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A high-resolution speleothem record of climatic variability at the Allerød–Younger Dryas transition in Missouri, central United States

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Abstract

Stable isotopic analysis and $^{234}U/^{230}Th$ mass spectrometry dating of a single calcite stalagmite from Onondaga Cave, east-central Missouri, reveal climatic variability between 13 200 and 12 400 yr BP coincident with the Allerød–Younger Dryas transition in the GISP2 Greenland ice cores. $\delta^{18}O$ values of speleothem calcite mark decreasing temperatures between 13 200 and 13 000 yr BP while more rapidly decreasing $\delta^{13}C$ values record a sharp shift toward C_3 vegetation at approximately 13 100 yr BP. Because of uncertainties in the chronological control of both this speleothem and the GISP2 record and the brevity of this speleothem record ($\sim$800 years), direct comparison of the two sequences is difficult. The interval preserved by this speleothem may span the inter-Allerød cold period (IACP) alone, both the IACP and the start of the Younger Dryas, or only the first half of the Younger Dryas. In any case, these data demonstrate that the climate of the North American interior deteriorated significantly and rapidly during the interval associated with the Allerød–Younger Dryas transition.

Keywords: Younger Dryas; Allerød; speleothem; Missouri; isotope

1. Introduction

Multi-component analyses ($\delta^{18}O$, δD, trace elemental abundance, electrical conductivity) of annually laminated Greenland ice sequences illustrate the rapid reorganization of climatic systems associated with the Younger Dryas (YD) (12 940–11 640 yr BP/11 000–10 200 ^{14}C yr BP) and the preceding Allerød warm period. Analysis of the GRIP (Dansgaard et al., 1993) and GISP2 (Grootes et al., 1993; Mayewski et al., 1993; Taylor et al., 1993; Stuiver et al., 1995) ice cores demonstrates that the Allerød–YD transition occurred rapidly (in the span of a few decades) and that significant climatic variability accompanied this shift (Alley et al., 1993; Grootes et al., 1993). This climatic instability extends back into

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the late Allerød to the inter-Allerød cold period (IACP) which is characterized by a ~4‰ decrease and subsequent rise in glacial ice $\delta^{18}O$ values between 13 250 and 13 050 yr BP (Mayewski et al., 1993).

The YD is proposed to have originated when glacial meltwater was diverted via the St. Lawrence River to the North Atlantic ocean where it interrupted thermohaline circulation (Rooth, 1982; Broecker et al., 1989). Although the YD is recorded in other regions (Kuhry et al., 1993; Denton and Hendy, 1994; Thompson et al., 1995), the YD is best known from studies of the North Atlantic ocean and North Atlantic continental margins (Wright, 1989; Alley et al., 1993; Dansgaard et al., 1993; Bjorck et al., 1996) and Europe (Eicher et al., 1981; Dansgaard et al., 1989; Lehman and Keigwin, 1992). Few late Pleistocene high-resolution paleoclimatic records preserve the effects of the YD on the North American interior, however. Paleoclimatic records from western North America have demonstrated a YD climatic signal (Engstrom et al., 1990; Mathewes et al., 1993; Reasoner et al., 1994; Gosse et al., 1995), but paleoclimatic records spanning the YD chronozone are particularly rare in the southern Midwest where pre-Holocene pollen spectra are scarce and many other sequences used to constrain climatic conditions related to the YD (e.g., alpine glacial deposits, Thompson et al., 1995) are nonexistent.

Speleothems of YD age have been documented from areas including New Zealand (Hellstrom et al., 1998) and Israel (Frumkin et al., 1999) but they are extremely rare in the North American continental interior. Here we present a $^{230}Th/^{234}U$ chronology of isotopic variation in a single calcite stalagmite from east-central Missouri (Fig. 1) that preserves a detailed record of climatic variability in central North America at the Allerød–YD transition.

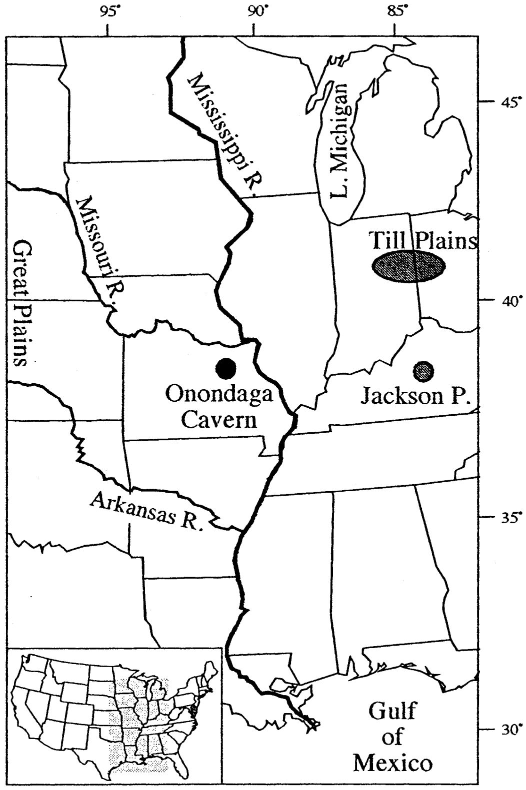

Fig. 1. Map of the central USA showing location of Onondaga Caverns, Missouri, Jackson Pond, Kentucky, and the Till Plains of Ohio and Indiana.

2. Methods

Stalagmite ON-3-B was collected from Onondaga Caverns (38°3′N, 91°13′W) which formed in the Ordovician Gasconade Dolomite near Leasburg, MO, central USA, on the northeastern edge of the Ozark Highlands (Bretz, 1956). ON-3-B was sawed vertically in half, polished and sampled at 1-mm intervals for carbon and oxygen isotopic analysis using a drill with a 500-μm tip (Fig. 2). Samples were drilled from the central portion of the speleothem. The entire 16.5 cm of ON-3-B is composed of dense, optically clear calcite characterized by visible banding on the scale of micrometers and is without any signs of recrystallization. Stable isotopic analyses were performed at the University of Michigan Stable Isotope Laboratory with analytical precision better than 0.1‰. Carbon and oxygen isotopes are normalized to the Pee Dee Belemnite isotopic standard (Fig. 3).

Radiometric dating was conducted on 200–300-mg-sized samples at the University of New Mexico Radiogenic Isotope Laboratory using a Mi-

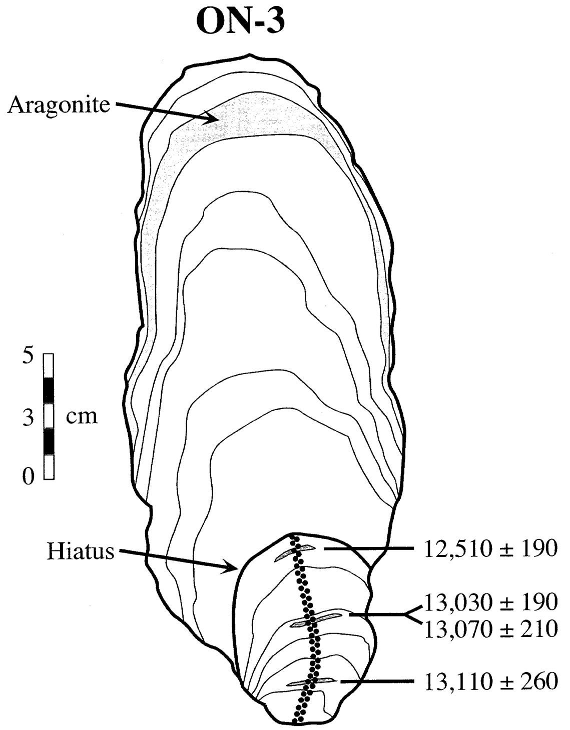

Fig. 2. Representation of stalagmite ON-3-B showing growth banding and location of pits milled for stable isotopic and radiometric analysis.

cromass Sector 54 thermal ionization mass spectrometer (Table 1). Analytical procedures are modifications of those of Edwards et al. (1987) as discussed in Edwards et al. (1993). All isotopes of interest (^{229}Th, ^{230}Th, ^{232}Th, ^{233}U, ^{234}U, ^{235}U, ^{236}U) were measured on an ion-counting Daly multiplier with abundance sensitivity in the range of 20 ppb at one mass distance in the mass range of U and Th, requiring very little background correction even for samples with large ^{232}Th content. Multiplier dark noise was about 0.12 cps. NBL-112 U standard was measured during the course of this study and was always in the range of 0.1% of the accepted ^{234}U/^{238}U ratio. Ages, determined using linear interpolation between dated intervals and linear extrapolation beyond dated intervals, are in years before present (yr BP) where present is 1998–2000, the years of these isotopic analyses. Radiocarbon ages from the literature were converted to calendar years before present (yr BP) using the CALIB radiocarbon calibration program (Stuiver and Reimer, 1993). Ages reported here are calendar years before present unless noted otherwise.

3. Results

3.1. Chronology

Based on annual layer counting in Greenland ice sheets, the YD has been dated from 12 940 ± 260 to 11 640 ± 250 calendar yr BP with a duration of 1300 ± 70 years (Alley et al., 1993; Meese et al., 1994). Mayewski et al. (1993) defined the IACP at 13 250 and 13 050 yr BP and noted

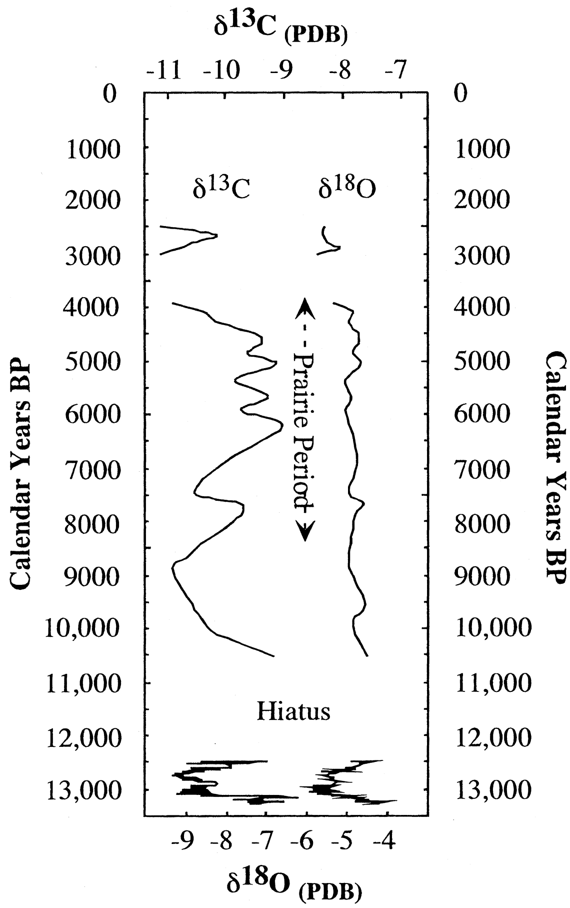

Fig. 3. Carbon and oxygen isotopic compositions of ON-3-B during both the late Pleistocene and Holocene. Holocene section from Denniston et al. (1999b).

that it was characterized by conditions similar to those of the YD. $^{230}Th/^{234}U$ mass spectrometry dates were obtained from the top, middle, and bottom of ON-3-B (Table 1). Based on the absence of hiatuses or sharp changes in speleothem morphology, deposition of stalagmite ON-3-B appears to have occurred continuously from 13 250 to 12 400 yr BP based on linear extrapolation from dated areas. A prominent hiatus marks a growth interruption that occurs after this time period and persisted from 12 400 to ~10 000 yr BP. The combined age uncertainties of the GISP2 and ON-3-B records prevent extremely precise comparison; we cannot resolve, therefore, whether the older portion of the ON-3-B record is coincident with the IACP only, both the IACP and the YD, or just the YD as defined by the GISP2 sequence (Fig. 4).

3.2. Stable isotopes

Both carbon and oxygen isotopic compositions demonstrate considerable temporal variability throughout ON-3-B (Fig. 3). Ages for each carbon and oxygen isotope analysis were calculated based on linear interpolation from dated intervals. Average oxygen isotopic values decrease from −4.1‰ to −5.6‰ by 13 050 yr BP, increase briefly to −5.2‰ at 12 800 yr BP, decrease to −5.6‰ by 12 700 yr BP, increase gradually to −4.3‰ by 12 450 yr BP, and finally decrease to −4.9‰ by 12 400 yr BP. Carbon isotopic values follow a similar overall trend as oxygen isotopic values although the magnitudes of the carbon isotopic shifts are larger and the changes occur more rapidly. While oxygen isotopic values decrease more or less continuously from 13 250 to 13 050 yr BP, $\delta^{13}C$ values are more erratic and do not begin a systematic decline until 13 100 yr BP, approximately 150 years after declines in $\delta^{18}O$ values, reaching −10.4‰ by 13 050 yr BP. $\delta^{13}C$ values then rise to −9.9‰ at 12 850 yr BP, fall to −11.0‰ by 12 700 yr BP, rise again to −8.6‰ by 12 450 yr BP, and finally decrease to −10.3‰ by 12 400 yr BP.

Interpreting speleothem oxygen and carbon isotopic ratios in a paleoclimatic context requires that speleothem calcite crystallized under equilibrium conditions where high humidity, CO_2 levels, and temperature remained constant in the cave throughout the year (Hendy, 1971). $\delta^{18}O$ and $\delta^{13}C$ values obtained from samples drilled along the same growth horizon(s) are not covariant, suggesting that ON-3-B crystallized under isotopic equilibrium (Hendy, 1971). However, the evolution of carbon (Baker et al., 1997) and oxygen isotopes (Denniston et al., 1999a) in infiltrating fluids cannot necessarily be accurately constrained through the analysis of a single speleothem. Therefore, cave drip waters may have been enriched in ^{18}O by pre-infiltration evaporation or ^{13}C due to CO_2 out-gassing before entering the

Table 1
Uranium and thorium isotopic ratios and $^{230}Th/^{234}U$ ages

Sample	Distance from bottom[a] (mm)	^{238}U (ng/g)	^{232}Th (pg/g)	$\delta^{234}U^{b,c}$ measured	$^{230}Th/^{238}U$ activity	$^{230}Th/^{232}Th$ atomic	Uncorrected age (yr)	Corrected age[d] (yr)
ON-3-B	19	266	9740	1599 (7)	1.18E-2 (0.6)	1.39E-4 (1)	13 580 (180)	13 110 (260)
ON-3-B-a	48	283	9480	1870 (7)	3.35E-1 (3)	1.65E-4 (2)	13 400 (130)	13 070 (210)
ON-3-B-b	48	282	7920	1870 (8)	3.33E-1 (3)	1.96E-4 (2)	13 310 (140)	13 030 (190)
ON-3-B	75	268	4840	1826 (8)	3.15E-1 (4)	2.86E-4 (4)	12 730 (160)	12 510 (190)

ON-3-B-a and ON-3-B-b are replicate samples analyzed from a homogenized powder drilled from a single area of the speleothem.

[a] The total length of ON-3-B is 16.5 cm.

[b] $\delta^{234}U_{measured} = [(^{234}U/^{238}U)_{measured}/(^{234}U/^{238}U)_{eq} - 1] \times 10^3$, where $(^{234}U/^{238}U)_{eq}$ is the secular equilibrium atomic ratio: $\lambda_{238}/\lambda_{234} = 5.472 \times 10^{-5}$.

[c] Values in parentheses represent 2σ errors in the last significant figure.

[d] Unsupported ^{230}Th was corrected using an initial $^{230}Th/^{232}Th$ ratio of 4.4×10^{-6} ($\pm 2.2 \times 10^{-6}$) representative of average crustal silicates.

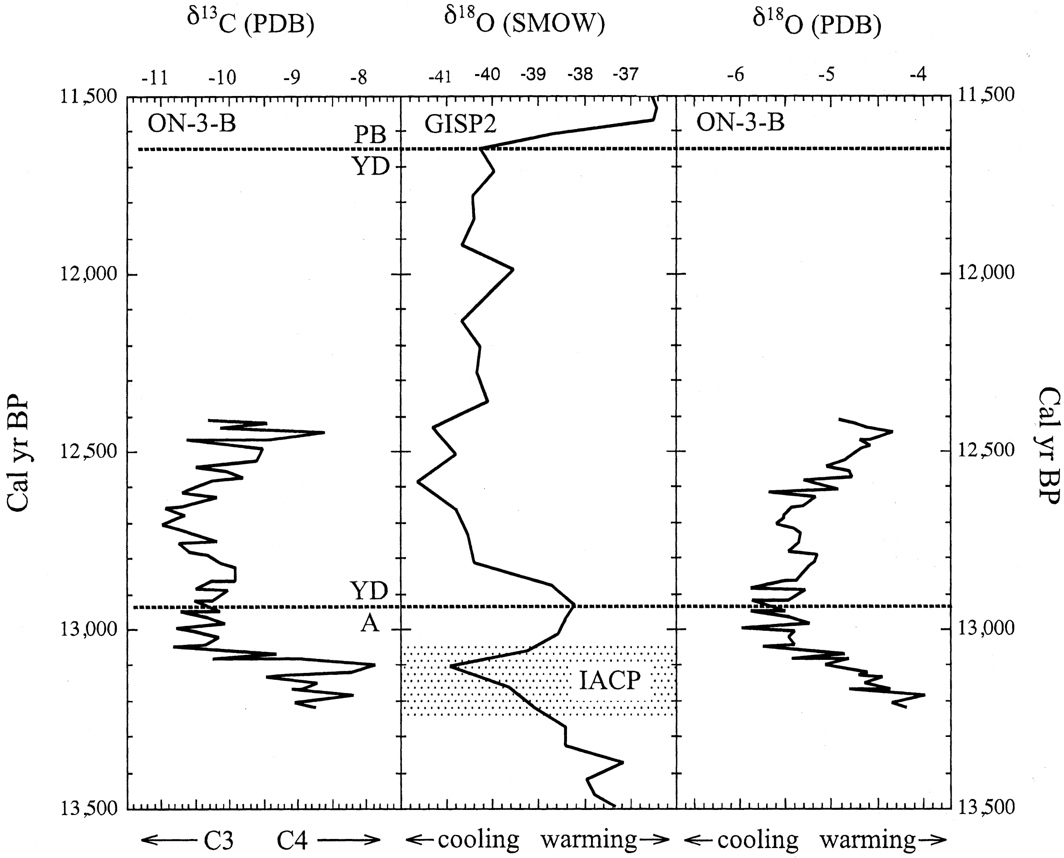

Fig. 4. Carbon and oxygen isotopic compositions of stalagmite ON-3-B and oxygen isotopic composition of GISP2 Greenland ice record (Grootes et al., 1993; Mayewski et al., 1993; Meese et al., 1994; Stuiver et al., 1995) plotted against both calendar years and radiocarbon years before present. YD = Younger Dryas; A = Allerød; PB = Pre-Boreal; IACP (shaded area) = inter-Allerød cold period. C_4 and C_3 along bottom axis of $\delta^{13}C$ plot correspond to temporal changes in vegetation type over the cave.

cave. The impact of these effects on speleothem isotopic values is beyond our ability to detect with only one speleothem.

4. Discussion

4.1. Constraining Allerød–YD temperature change in the North American mid-continent

The $\delta^{18}O$ values of speleothem calcite may be linked to mean annual temperature through the temperature dependence of calcite–water ^{18}O fractionation (Thompson et al., 1976; Friedman and O'Neil, 1977). The link between temperature and fractionation requires that (1) the cave must be sufficiently closed to the outside atmosphere so that it maintains a constant temperature throughout the year, (2) in-cave evaporative effects do not significantly alter the $\delta^{18}O$ of meteoric waters, (3) precipitation seasonality is understood, and (4) the $\delta^{18}O$ of the moisture source is constrained.

Onondaga Caverns is a long cave with little atmospheric exchange with the outside atmosphere in the area where this speleothem grew. The ~150-year separation between the decreases in carbon and oxygen isotopic values suggests that it is unlikely that in-cave evaporation affected speleothem $\delta^{18}O$ and $\delta^{13}C$ values. Also, the lack of covariance between carbon and oxygen isotopic values has been used as an indicator of isotopic equilibrium (Hendy, 1971). Cooler conditions that decrease speleothem calcite $\delta^{18}O$ values may also result in a more C_3-rich plant community over the cave, which can lower speleothem $\delta^{13}C$ values. The primary obstacle to constraining mean annual temperatures for the YD from $\delta^{18}O$ values of ON-3-B, therefore, remains establishing the oxygen isotopic composition of precipitation. It remains unclear how much meltwater drained into the Gulf of Mexico immediately prior to, during, and after the YD and thus oxygen isotopic values of Gulf of Mexico surface waters are poorly constrained (Flower and Kennett, 1990). As the predominant source of precipitation for southern Missouri is the Gulf of Mexico, a clear understanding of temporal changes in the oxygen isotopic composition of Gulf of Mexico surface waters is critical in assessing temperature/precipitation oxygen isotopic relationships in Onondaga speleothem calcite (Bryson, 1966). However, over short time scales, speleothem $\delta^{18}O$ values may reflect the magnitude of temperature change (Dorale et al., 1998).

The modern relationship between mean annual temperature and average precipitation oxygen isotopic composition is 0.6‰/°C (Dansgaard, 1964). When coupled with calcite–water fractionation effects, speleothem $\delta^{18}O$ values record temperature changes at 0.3‰/°C (Friedman and O'Neil, 1977). Therefore, the ~1.3‰ decrease in $\delta^{18}O$ values of ON-3-B suggests that mean annual temperatures declined ~4°C between 13 200 and 13 000 yr BP in east-central Missouri. The range of $\delta^{18}O$ values of ON-3-B is consistent with the 25 000–75 000 yr BP speleothem record of Dorale et al. (1998) from nearby Crevice Cave, Missouri. However, such a single, large shift in oxygen isotopic composition is not apparent in the Crevice Cave series.

Previous paleoclimatic reconstructions report similar temperature estimates for the YD in North America. Pollen spectra from the Till Plains of Ohio and Indiana were interpreted as indicating a 1–5°C drop in January temperatures and a 1–2°C drop in July temperatures (Shane and Anderson, 1993). Peteet et al. (1993) report a mean annual temperature depression of 3–4°C based on New Jersey pollen sequences. Mathewes et al. (1993) calculated a 2–3°C decline in summer temperatures during the YD on the British Columbia coast using pollen transfer functions. An alternative explanation for the decrease in speleothem $\delta^{18}O$ values during the YD is increased contributions of ^{18}O-depleted moisture derived from Pacific air masses. At present, moisture derived from the Gulf of Mexico is enriched in ^{18}O relative to a mixed Gulf–Pacific source in the north-central Midwest (Simpkins, 1995). Possible evidence for increased Pacific moisture during the YD has been obtained from elevated lake levels in the Estancia Basin, central New Mexico (Allen and Anderson, 1996). A shift in moisture source would complicate paleotemperature estimates by increasing the uncertainty of the oxygen isotopic composition of precipitation, and thus temperatures constrained here would represent maximum cooling. However, based on the dominance of Gulf of Mexico moisture in southern Missouri today (Bryson, 1966), the gulf most likely remained the primary moisture source throughout the late Allerød and YD; temperature estimates are therefore probably not significantly affected by air mass changes.

4.2. Vegetation changes associated with Allerød–YD cooling

Speleothem carbon is derived from dissolution of carbonate bedrock and from organic carbon liberated during plant respiration and decomposition (Dorale et al., 1992). During photosynthesis, nearly all trees, shrubs, forbs, and cool season grasses discriminate against ^{13}C in favor of ^{12}C. Many warm and dry season grasses utilize the C_4 (less discriminatory) photosynthetic pathway. The carbon isotopic composition of C_3 vegetation ranges from −32‰ to −22‰ while C_4 vegetation falls between −18‰ and −12‰ (Boutton, 1991). For example, as North American prairies are rich in C_4 vegetation (Teeri and Stowe, 1976), when forest replaces prairie, $\delta^{13}C$ values of soil organic matter decrease. Changes in the carbon isotopic composition of soil organic matter are then translated downward through infiltrating meteoric water and preserved in calcite speleothems (Dorale et al., 1992). On short time scales, moisture balances and summer temperatures determine the distribution of these vegetation types, with C_3 vegetation predominant in cooler and/or wetter climates and C_4 vegetation preferring warmer and/or drier conditions.

Speleothem carbon isotopic compositions can also change in response to factors other than vegetation dynamics (e.g., fluids out-gassing CO_2 before entering the cave) (Baker et al., 1997). Ideally, carbon isotopic values of several similarly aged speleothems from the same cave would be compared as a means of identifying samples influenced by such effects (Denniston et al., 1999a; Dorale et al., 1998). However, such a comparison is not possible in this circumstance because it is extremely rare to find even one stalagmite from this region that grew during this time interval. However, Holocene sections of this and other speleothems from Onondaga Cavern and four other regional caves are consistent in the direction and timing of shifts in carbon isotopic values and thus likely record regional vegetation change (Denniston et al., 2000).

Because the carbon isotopic composition of speleothem calcite is sensitive to changes in soil organic matter, the variability in $\delta^{13}C$ values throughout ON-3-B (e.g., 12 850 yr BP and 12 650–12 450 yr BP) likely represents shifts in the relative abundance of C_3 vegetation and thus alternating cooler/moister and drier/warmer conditions. The $\sim$2‰ decrease in $\delta^{13}C$ values occurring at 13 100 yr BP is similar in magnitude to the changes that occurred during the middle Holocene, when prairie replaced deciduous forest over Onondaga Cave (Denniston et al., 2000). A pollen sequence from Jackson Pond, central Kentucky also supports highly variable climatic conditions during the YD. At Jackson Pond, climate oscillations are documented by at least three pronounced changes in the ratio of spruce to oak pollen abundance between 13 200 and 11 200 yr BP (11 300 to 10 000 ^{14}C yr BP), which likely reflect shifts in the boreal/deciduous forest ecotone (Wilkins et al., 1991).

4.3. Comparison of the ON-3-B and GISP2 $\delta^{18}O$ records

Uncertainties in the dating of both the GISP2 and ON-3-B records and the fact that ON-3-B spans only $\sim$800 years preclude comparing the details of each sequence. However, significant cooling in southern Missouri is associated with the IACP and/or the YD (Fig. 4). Assuming that maximum cooling at 13 050 yr BP in the ON-3-B record is essentially synchronous with the IACP in the GISP2 record at 13 100 yr BP, then the warming that separated the IACP from the YD was dramatically larger in the North Atlantic than in the North American interior. During the interval separating the IACP from the YD (13 050 to 12 940 yr BP), $\delta^{18}O$ values of glacial ice rose by 3‰, corresponding to a maximum temperature increase of $\sim$4°C (Mayewski et al., 1993). During the same period, the $\delta^{18}O$ values of ON-3-B increased by $\sim$0.2‰, corresponding to a mean annual temperature increase of $<$1°C. Several distinct paleoclimatic records from western North America (e.g., alpine glacial moraines in the Rockies (Gosse et al., 1995), pollen spectra in the Pacific Northwest and southern Alaska (Engstrom et al., 1990; Mathewes et al., 1993), cold-water indicator benthic foraminifera species from the Pacific coast of Canada (Mathewes et al., 1993)), and from the mid-continent (e.g., Jackson Pond pollen record (Wilkins et al., 1991)) record a similar timing for the onset of cooling (e.g., 13 100 yr BP/11 200 ^{14}C yr BP). In addition, ice sequences from both the Tibetan Plateau (Guliya ice cap) and the Peruvian Andes (Huascarán glacier) also lack the higher $\delta^{18}O$ (warming) trends prior to the YD and instead record cooling from the early Bølling ($\sim$15 000 calendar yr BP) to the YD (Thompson et al., 1995, 1997).

Alternatively, the beginning of the ON-3-B record, and thus the onset of cooling in Missouri could correspond with the start of the YD. Beginning at 12 700 yr BP, $\delta^{18}O$ values in ON-3-B rose by 0.7‰ by 12 450 yr BP, implying $\sim$2°C of mean annual warming. While evidence for subtly rising temperatures during the first half of the YD also exists in the Greenland ice record (Mayewski et al., 1993), $\delta^{18}O$ values rise by approximately 1.5‰ between 12 700 and 12 350 yr BP (Fig. 4), corresponding to a temperature increase of $\sim$2°C, possibly similar to warming experienced in the mid-continent.

Finally, ON-3-B may have been precipitated only during the IACP. If this was the case, then growth may have ceased with the onset of the YD. The lack of documented speleothems of Younger Dryas age from the Midwest may indicate increased regional aridity.

5. Conclusions

The interval associated with the Allerød–YD transition was characterized by a shift toward a cooler climate and increased C_3 vegetation in southern Missouri, central USA. Oxygen isotopic shifts recorded in ON-3-B appear to parallel those in the GISP2 sequence but demonstrate that cooling in central North America was not as rapid or as pronounced as that recorded in Greenland ice.

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