that span the late-middle to latest Holocene (Figs. 7–9). Sub-stage Ia (4.3–2.0 cal ka BP) shows negative $\Delta\delta^{13}\text{C}_{\text{som}}$ ($\Delta\%C_4$) values using the measurement error model with a minor positive episode at ~3.0 cal ka BP. These values indicate a period of below-average $\delta^{13}\text{C}_{\text{som}}$ and a period of time with decreased $\%C_4$ plants. The less than average $\delta^{13}\text{C}_{\text{som}}$ and $\%C_4$ plants suggest decreasing growing-season temperatures and an increase in available moisture. Middle to late Holocene $\delta^{13}\text{C}_{\text{som}}$ values from West Virginia, southeastern Tennessee and the Great Plains all show a shift towards below-average $\delta^{13}\text{C}_{\text{som}}$ and $\%C_4$ plants (Driese et al., 2008, after Kocis, 2011; Nordt et al., 2008; Springer et al., 2008). This is thought to be related to the onset of the cooler Neoglaciation (Viau et al., 2002) and increased moisture (Parris et al., 2010) as discussed previously.

Lacustrine sediment and pollen data from the New England region indicate a transition towards a cool/wet climate interval much later around 3.2 cal ka BP (Shuman et al., 2004). The minor above-average $\delta^{13}\text{C}_{\text{som}}$ ($\%C_4$) centered at 3.0 cal ka BP coincides with the onset of the Sub-Atlantic, a widespread vegetation change (T3) and an ice-rafting debris episode, Bond event 2 (Wendland and Bryson, 1974; Bond et al., 1997; Bond et al., 2001; Viau et al., 2002). This episode has been documented as a widespread cooling event, which runs counter to an increase growing season temperature interpretation based on sub-stage Ia $\Delta\delta^{13}\text{C}_{\text{som}}$ ($\Delta\%C_4$) values. Rather the slightly above-average Ia $\delta^{13}\text{C}_{\text{som}}$ values indicate a dry interval. Coeval dry conditions were documented in nearby Lake Grinnell, NJ (Li et al., 2007), upstate New York (Mullins et al., 2011), and Davis Pond, MA (Newby et al., 2011). West Virginia speleothem Sr/Ca ratios also indicate a dry interval around 3.0 cal ka BP (Springer et al., 2008).

Sub-stage Ib (2.0–1.8 cal ka BP) shows slightly negative $\Delta\delta^{13}\text{C}_{\text{som}}$ ($\Delta\%C_4$) values using the measurement error model, but with notably high $\delta^{13}\text{C}_{\text{som}}$ values previously recorded in the MAN1 trench (Stinchcomb et al., 2011). Although the measurement error model indicates a period of below-average $\delta^{13}\text{C}_{\text{som}}$ and lower $\%C_4$ plants, the positive $\delta^{13}\text{C}_{\text{som}}$ values suggest a period of elevated C_4 plants at the MAN1 locality. These data coincide with a dry interval in Sr/Ca ratios from a West Virginia speleothem (Springer et al., 2008) and precede a major vegetation change (T2) at 1.6 ka (Viau et al., 2002). Other than these examples, it is unclear whether or not Sub-stage Ib correlates with a known climate event. Alternatively, the sub-stage Ib excursion could reflect either forest clearance and dominance of early-successional grass species or an introduction of maize along the floodplain (or both). The earliest archaeological evidence of maize (a C_4 Panicoid grass) derives from neighboring central New York and dates to the beginning of this sub-stage (Hart et al., 2007). The carbon isotopic signature of forest clearance was documented previously by Springer et al. (2010) (Fig. 8), whereas Stinchcomb et al. (2011) documented evidence of land-use related to maize-based agriculture along the Delaware River valley, albeit 1000 years later.

Sub-stage Ic (1.8–1.0 cal ka BP) shows an interval of negative $\Delta\delta^{13}\text{C}_{\text{som}}$ ($\Delta\%C_4$) values, using the measurement error model. These data indicate an interval of below-average $\delta^{13}\text{C}_{\text{som}}$ and a period of time with decreased $\%C_4$ plants. The beginning of sub-stage Ic (1.8 cal ka BP)

coincides with a peak in lacustrine-based flood events in the New Hampshire and Maine region (Parris et al., 2010). However, West Virginia Sr/Ca data from a speleothem infer a drought episode (Springer et al., 2008). It is possible that some of the dry intervals are not as widespread, and the paleoclimate records from Maine to West Virginia record local variability in centennial to millennial-scale hydroclimatic variability.

Sub-stage Id (1.0–0.3 cal ka BP) indicates an interval of mostly positive $\Delta\delta^{13}\text{C}_{\text{som}}$ ($\Delta\%C_4$) values as shown in the measurement error model and raw $\delta^{13}\text{C}_{\text{som}}$ values. These data indicate an interval of above-average $\delta^{13}\text{C}_{\text{som}}$ and increasing $\%C_4$ plants through time. The trend of increasing $\delta^{13}\text{C}_{\text{som}}$ and increasing $\%C_4$ recorded from sub-stage Id-Ie (1.8–0.3 cal ka BP) coincides with increasing $\delta^{13}\text{C}_{\text{som}}$ from alluvium in West Virginia and southeastern Tennessee (Driese et al., 2008; Springer et al., 2010). These data suggest a widespread increase in the $\%C_4$ and may indicate Pollen-based temperature reconstructions document an increase in the mean July temperature further north (Medieval Warm Period equivalent) and the above-average $\%C_4$ for sub-stage Id may reflect this. Lacustrine-based paleohydrology data from Maine and New Hampshire show a decrease in the frequency of flooding events (Parris et al., 2010), whereas West Virginia Sr/Ca data from speleothems suggest a drought interval. The above-average $\delta^{13}\text{C}_{\text{som}}$ could also reflect this decrease in moisture availability. However, lake lowstands from the New England region are conflicting (Shuman et al., 2009; Newby et al., 2011) and the interval from 3.2–0 ka has been documented as a wet interval during the Holocene (Shuman et al., 2004; Shuman and Plank, 2011).

Alternatively, the above-average $\%C_4$ and $\delta^{13}\text{C}_{\text{som}}$ for substage Id could reflect an increasing reliance on maize-based agriculture and land clearance in the study area (Stinchcomb et al., 2011). Maize (*Zea mays*) is a C_4 grass of the Panicoideae subfamily and the increased use would have altered the underlying $\delta^{13}\text{C}_{\text{som}}$. Additionally, archaeological data suggest an increase in prehistoric human population that peaks approximately ~0.4 ka (Peros et al., 2010), and coincides with a shift towards maize-based agriculture throughout much of eastern North America (Custer, 1996). Increasing population pressure and land-use change likely resulted in increased forest clearance for farming, hunting, mast resource management, and habitation (Ekdahl et al., 2004; Fesenmyer and Christensen, 2010; Munoz and Gajewski, 2010). More frequent and widespread forest clearance likely created open-canopy ecosystems where early-successional C_3 and C_4 grasses could thrive, thus enriching the belowground $\delta^{13}\text{C}_{\text{som}}$ signature. The sub-stage Id isotope record from the Delaware River valley reflects this prehistoric anthropogenic input. Studies of calcite $\delta^{18}\text{O}$ from West Virginia speleothems also depict a heightened level of environmental manipulation during this era (Springer et al., 2010). These data show that prehistoric land-use practices, in addition to other sources of vegetation change, can influence the $\delta^{13}\text{C}_{\text{som}}$ record in alluvium.

Substage Ie (0.3–0 cal ka BP) is poorly constrained, having 3 $\delta^{13}\text{C}_{\text{som}}$ observations from three separate trenches. Given this, the measurement error model shows a trend towards negative $\Delta\delta^{13}\text{C}_{\text{som}}$ ($\Delta\%C_4$) values, implicating below average $\delta^{13}\text{C}_{\text{som}}$ ($\%C_4$) values for this time period. The return to a more C_3 -dominated ecosystem could have resulted from post-pandemic forest recovery. The ~200 years that followed rapid population decline due to European-introduced diseases would have been sufficient time for forest growth and regeneration. This hypothesis cannot be properly addressed due to few high-resolution $\delta^{13}\text{C}_{\text{som}}$ alluvial records spanning the past 1.0 ka.

$\delta^{13}\text{C}_{\text{som}}$ as a paleoenvironmental proxy for eastern North America

The Delaware River valley $\delta^{13}\text{C}_{\text{som}}$ curve shows that variations in C_3 and C_4 plant communities (driven by changes in summer growing season temperature) and environmental effects (e.g., moisture stress) correlate with established climate changes on both orbital and millennial time-scales. Orbital scale climate changes such as the Northern

Hemisphere insolation maximum during the early Holocene and middle Holocene climatic optimum appear to have influenced the $\delta^{13}\text{C}_{\text{som}}$ via changes in the relative abundance of C_3 and C_4 plants by means of long-term changes in growing season temperatures. Drought intervals resulting from both orbital and millennial-scale climate changes also coincide with variations in the $\delta^{13}\text{C}_{\text{som}}$, suggesting moisture stress affected the plant and resulting below-ground $\delta^{13}\text{C}$. If environmental, biological, and geomorphic influences on the $\delta^{13}\text{C}_{\text{som}}$ can be accounted for, this isotope application holds great promise as a paleotemperature proxy for eastern North America.

Carbon isotope research in the Great Plains indicates a quantitative link between the $\delta^{13}\text{C}$ of topsoil organic matter and mean July temperature (Nordt et al., 2007). Excluding the BRN1 trench (no topsoil data) and incipient A horizons forming on recent flood deposits, the average $\delta^{13}\text{C}_{\text{som}}$ from topsoil A horizons is -24.0% (± 0.9). Using a paleotemperature transfer function developed for forest-grassland ecosystems occurring along the eastern Great Plains this would yield a mean July temperature of 21.7°C (± 1.1). This estimated mean July temperature value is within range of the current mean July temperature of 22.2°C , based on 1981 and 2010 climate data (NOAA, 2011). This preliminary test suggests the $\delta^{13}\text{C}_{\text{som}}$ from buried A horizons in eastern North America could contain quantitative paleotemperature information that will enhance our understanding of regional climates during the Holocene.

Conclusions

The results from this study demonstrate that the novel approach of combining $\delta^{13}\text{C}$ of soil organic matter ($\delta^{13}\text{C}_{\text{som}}$) and phytolith analysis is a useful proxy for estimating the above-ground concentrations of C_3 : C_4 biomass in eastern North America. The $\delta^{13}\text{C}_{\text{som}}$, when standardized, reveal distinct intervals of above and below average $\delta^{13}\text{C}_{\text{som}}$ ($\%C_4$). The raw $\delta^{13}\text{C}_{\text{som}}$ and novel deconvolution curve method, which accounts for uncertainty in the time direction, reveal two major isotope stages and nine sub-stages for the past 10,700 years. These stages coincide with previously published paleoclimate and archaeological phenomena and suggest that variations in $\delta^{13}\text{C}_{\text{som}}$ are caused by changes in growing-season temperature, moisture availability, and human alterations of the ecosystem. $\delta^{13}\text{C}_{\text{som}}$ is relatively fast and easy to measure in floodplain sediments and buried soils, especially in deposits lacking pedogenic and lithogenic carbonates. Close sample spacing (between 5 and 10 cm) in a depth transect ensures that inflections recording important changes are not missed. Further research on progressive changes in SOM with depth and time in buried soil profiles across landscapes (e.g., Fang et al., 2010) may help improve understanding of $\delta^{13}\text{C}_{\text{som}}$ values. More comprehensive floodplain-based $\delta^{13}\text{C}_{\text{som}}$ studies will provide a higher resolution more regionally based paleoclimate than before.

Supplementary data to this article can be found online at <http://dx.doi.org/10.1016/j.yqres.2013.02.004>.

Acknowledgments

Support for this study was provided by NSF EAR 0823131 awarded to S. G. Driese, which supported radiocarbon dating, thin-sections, and $\delta^{13}\text{C}$ analysis, and a GSA student-aid-in research grant and Baylor research grant awarded to G. E. Stinchcomb which supported radiocarbon dating. We thank Brinnen Carter and the Delaware Water Gap National Recreation Area for providing additional funds for radiocarbon dating, Chase Gerken, John McFadden, Steve Dworkin, Ren Zhang, Lauren Michel, Sally Horn, Zheng-Hua Li, and Jason Mintz for field and lab assistance, James Lee, Michael Stewart, and John Stiteler for introducing us to the Browning Site, and Jane Harvill and Jack Tubbs for statistical assistance. Derek Booth, John Dodson, and two anonymous reviewers provided comments and suggestions that greatly improved an earlier version of this manuscript.

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