Climatic backdrop to the terminal Pleistocene extinction of North American mammals

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ABSTRACT

North American terminal Pleistocene mammal extinctions are the subject of a long-running scientific debate. Although the role of climate has figured centrally, we lack clear knowledge of the timing and nature of terminal Pleistocene climate variability. Herein we document lengthy terminal Pleistocene drought in the southwestern United States (USA) using $\delta^{13}C$ and $\delta^{234}U$ effective moisture proxy data in speleothem calcite (stalagmite FS2) from Fort Stanton Cave, New Mexico, supplemented with age data from pool basin shelfstone speleothems from the Big Room in Carlsbad Cavern. This terminal Pleistocene drought, defined by a sharp rise in both $\delta^{13}C$ and $\delta^{234}U$ values, began just before 14.5 k.y. ago and lasted at least until 12.9 k.y. ago, when it was briefly and only mildly interrupted by the Younger Dryas. The timing and length of this drought (~1500 yr) match the Northern Hemisphere Bølling-Allerød oscillation preserved in Greenland ice cores and exhibited in the $\delta^{18}O$ record of stalagmite FS2. Rapid transition from cool moist Late Glacial to warm dry Holocene-like climatic conditions was likely unfavorable to many species of Pleistocene mammals in the southwestern USA. A climate-induced extinction implies that this last glacial cycle and its termination were more extreme than previous glacial cycles and/or glacial terminations.

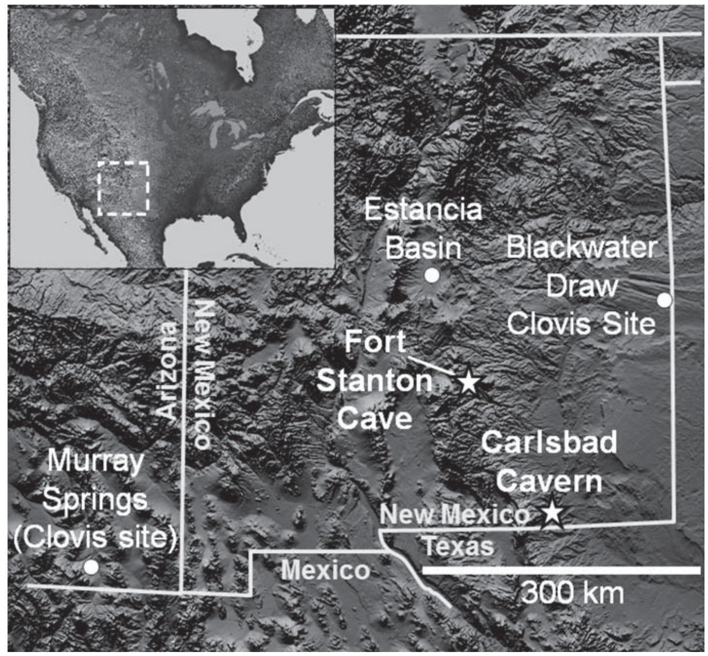

Figure 1. Location of Fort Stanton Cave relative to other important paleoclimate and/or cultural sites in southwestern United States (USA). Estancia Basin (to north) and Carlsbad Cavern (to south) yield results that support those from stalagmite FS2 from Fort Stanton Cave. Murray Springs and Blackwater Draw are two significant Clovis cultural sites where Clovis-aged drought, and Younger Dryas–aged black mats and cometary evidence, were reported (Haynes, 1991; Haynes et al., 1999; Firestone et al., 2007). Fort Stanton Cave is ideally situated for paleoclimate interpretations applicable to early human culture and terminal Pleistocene mammal extinctions in southwestern United States. (Image courtesy of NASA Jet Propulsion Laboratory at the California Institute of Technology.)

INTRODUCTION

The late Pleistocene–Holocene transition was a time of profound environmental and cultural changes in North America. Earliest human occupation is best documented during this time (Waters and Stafford, 2007; Waters, 2011), roughly coincident with the extinction of 35 genera of mostly large mammals, 16 of which have last appearance dates occurring within a period from 13.8 to 11.4 k.y. ago (Grayson and Meltzer, 2003; Faith and Surovell, 2009). The possible roles that environmental and cultural changes played on mammal extinction are still intensely debated globally (e.g., Barnosky et al., 2004). Climatic-induced causes for extinction include slow change from mosaic vegetation to a more zonal pattern (Guthrie, 1984), rapid cooling during the Younger Dryas (YD) cold event (Berger, 1991), and the Clovis drought (Haynes, 1991; Haynes et al., 1999; Polyak et al., 2004). It was suggested that both the Pleistocene mammal extinction and the YD climate event were caused by cometary impact (Firestone et al., 2007). Others have argued for human overhunting; variations of the "blitzkrieg hypothesis" (Martin, 1967). Much of the discussion of late Pleistocene North American mammalian extinction to date has taken place in the absence of clear understanding of extent and time scale of moisture variability. For example, rapid cooling during the YD has been blamed as a cause of extinction, but the YD was probably a period of greater effective moisture in the southwestern (USA) and therefore probably favorable for fauna (Haynes, 1991; Polyak et al., 2004; Haynes et al., 2010). Any conclusion about the role of climate on North American mammalian extinction has to be ultimately based on high-resolution data that give a sense of the timing and magnitude of moisture variability.

A late Pleistocene high-resolution $\delta^{18}O$ isotope record from Fort Stanton Cave (stalagmite FS2), New Mexico (Fig. 1), shows a remarkable positive comparison with the $\delta^{18}O$ isotope record from the Greenland ice cores (Greenland Ice Core Chronology 2005, GICC05) (Asmerom et al., 2010; Svensson et al., 2006). While this similarity with GICC05 indicates a teleconnection between the Pacific source of moisture for western North America and the North Atlantic, the fundamental question of effective moisture amount still remains, because the correlation between $\delta^{18}O$ data and GICC05 primarily reflects variability in moisture sources and is strongly weighted to changes in winter precipitation (Asmerom et al., 2010). Addressing terminal Pleistocene mammal extinction requires highly resolved proxies that yield records of effective moisture and vegetation.

In contrast to the oxygen isotope time series, which we interpret as an indicator of moisture source, carbon isotope variability reflects local environmental conditions responding to regional climatic changes. Drier climatic conditions in the southwestern USA likely result in lower soil productivity, slower infiltration rates with greater bedrock carbon contribution (higher $\delta^{13}C$ values relative to soil-derived carbon), possible change from C3 to C4 vegetation, and enhanced kinetic fractionation, all of which produce higher speleothem calcite $\delta^{13}C$ values (Frisia et al., 2011). Effectively wetter climatic conditions favor more vegetation, thicker soils, changes from C4 to C3 plants, and more productive soil CO_2 production, all producing more negative $\delta^{13}C$ values. In conjunction with carbon isotopic data, we also have obtained a novel independent proxy for precipitation

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based on the variation in the $^{234}U/^{238}U$ ratio, expressed as the per mil (‰) variation of the $^{234}U/^{238}U$ atomic ratio of a sample from the secular equilibrium ratio, which is equal to the inverse ratio of the decay constants (λ) of ^{234}U and ^{238}U ($\delta^{234}U = ([^{234}U/^{238}U / (\lambda_{238}/\lambda_{234})] - 1) \times 1000$ of speleothem calcite). $\delta^{234}U$ in speleothems as a climate proxy for groundwater-bedrock interaction times has been suggested (Ayalon et al., 1999; Zhou et al., 2005; Griffiths et al., 2010), and is reported to have potential as a long-term record of past precipitation (Ayalon et al., 1999). The decay of ^{238}U to ^{234}U (through an intermediate step of decay to ^{234}Th) results either in direct injection of ^{234}U nuclide from crystal surface to water or weakening of the daughter ^{234}U crystal site, making it susceptible to selective weathering by water. Thus, calcite precipitating from such water will typically have $\delta^{234}U > 0$, especially in more arid regions such as the southwestern USA. As a result, speleothem calcite $\delta^{234}U$ values corrected to their initial values can directly reflect the extent of water-rock interaction. Change from wet to dry climate regimes likely results in slowing of water infiltration rates through the bedrock. These higher water residence times can produce longer water-rock interaction, resulting in more ^{234}U relative to ^{238}U in the water, and vice versa. Advances in multicollector inductively coupled plasma mass spectrometric analysis of uranium-series isotopes from small samples (Asmerom et al., 2006) has now made it possible to get high-resolution $\delta^{234}U$ data. One of the significant advantages of this isotope system is that it is not susceptible to kinetic fractionation. Our results demonstrate potential usefulness of this relatively new proxy.

In addition to $\delta^{13}C$ and $\delta^{234}U$, we present chronologic data on speleothems representing cave pool basin highstands from the Big Room in Carlsbad Cavern (170 km south-southeast of Fort Stanton Cave; Fig. 1; Fig. DR1 in the GSA Data Repository[1]). We use this record, which is a measure of effective moisture, in combination with the period of highstands of Pleistocene Lake Estancia (Allen and Anderson, 1993, 2000) to provide support for our high-resolution stalagmite isotopic proxy. Additional isotope and age data are in Tables DR1–DR4 and Figures DR1–DR3.

RESULTS

Figure 2 shows large changes in the $\delta^{13}C$ and $\delta^{234}U$ records from the Late Glacial to the Pleistocene-Holocene transition. The carbon isotope system for caves in drier areas such as the southwestern USA seems to drive $\delta^{13}C$ values higher during drier climate and vice versa, with seemingly few if any other scenarios resulting. However, carbon isotope signatures in speleothems are not fully understood, and independent proxies such as our $\delta^{234}U$ record supporting climatic interpretations are recommended (Frisia et al., 2011). Stalagmite FS2 $\delta^{13}C$ values are relatively stable at −5‰ to −4‰ from 24 to 18 k.y. ago (Fig. 3; Table DR2), but exhibit a large variation (~−7‰ to ~−1.5‰) from 18 to 12 k.y. ago not likely related to kinetic fractionation in the cave (Asmerom et al., 2010). Stalagmite FS2 $\delta^{234}U$ values during the Last Glacial Maximum (LGM), a period of lake highstands with cool and wet climatic conditions prevailing throughout the western United States (Smith and Street-Perrott, 1983; Allen and Anderson 1993, 2000), vary from 500‰ to 550‰ until ca. 18.5 k.y. ago, when a sharp rise of ~60‰ in the $\delta^{234}U$ occurs (Fig. 3; Fig. DR3B). We interpret the higher $\delta^{234}U$ values to reflect a lower water/rock ratio tied to drier surface climate; therefore the rise in the $\delta^{234}U$ values from 18.5 to 17 k.y. ago represents a drier surface climate, also recorded as the first post-LGM lowstand of Pleistocene Lake Estancia (Allen and Anderson, 1993, 2000), referred to as the Estancia Big Dry (Broecker et al., 2009). The $\delta^{13}C$ trend from 18.5 to 17 k.y. ago does not rise with the $\delta^{234}U$, probably reflecting differences in sensitivities of the two proxies. The combination of the two, $\delta^{13}C$ and $\delta^{234}U$, are more useful for interpreting wetter and drier climate. Both the $\delta^{13}C$ and $\delta^{234}U$ values drop at 16.5 k.y. ago, reflecting the return of wetter conditions coinciding with the Heinrich stadial event 1 (HS1) Northern Hemisphere cooling excursion (Fig. 3) and Late Glacial Pleistocene Lake Estancia highstands. At 14.5 k.y. ago, both the $\delta^{13}C$ and $\delta^{234}U$ values increase markedly immediately following the beginning of the Bølling oscillation (Figs. 2 and 3). Carlsbad Cavern Big Room pool basins, periodically full during the LGM and Late Glacial, dry up at that time, and never return to fullness (Fig. 3; Fig. DR1). This period is also recorded as the first desiccation of Pleistocene Lake Estancia (Allen and Anderson, 1993), and is coincident with the initiation of drying interpreted from significant changes in vegetation preserved in south-central New Mexico packrat middens (Betancourt et al., 2001).

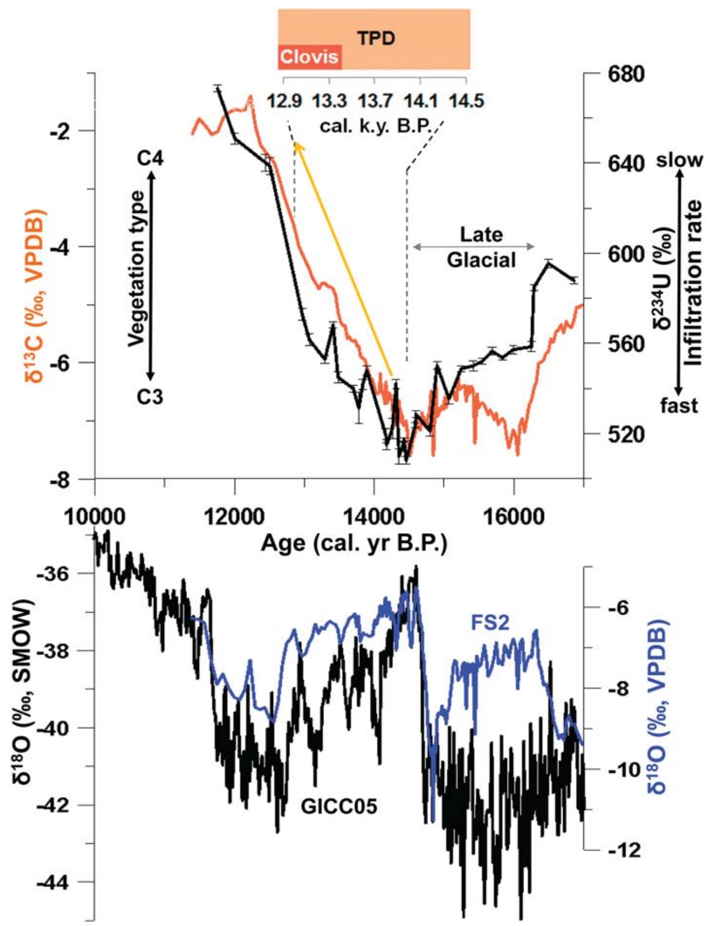

Figure 2. Isotopic profile of stalagmite FS2 (from Fort Stanton Cave) showing evolution of $\delta^{13}C$ (brown curve) relative to $\delta^{234}U$ (black dashed curve) from Late Glacial to beginning of Holocene. Periods of slow infiltration of vadose groundwater from surface into cave are expressed as higher $\delta^{234}U$, and slope along which $\delta^{234}U$ values increase (gold arrow) defines a significant dry episode at terminal Pleistocene. Changes in $\delta^{13}C$ reflect changes in vegetation and/or soil driven by climate shifts and follow the $\delta^{234}U$ curve. This southwestern United States terminal Pleistocene period of drought started ~1 k.y. before well-established evidence of Clovis occupation. Late Glacial to Holocene shift in $\delta^{13}C$ and $\delta^{234}U$ began soon after initiation of Bølling-Allerød ca. 14.8 k.y. ago (lower curve). Comparison of Greenland ice core (Greenland Ice Core Chronology 2005, GICC05) (Svensson et al., 2006) and Fort Stanton Cave FS2 $\delta^{18}O$ (Asmerom et al., 2010) records shows that rapid shift took place synchronously within errors of each record. FS2 $\delta^{18}O$ represents source of moisture (Gulf of Mexico or Pacific) into cave. GICC05 $\delta^{18}O$ values are reported with respect to standard mean ocean water (SMOW), and FS2 $\delta^{18}O$ values are reported with respect to Peedee belemnite (PDB). TPD—terminal Pleistocene drought.

[1]GSA Data Repository item 2012286, Figures DR1–DR3 and Tables DR1–DR4, is available online at www.geosociety.org/pubs/ft2012.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

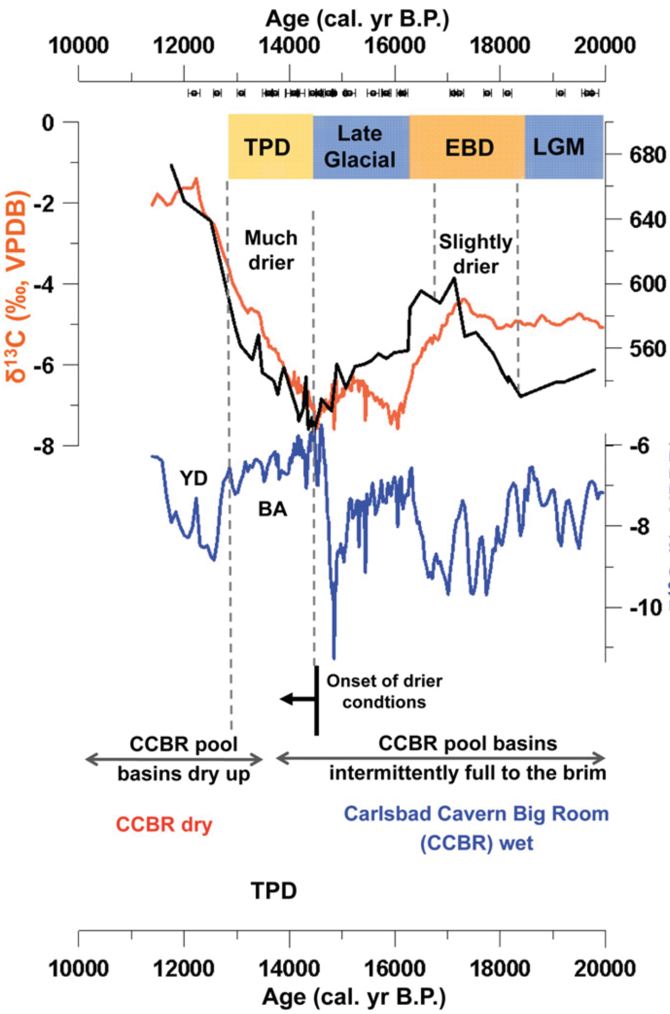

Figure 3. Stalagmite FS2 $\delta^{13}C$ (from Fort Stanton Cave; brown curve) and $\delta^{234}U$ (black) in relation to climatic events from Last Glacial Maximum (LGM) to Holocene. Estancia Big Dry (EBD; D zone, Allen and Anderson, 2000; Broecker et al., 2009) and Late Glacial are well characterized by our data. We used Fairbanks et al. (2005) to compare FS2 U-series and ^{14}C-based Estancia Basin chronologies. Dashed lines show two periods when $\delta^{234}U$ values rose significantly, roughly equivalent in time to two dry periods, EBD and Clovis drought. Period of drying at end of Late Glacial, terminal Pleistocene drought (TPD) in southwestern United States, is well represented by shelfstone samples from 7 pool basins in Carlsbad Cavern showing that pool basins in Big Room dried up by 13.5 k.y. ago (Fig. DR3 [see footnote 1]), consistent with end of Late Glacial defined by $\delta^{234}U$ and $\delta^{13}C$ values (see Fig. 2). Overall, onset of drying at terminal Pleistocene is synchronous with start of Bølling-Allerød (BA; bottom curve is FS2 $\delta^{18}O$ for comparison) (Asmerom et al., 2010). Source of moisture represented by FS2 $\delta^{18}O$ responded quickly in synchronization with North Atlantic climate, while regional climatic conditions represented by both $\delta^{234}U$ and $\delta^{13}C$ responded more slowly. U-series ages and their errors are shown along top *x*-axis. YD—Younger Dryas.

DISCUSSION

The coherency in multiple proxies, both physical and chemical, gives us confidence in our inferences about moisture variability during the time of interest. We interpret the large shift in $\delta^{13}C$ from 14.5 to ca. 12 k.y. ago as a change from Late Glacial C3 vegetation to Holocene C4 vegetation and/or decrease in organic component of speleothem carbon via lower soil respiration rates (Quade et al., 1989) and added bedrock carbon, and the corresponding large 150‰ shift in $\delta^{234}U$ to significantly slower vadose groundwater infiltration, both driven by a rapid change from very moist glacial-like to very dry Holocene-like conditions. The combined isotopic, pool water level, and lake highstand data together with the growth record of stalagmite FS2 point to unprecedented change in the moisture budget, the severity and rapidity not present to that point in the last 46 k.y. of the Pleistocene (Asmerom et al., 2010).

Stalagmite FS2 $\delta^{13}C$ and $\delta^{234}U$ data show precisely the beginning of sharply drier conditions in the southwestern USA starting at 14.5 k.y. ago and lasting for ~1500 yr, when this dry period was briefly and only mildly interrupted by wetter and cooler climatic conditions during the Younger Dryas (Polyak et al., 2004). We interpret this period as a terminal Pleistocene drought in the southwestern USA, and as a rapid change from exceptionally moist conditions to particularly dry Holocene-like conditions, as evidenced by western lake records (Smith and Street-Perrott, 1983; Benson et al., 1990), Carlsbad Cavern pool basin highstands, and the nearby Estancia Basin record (Allen and Anderson, 2000). Drier conditions at that time probably extended beyond the southwestern USA. Charcoal data from across the continent suggest rapid change to arid conditions around this time (Marlon et al., 2009), not unlike recent megadroughts that seem to be continental in scale (Cook et al., 2004). Following a long glacial-driven pluvial period, this transition to sustained aridity has to count as a significant environmental factor in the late Pleistocene extinction of mammals. The last appearance of 16 out of 35 genera between 13.8 and 11.4 k.y. ago (Faith and Surovell, 2009) show that this event happened during the terminal Pleistocene drought defined by our results. Other data push back the decline of large mammals to ca. 14.2 k.y. ago (Gill et al., 2009). Although difficult to establish causality, there is strong coincidence between this distinct change from moist to arid conditions and North American mammalian extinction. The last glacial was characterized by more extreme climate than previous glacials (Asmerom et al., 2010; Cheng et al., 2009). Such extreme climate variability might have preconditioned the megafauna for final extinction during the terminal Pleistocene drought starting ca. 14.5 k.y. ago. Climate change as a cause of mammalian extinction does not rule out the contribution of other factors during any interval of the extinction episode. Most extinctions in the geologic past, by necessity, were not caused by humans. However, we do not rule out human contribution to Pleistocene extinction. Whether climate-caused rapid change of flora and fauna resulting in food-source scarcity would lead to hunting of large mammals and thus provide positive feedback is an arguable scenario. The size selection does not require human agents. It has been suggested that large animals have dual vulnerability for extinctions, i.e., external environmental factors and inherent species traits having to do with decrease in productivity with body size (Cardillo et al., 2005).

The most compelling argument linking extinction to aridity rests on the timing of North American mammalian extinction and that of a potential cause or causes. Here we have presented the climate context for the period of interest, showing broad overlap between a pronounced postglacial drought and the period of mammalian extinction. Both the cometary impact (Firestone et al., 2007) and human overhunting (i.e., the blitzkrieg hypothesis; Martin, 1967), suffer from this lack of event-cause overlap. Even with evidence (Waters, 2011) of pre-Clovis human presence in the continent, it is unlikely that human footprint was heavy enough to have been the cause of the early extinction 13.8 k.y. ago (Faith and Surovell, 2009). Even though it is difficult to demonstrate causality, environmental stress as the cause of extinction, in this case driven by drought, is a type of extinction probably common in the geologic past.

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