Millennial-scale variations in western Sierra Nevada precipitation during the last glacial cycle MIS 4/3 transition


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ABSTRACT

Dansgaard–Oeschger (D–O) cycles had far-reaching effects on Northern Hemisphere and tropical climate systems during the last glacial period, yet the climatic response to D–O cycles in western North America is controversial, especially prior to 55 ka. We document changes in precipitation along the western slope of the central Sierra Nevada during early Marine Oxygen Isotope Stages (MIS) 3 and 4 (55–67 ka) from a U-series dated speleothem record from McLean’s Cave. The timing of our multi-proxy geological dataset is coeval with D–O interstadials (15–18) and stadials, including Heinrich Event 6. The McLean’s Cave stalagmite indicates warmer and drier conditions during Greenland interstadials (GISs 15–18), signified by elevated δ18O, δ13C, reflectance, and trace element concentrations, and less radiogenic 87Sr/86Sr. Our record extends evidence of a strong linkage between high-latitude warming and reduced precipitation in western North America to early MIS 3 and MIS 4. This record shows that the linkage persists in diverse global climate states, and documents the nature of the climatic response in central California to Heinrich Event 6.

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Introduction

Irregular, millennial-scale climate fluctuations known as Dansgaard–Oeschger (D–O) cycles are prominent features of Marine Oxygen Isotope Stages (MIS) 4, 3, and 2 (73.5–17.8 ka). First identified in the Greenland ice-core records (e.g., Dansgaard et al., 1984; Johnsen et al., 1992), D–O cycles are characterized by abrupt temperature increases of up to 10°C in as little as 50 yrs, followed by gradual cooling, and terminating in brief periods of accelerated cooling and cold conditions in the North Atlantic (Steffensen et al., 2008; Wolff et al., 2010). The most prominent of these abrupt terminal coolings are associated with Heinrich Events, large discharges of icebergs that are inferred from an influx of detrital grains in North Atlantic sediment cores (Heinrich, 1988).

High-resolution paleoclimatic proxy records document the response of the Intertropical Convergence Zone (ITCZ) (Peterson et al., 2000) and East Asian (Wang et al., 2001; Cheng et al., 2009) and South American monsoons (Wang et al., 2007; Kanner et al., 2012) to D–O cycles and Heinrich Events across MIS 2 through 4 and beyond. However, few continuous records with sufficient resolution to capture D–O variability have been developed for terrestrial mid-latitude regions outside the direct influence of the ITCZ and monsoon systems (Voelker and workshop participants, 2002; Genty et al., 2003; Harrison and Sanchez Goñi, 2010; Jo et al., 2014). Two speleothem δ18O records from the southwestern United States reveal a strong regional correlation with D–O events during MIS 2 and 3, interpreted to reflect elevated winter precipitation during Greenland stadials (cold periods) and reduced winter precipitation during interstadials (warm periods) (Asmerom et al., 2010; Wagner et al., 2010). Less is known about the relationship between D–O cycles, Heinrich events and western North American climate prior to MIS 3 (~55 ka), a time period that is especially difficult to investigate given that it is beyond radiocarbon age control. Lake records from the Great Basin suggest shifts to colder and wetter conditions during Heinrich Events, including the oldest Heinrich Event 6 (H6) (Bischoff and Cummins, 2001; Jiménez-Moreno et al., 2007). In contrast, the Devils Hole δ18O record, which apparently has resolution adequate to capture D–O events during MIS 3 and 4, does not display millennial-
scale variability consistent with D–O cycles and Heinrich Events (Winograd et al., 2006). Thus, it remains unclear how (and if) climate in western North America varied in response to these events prior to 55 ka.

Stalagmites from caves on the western slope of the central Sierra Nevada provide proxy records of past precipitation (Oster et al., 2009, 2010) with sufficient continuity and temporal resolution to assess climatic responses to D–O cycles and Heinrich Events. Here, we present U-series-dated records of growth rate, reflectance, stable isotope and $^{87}$Sr/$^{86}$Sr ratios, and trace elements from a stalagmite from McLean’s Cave, California (Fig. 1), a ~290 mm segment of which grew from 67 to 55 ka. Stable and radiogenic isotope and reflectance records from McLean’s Cave share many features coincident with millennial-scale climate events, including D–O cycles, documented in time-equivalent Greenland ice-core records (Dansgaard et al., 1984; NGRIP Members, 2004), Cariaco Basin sediments (Peterson et al., 2000), and Chinese, Brazilian, and European speleothems (Wang et al., 2001, 2004, 2007; Genty et al., 2003). Our new record also documents a significant climate response in central California to H6 and provides new constraints on the timing and duration of the effects of H6 in the mid latitudes. The McLean’s Cave stalagmite provides the first precisely dated record of the dynamic and variable climate in western North America that prevailed during MIS 4 and early MIS 3.

Site and sample description

McLean’s Cave (Fig. 1) is developed within the Columbia carbonate lens, one of several discrete metamorphosed pre-Jurassic limestone and dolomite masses in the Sierra Nevada foothills that are tectonically intercalated with other metasedimentary and metavolcanic rocks of the Calaveras Complex (Clark and Lydon, 1962; Bowen, 1973). The Columbia carbonate lens is oriented NE–SW and is 6.5 km long by 2.4–4 km wide (Clark and Lydon, 1962). The cave entrance, which is 300 m above sea level (asl), is located near the bottom of the South Fork Stanislaus River canyon. Speleothem ML-2, a ~400 mm long stalagmite, was collected in 1979 from the upper passage of McLean’s Cave, below 30–60 m of carbonate bedrock (Fig. 1C), just prior to cave inundation following construction of the New Melones Dam.

The climate above McLean’s Cave is characterized by cool, wet winters and warm, dry summers. From 1951 to 2011, a University of California weather station located 11.2 km southeast of McLean’s Cave at 533 m asl, experienced an average annual precipitation of ~850 mm. On average, 93% of this precipitation occurred between October and April. Only 2% of annual precipitation fell between July and September, suggesting minor influence by localized summer convective storms. McLean’s Cave is not significantly influenced by precipitation related to the North American Monsoon (Higgins et al., 1999).

Figure 1. Location of McLean’s Cave (ML; 38°4.20′N, 120°25.20 W; elevation of 300 m asl) in the western United States (A) and the western foothills of the Sierra Nevada (B) with location of Black Chasm Cavern (BL; Oster et al., 2012) and Moaning Cave (MC; Oster et al., 2009). (C) Cross-sectional view of McLean’s Cave (adapted from McEachern and Grady, 1978). Star shows location of sample ML-2 at time of collection. (D) Cross-sectional location of the cave in relation to level of the Stanislaus River prior to construction of the New Melones Dam.
Daily average temperature is 22.7°C during the summer (June–September) and 7.6°C during the winter (November–February). Present vegetation in this area consists of C3 plants, including manzanita (Arctostaphylos sp.), toyon (Heteromeles arbutifolia), and other chaparral species, and there is no evidence of transitions to C4 plant assemblages during the Pleistocene in this region (Cole, 1983; Davis, 1999).

Methods

U-series

Twelve U–Th dating samples were drilled from ML-2 following visible growth bands (Fig. 2). A ~1 mm wide dark band signifies a hiatus 85 mm below the top of the stalagmite. Dating of material on either side of this dark band indicates a growth hiatus occupying most of the glacial period between 55.7 ± 0.52 and 16.7 ± 2.55 ka. More recent calcite growth above this hiatus is porous and incorporates abundant detrital material, so dating efforts and geochemical analyses were focused on the older material.

Analyses were performed at the Berkeley Geochronology Center first using a Micromass Sector-54 Thermal Ionization Mass Spectrometer (TIMS) and then a Thermo NEPTUNE Plus Multi-Collector-Inductively-Coupled-Mass-Spectrometer (MC-ICP-MS). Samples of ~750–3300 mg for TIMS and ~200 mg for MC-ICP-MS were totally dissolved in 7 N HNO3 and equilibrated with a mixed spike containing 229Th, 233U, and 236U. The spike was calibrated against solutions of NBL CRM 145 and solutions prepared from a 69 Ma U ore that has been demonstrated to yield concordant U/Pb ages (Schwartzwalder Mine, Colorado, USA; hereafter, SM; Ludwig et al., 1985) and sample-to-sample agreement of 234U/238U and 230Th/238U ratios. U and Th were separated using two stages of HNO3–HCl cation exchange chemistry followed by reaction with HNO3 and HClO4 to remove any residual organic material. MC-ICP-MS analyses followed the procedures outlined in Mertz-Kraus et al. (2012). Measured peak heights were corrected for multiplier dark noise/Faraday baselines, background intensities, ion counter yields, tail contributions, and interfering spike isotopes. Mass fractionation was determined using the gravimetrically determined 231U/238U ratio of the spike. The external reproducibility of 234U/238U and 230Th/238U ratios of SM solutions measured during each run was better than 0.2%. Ages were calculated using the half-lives of Jaffey et al., (1971) for 238U, Holden (1990) for 232Th, and Cheng et al. (2000) for 230Th and 234U. Measured 230Th/232Th activity ratios, which scale with contamination by detritus, range from about 10 to 830 (median = 66), indicating that some samples require significant correction for U and Th from detritus. Correction was made assuming activity ratios of (232Th/238U) = 1.2 ± 0.6, (230Th/238U) = 1.0 ± 0.1, and (234U/238U) = 1.0 ± 0.1, which correspond to average silicate crust in secular equilibrium. An isochron constructed for five fractions of sample ML12, though somewhat scattered (MSWD = 6.5), yielded an age of 55.4 ± 1.2 ka. This age is in good agreement with the mean corrected age of this sample (55.70 ± 0.49 ka), indicating that the detritus correction is appropriate. Age uncertainties are stated at the 2σ level and include measurement errors as well as uncertainties associated with the initial isotope correction.

An age-depth model for ML-2 was constructed using the StalAge algorithm of Scholz and Hoffmann (2011). Comparison of StalAge and other modeling approaches using both synthetic and natural data sets, demonstrated that it performs well in dealing with complexities such

Figure 2. Photograph of stalagmite ML-2 (top) with sample locations for U-series dating marked by circles (closed circles: TIMS, open circles: MC-ICP-MS). U-series ages versus depth (bottom) with associated 2σ errors analyzed by TIMS (closed squares) and MC-ICP-MS (open squares). Age model for ML-2 was computed using the StalAge algorithm in R (Scholz and Hoffmann, 2011). Green line shows the final age model and red lines show associated 95% confidence limits.
as outliers, age inversions, large and abrupt changes in growth rate, and hiatuses that are found in many speleothem time series constrained by $^{230}$Th/U dating (Scholz et al., 2012). We selected StalAge for our study because it considers stratigraphic information in addition to ages and their assigned uncertainties, is objective, can be applied to data sets containing large changes in growth rate, performs well when applied to synthetic data sets where the “true” age model is known, and accounts for complexities by increasing the uncertainties assigned to the age model.

**Stable isotope compositions and reflectance measurements**

Using a binocular microscope and modified dental drill, 370 samples weighing ~10 μg each were taken at 1 mm intervals along the central growth axis of ML-2. Samples were baked at 100°C under vacuum for 30 min to drive off water vapor, reacted in 105% orthophosphoric acid at 90°C, and analyzed using either a PRISM or an OPTIMA gas source isotope ratio mass spectrometer at the University of California Santa Cruz. Analytical precision based on replicate analyses of in-house standard (CM) was better than ±0.05 and 0.1‰ (1σ) for C and O respectively. Oxygen and carbon isotope ratios are expressed relative to VPDB.

Reflectance data for ML-2 were obtained using high-resolution flatbed scanner imaging, sampling along the central growth axis of the stalagmite. Using the software NIH Image (http://rsb.info.nih.gov/nih-image/), a single pixel profile was generated along the growth axis of the ML-2 approximately 1 pixel width from the stable isotope sampling transect. This profile was converted to grayscale values where 0 represents black and the maximum value of 255 corresponds to white. Gray-scale values and variations were reproducible along multiple profiles within several pixels of the stalagmite growth axis.

**Strontium isotopes and trace elements**

Three to six mg of calcite from ML-2, sampled at 1–2 mm intervals, and host calcite and dolomite marbles taken from the Columbia Quarry directly across the South Fork of the Stanislaus River from McLean’s Cave, were processed for Sr isotope analysis. Strontium was isolated using Eichrom Sr Spec cation exchange resin (Horwitz et al., 1992). $^{87}$Sr/$^{86}$Sr ratios were measured in solution mode on a Nu Instruments Multi-Collector-Inductively Coupled Plasma Mass Spectrometer (MC-ICP-MS) in the Earth and Planetary Sciences Department at the University of California, Davis. The measured $^{87}$Sr/$^{86}$Sr ratio of NIST SRM 987 was 0.710251 ± 0.000025 (2σ, n = 30) during this study. All measured sample $^{87}$Sr/$^{86}$Sr ratios were normalized to a nominal value for NIST SRM 987 of 0.710248. A modern coral from the South China Sea was used as an in-house consistency standard. Its value for the measurement period was 0.709183 ± 0.000040 (2σ, n = 10), which falls within the range of modern seawater $^{87}$Sr/$^{86}$Sr (0.70916–0.70923) (DePaolo and Ingram, 1985; Palmer and Edmond, 1989).

Trace-element analyses conducted on a subsection of the ML-2 stalagmite (interval from 130 to 224 mm) were carried out using a New Wave Research UP-213 ablation system coupled to an Agilent 7500 series quadrupole ICP-MS at the Interdisciplinary Center for Plasma Mass Spectrometry, UC Davis. This interval was analyzed because it covers a prominent shift in stalagmite $\delta^{13}$C. We analyzed a polished thick section (~400 μm) at 70% laser energy and 10 Hz using a spot size of 20 μm and spacing of 150 μm. We used the multi-element synthetic glass NIST SRM 612 for calibration of the element concentrations, and the preferred values reported in the GeoReM database (http://geoem.mpch-mainz.gwdg.de/)(Jochum et al., 2011) as the “true” concentrations in the reference glass. Data were reduced using the program GLITTER (www.glitter-gemoc.com). USGS MACS-3, a synthetic carbonate base was used as a matrix-matched quality control material. Concentrations determined for USGS MACS-3 during this study were compared to the recommended values of the preliminary Certificate of the USGS (2010). The measured values for USGS MACS-3 agree within 11% for Mg, 4% for Sr, and 8% for Ba with the certified values. Precision (2σ, n = 71) on the MACS-3 measurements was 17% for Mg, 13% for Sr, and 11% for Ba throughout the course of the measurement period.

**Statistical tests**

In order to characterize the sign and strength of relationships among isotopic and geochemical proxies, Spearman rank-order correlation coefficients (“r_s”) and one-tailed tests of significance (“p”) between pairs of measured parameters were calculated using the statistics program R 2.15.1 (R Core Team, 2012). In order to calculate correlation coefficients between time-series of different resolution, proxy data were binned according to the median time step for the lower resolution data set and data within each bin was averaged. Linear trends were removed from the data sets prior to calculating $r_s$ values in order to evaluate shorter-term correlations between parameters without the bias of long-term linear trends.

To facilitate visual comparison, higher-resolution (trace element) time-series were smoothed using a Gaussian kernel density estimation, a robust method for comparison of irregularly sampled time-series (Rehfeld et al., 2011). Kernel smoothing bandwidths were selected using the Sheather-Jones method (Sheather and Jones, 1991) included in the SIZer package in R (Sondereregger, 2011).

**Results**

**U-series**

Analytical results and calculated ages are given in Table 1. Measured U concentrations were relatively low, ranging from 46 to 65 ppb (median = 53 ppb). U-series ages for the ML-2 stalagmite preserve stratigraphic order within dating errors (Table 1; Fig. 2). Corrected U-series ages for ML-2 below its growth hiatus range from 67.80 ± 1.5 to 55.70 ± 0.49 ka, indicating that our proxy records span the transition between MIS 4 and 3 at 59.44 ± 1.287 (Wolf et al., 2010).

**Stable isotope compositions and reflectance**

The McLean’s Cave stalagmite $\delta^{13}$C time-series exhibits systematic fluctuations between −11.8‰ and −5.9‰, while $\delta^{18}$O displays a smaller range of values between −11.0‰ and −8.1‰. Both datasets define similar trends (Fig. 3). Mickler et al. (2004, 2006) showed that non-equilibrium fractionation of stable isotopes can occur during speleothem growth due to rapid and extended degassing of CO₂ driven by decreased seepage-water flow rates, increased evaporation, and/or changing cave air–water CO₂ ratios. Such kinetic effects lead to $^{13}$C and $^{18}$O-enrichment in the speleothem and elevated concentrations of trace elements such as Mg, Sr, and Ba (Lorens, 1981). As McLean’s Cave has been inundated since 1979, it was not possible to collect modern calcite or drip water for calibration purposes. “Hendy Tests” have also been used to check for isotopic equilibrium during speleothem precipitation. A speleothem will pass the Hendy Test if $\delta^{18}$O values remain constant along a growth layer and there is no relationship between $\delta^{18}$O and $\delta^{13}$C along the growth axis (Hendy, 1971). Three series of stable isotope measurements were performed along ML-2 growth bands. Variability in $\delta^{18}$O along each band is 0.25, 0.32, and 0.63‰ respectively, and $\delta^{18}$O and $\delta^{13}$C are not significantly correlated along any of the bands (Supplementary Fig. 1). The results of the Hendy Tests suggest that the moderate down-axis correlation between $\delta^{18}$O and $\delta^{13}$C ($r_s = 0.56$; Table 2) is a result of environmental processes that influence both proxies independently rather than non-equilibrium calcite precipitation (e.g. Genty et al., 2003; Frappier et al., 2007; Oster et al., 2009).

McLean’s Cave stalagmite reflectance varies between grayscale values of 58 and 248, with higher values corresponding to brighter areas of the stalagmite, and lower values corresponding to darker...
Correlations with stalagmite growth rate (Figs. 3D,F) and display moderate positive relationships. Several sub-mm scale detritus-rich bands between 303 μm and 4 μm. Stalagmite [Sr] and [Ba] display a strong positive correlation with δ18O, and no correlation between [Sr] and [Ba]. δ13C, a moderate positive correlation with δ18O, and a moderate negative correlation with δ13C (Table 2).

**Strontium isotopes and trace elements**

McLean's Cave stalagmite δ57Sr/86Sr varies between 0.70715 and 0.70668 (Fig. 3). Calcite marble from the Columbia Quarry developed in the host carbonate unit has an average δ57Sr/86Sr value of 0.707233 ± 0.00009 (1σ, n = 3), whereas dolomite marble is less radiogenic, with an average δ57Sr/86Sr value of 0.706773 ± 0.00059 (1σ, n = 3).

McLean's Cave stalagmite trace element concentrations were measured for the interval 61.3 to 59.4 ka, across the MIS 4/3 transition where prominent shifts in stalagmite δ13C, δ18O, and reflectance occur. No analyses were made during the ~90-yr interval of 60.02 to 60.98 ka due to a break in the sample section used for trace element analysis. [Sr] for the sampled interval varies between 26 and 76 μg/g, [Ba] between 6 and 28 μg/g, and [Mg] between 800 and 3710 μg/g (Figs. 3 and 4). Stalagmite [Sr] and [Ba] display a strong positive correlation (Table 2) and exhibit shifts that are visually similar to stalagmite δ13C, most notably during the MIS 3 portion of the record younger than ~60 ka. [Mg] displays a weak negative correlation with [Sr] and no correlation with [Ba]. [Mg] also displays a strong positive correlation with δ13C, a moderate positive correlation with δ18O, and a moderate negative correlation with δ13C (Table 2).

**Discussion**

**Interpretation of McLean's Cave proxy records**

We interpret the McLean-2 isotopic, trace element, and reflectance time-series, to record variations in infiltration, soil respiration, water-rock interaction, temperature and sources of precipitation. Despite multiple influences on each proxy, these records point to significant variations in Sierra Nevada hydroclimate across the MIS 4/3 transition, during O cycles 15 through 18, and during additional abrupt millennial-scale events not recorded in NGRIP, but present in other records (e.g. Peterson et al., 2000; Wang et al., 2001; Boch et al., 2011a). For comparison, we use the NGRIP δ18O record plotted on the extended GICC05 chronology of Wolff et al. (2010) (Fig. 5G), which is generally accepted as the standard Greenland ice-core chronology. In this record, the onset of GIS17 and placement of the MIS 4/3 transition has an age of 59.44 ± 1.287 ka (1σ) (Wolff et al., 2010). We also compare the ML-2 record to absolute dated speleothem records from Hulu Cave in China (Wang et al., 2001) (Fig. 5D), and the Villars Cave and NALPS records from Europe (Genty et al., 2003; Boch et al., 2011a) (Fig. 5E), which display shifts synchronous with D-0 events. Correspondence between shifts in the Hulu Cave δ18O record and Greenland interstadials events (GISs) was originally determined against the GISP2 δ18O record, and the estimated error in the GISP2 timescale in this interval is ±10‰ (Stuiver and Grootes, 2000). Similarly, the age-model for the Cariaco Basin sediment reflectance record (Fig. 5F) has been tuned to the GISP2 timescale during this interval (Peterson et al., 2000), and this may account for the visible offset between the Cariaco Basin and the other records presented in Figure 5.

ML-2 stable isotopes and reflectance shifts display shifts synchronous with interstadial (Giss 16, 17, 18) and stadial events of MIS 4 and 3, that are documented in NGRIP, the Chinese and European stalagmite stable isotope records, and the Cariaco Basin reflectance record (Fig. 5). Transitions in the ML-2 proxy records that we interpret to be associated with GIS 17 occur at 59.48 ± 0.7 ka, within error of shifts in the NGRIP, Hulu, and Villars Cave records (Wang et al., 2001; Genty et al., 2003). Of the ML-2 proxy records, reflectance is most strongly correlated with NGRIP δ18O (r = 0.69, p < 0.001) (Table 3). GIS 16 and 17 are each represented by double peaks in the NGRIP δ18O record, and similar structures are also seen in the ML-2 reflectance record. Peaks in ML-2 stable isotope associated with GIS 16–18 exhibit an average positive shift of 2.4‰ in δ13C and 1.7‰ in δ18O. Increases of 2.1‰ in δ13C and
0.7‰ in δ¹⁸O beginning at 55.7 to 60 ka are synchronous with GIS 15 in the NGRIP record.

The oldest Heinrich Event, H6, occurs during the interval covered by the ML-2 proxy records. Its age is poorly constrained in marine sediment records, but has been estimated as 60 ± 5 ka (Hemming, 2004). Based on comparisons with ice core and speleothem records (Fig. 5), we suggest that climate perturbations relating to H6 began in the central Sierra Nevada at 60.9 ± 0.5 ka, when δ¹⁸O and δ¹³C show abrupt negative shifts of 1.5 and 2.1‰ respectively and growth rate substantially increases. H6 lasted ~1100 yr until 59.8 ± 0.6 ka when δ¹³Cs show an abrupt positive shift of 2.1‰, and growth rate decreases (Figs. 4, 5).

The ML-2 proxy records also display shifts that do not correspond to changes in NGRIP δ¹⁸O. These include increases in δ¹⁸O and δ¹³C at 56.5 ka just following GIS 16, increases in δ¹⁸O, δ¹³C, and reflectance at 65.2 ka preceding GIS 18, and multiple shifts in δ¹⁸O, δ¹³C, reflectance, and ⁸⁷Sr/⁸⁶Sr from 63.5 to 60.5 ka between GIS 18 and H6. Many of these shifts in the ML-2 proxy records correspond with changes in the Hulu, Cariaco Basin, and NALPS records, suggesting climate variations occurred in the midlatitudes that were not associated with Greenland temperature change.

In stalagmite ML-2, calcite that formed during Greenland interstadials (GIS 15–18) exhibits higher reflectance and is characterized by...
open columnar fabrics with abundant porosity along crystallite boundaries and, during GIS 15, laminae of microcrystalline calcite. Calcite that formed during stadials exhibits lower reflectance and consists of columnar calcite in which the crystallites have more completely coalesced leaving fewer inclusions along crystallite boundaries (Supplementary Fig. 2). Variations between open and porous and dense stalagmite crystal fabrics have been linked to changes in drip rate, water film thickness, and calcite supersaturation, as well as changes in cave ventilation. Open columnar calcite fabrics have been related to variable drip rates and CaCO$_3$ supersaturation (Frisia et al., 2000; Frisia and Borsato, 2010; Belli et al., 2013) and reduced water excess (Genty and Quinif, 1996) but also to higher cave air pCO$_2$ and low drip water pH that promote vertical growth of crystals over lateral coalescence (Boch et al., 2011b). Microcrystalline calcite fabrics are also indicative of variable

### Table 2

Spearman rank-order correlation coefficients ($r_s$ values) for ML-2 proxies. Proxy data has been binned where necessary (see text) and linear trends have been removed.

<table>
<thead>
<tr>
<th></th>
<th>$\delta^{18}$O</th>
<th>$\delta^{13}$C</th>
<th>$\frac{^{87}Sr}{^{86}Sr}$</th>
<th>Reflectance</th>
<th>Mg</th>
<th>Sr</th>
<th>Ba</th>
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<tr>
<td>$\delta^{18}$O</td>
<td>0.56**</td>
<td>0.29*</td>
<td>-0.02</td>
<td>0.68*</td>
<td>-0.02</td>
<td>0.17</td>
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<tr>
<td>$\delta^{13}$C</td>
<td>0.29*</td>
<td>0.42**</td>
<td>-0.02</td>
<td>0.12</td>
<td>-0.29**</td>
<td>-0.31**</td>
<td></td>
</tr>
<tr>
<td>$\frac{^{87}Sr}{^{86}Sr}$</td>
<td>0.04</td>
<td>-0.62*</td>
<td>0.12</td>
<td>-0.09</td>
<td>-0.9</td>
<td>-0.29</td>
<td></td>
</tr>
<tr>
<td>Reflectance</td>
<td>0.42**</td>
<td>0.78**</td>
<td>-0.62*</td>
<td>-0.29**</td>
<td>-0.16**</td>
<td>0.01</td>
<td></td>
</tr>
<tr>
<td>Mg</td>
<td>0.68*</td>
<td>0.12</td>
<td>-0.29*</td>
<td>-0.16**</td>
<td>0.87**</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sr</td>
<td>-0.02</td>
<td>0.17</td>
<td>0.31</td>
<td>-0.31**</td>
<td>0.87**</td>
<td></td>
<td></td>
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<tr>
<td>Ba</td>
<td>0.17</td>
<td>0.31</td>
<td>-0.29</td>
<td>0.01</td>
<td>0.87**</td>
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** $p < 0.001$.

* $p < 0.05$.

Figure 4. For the period 59.0 to 61.5 ka, records of (A) Mg, (B) Ba, and (C) Sr concentrations in ML-2. Thick solid lines are Gaussian kernel smoothing functions calculated for a bandwidth (h = 0.058) determined by the Sheather Jones method. (D) and (E) show $\delta^{13}$C and $\delta^{18}$O records for the same time period. Note gap in elemental concentration records (60.02–60.1 ka) due to sample damage.
drip rates and the presence of impurities in the drip water (Frisia and Borsato, 2010). More compact columnar calcite fabrics have been shown to precipitate under moderate, uniform drip rates and low CaCO₃ supersaturation (i.e., always wet at the growth site; Frisia et al., 2000; Frisia and Borsato, 2010); increased water excess (Genty and Quinif, 1996), as well as lower cave air pCO₂ (Boch et al., 2011b).

We interpret the more open columnar and microcrystalline fabrics, signified by increased reflectance during Greenland interstadials to suggest variable drip rates and CaCO₃ supersaturation. In contrast, the denser columnar crystals, signified by lower reflectance, during stadials are compatible with a steady water supply and more continuous calcite precipitation under low levels of CaCO₃ supersaturation.
These interpretations are broadly consistent with long-term changes in ML-2 growth rate, which is generally slower during the MIS 3 portion of the stalagmite (younger than ~60 ka) when calcite fabrics indicate variable water supply, and faster during the MIS 4 portion of the record when fabrics indicate a more continuous water supply (Fig. 3F). However, the rate of calcite precipitation at a speleothem growth site and in the epikarst above the cave (prior calcite precipitation, or PCP) has been shown to respond to both water supply and the pCO2 gradient between the overlying soil and cave (Banner et al., 2007; Kowalczk and Froelich, 2010). Thus changes in the degree of seasonality might lead to variations in ventilation and in turn, changes in speleothem growth rates.

Summer insolation at the latitude of McLean’s Cave increases throughout the time period recorded by ML-2, while winter insolation decreases, and from this we infer that seasonality increased through the course of the ML-2 record (Fig. 3A). Ventilation of McLean’s Cave was likely elevated during MIS 3 relative to MIS 4, and cave air pCO2 was likely reduced. However, stalagmite growth rates are lowest between 58.5 and 55 ka during MIS 3 (~3.5 mm/ka), suggesting that potential changes in ventilation between MIS 3 and 4 were not a dominant control on stalagmite growth rate and crystal fabric, in contrast to the relationships observed in Katerloch Cave, Austria (Boch et al., 2011b). We infer that changes in water supply primarily controlled the calcite fabrics displayed by ML-2 and that stadials were characterized by levels of CaCO3 supersaturation and steady water supply due to an overall wetter climate while interstadials were characterized by variable drip rates and episodic supply of water to the cave.

Variations in ML-2 reflectance and average growth rate are accompanied by correlative changes in δ13C, indicating shared environmental controls on these proxies (Table 2). Stadials are characterized by decreased reflectance and more negative speleothem δ13C values likely reflecting the 12C-enriched soil CO2 source in seepage waters that is less modified by degassing and PCP compared to interstadials. In regions that are not water limited such as the tropics, temperature is the primary influence on soil respiration rate and soil pCO2. However, in Mediterranean climates such as the western slope of the Sierra Nevada, soil moisture can provide the primary control on soil respiration rate and pCO2 (Orchard and Cook, 1983; Quade et al., 1989; Terhune and Harden, 1991; Reichstein et al., 2003). Provided that well developed soils occurred above McLean’s Cave, the more negative δ13C values during stadials could reflect increased soil respiration as a result of increased soil moisture. However, it is also possible that lower temperatures during stadials tempered the effect of increased moisture on soil respiration. Dense columnar calcite textures and the more negative δ13C values indicate that stadials were characterized by reduced PCP and continuous stalagmite growth facilitated by a steady water supply.

In addition to δ13C, PCP has also been demonstrated to increase trace element concentrations of drip waters and speleothems in the central Sierra Nevada (Oster et al., 2009, 2012). The very strong positive correlation between ML-2 Sr and Ba concentrations (Table 2) suggests that PCP controls these parameters. Although there is no statistically significant relationship between δ13C and Sr or Ba concentrations (Table 2), there are consistent trends among the data sets, most notably marked increases around 60.4 and 59.8 ka near the MIS 4/3 transition (Fig. 4). The lack of relationship could arise due to the required binning of the trace element data at lower resolution for comparison with δ13C, which could obscure finer-scale relationships between these proxies. However, the marked increases in all records near the MIS 4/3 transition and similar patterns of variability between δ13C, Sr, and Ba concentrations especially during MIS3 (Fig. 4) further indicate that PCP was enhanced in the McLean’s Cave system during MIS 3 relative to MIS 4.

Variations in the 87Sr/86Sr and Mg concentration of the ML-2 stalagmite are also linked to changes in hydroclimate. These proxies likely reflect varying degrees of dolomite dissolution by seepage waters in the epikarst. The Columbia marble unit that hosts McLean’s Cave consists of up to 50% dolomite at the Columbia Quarry near the site of McLean’s Cave (Bowen, 1973; E. Haughy, Blue Mountain Minerals, pers. comm. 2010). Slower weathering of dolomite than calcite has been documented within karst systems in the field (Cowell and Ford, 1980; Atkinson, 1983) and experimentally (Chou et al., 1989). Seepage waters tend to reach supersaturation with respect to calcite before they reach dolomite saturation (Roberts et al., 1998), and changes in seepage-water flow rate can cause variations in the amount of dolomite dissolved and seepage-water Mg-content (Fairchild et al., 2000; Hellstrom and McCulloch, 2000). Thus, slower flow rates during dry periods could have led to longer water–rock contact times and waters that were supersaturated with respect to calcite while remaining undersaturated with respect to dolomite.

Given that host dolomite 87Sr/86Sr values (0.706773 ± 0.000059) are less radiogenic than host calcite values (0.707233 ± 0.000009), the overall lower 87Sr/86Sr values of the ML-2 stalagmite during MIS 3 (Fig. 3E), coincident with intervals of more open columnar and microcrystalline fabrics signified by higher reflectance and less negative δ13C is consistent with slower and episodic flow through the epikarst to the drip-site, leading to increased dolomite dissolution during dry periods. Conversely overall more radiogenic stalagmite 87Sr/86Sr values during MIS 4, coincident with intervals of dense columnar fabrics signified by lower reflectance and more negative δ13C are consistent with decreased dolomite dissolution during wet periods when seepage-flow rates through the epikarst were faster and more continuous. The negative correlation between 87Sr/86Sr and Mg concentration (Table 2) further supports this relationship.

Host rock limestone has a slightly higher δ13C value than host rock dolomite (2.22 vs. 1.63 ± 0.03‰, n = 3, respectively). Thus, the positive correlation between speleothem δ13C and Mg concentrations (Table 2) does not indicate that δ13C is significantly influenced by dolomite dissolution. Rather, this correlation suggests that δ13C and Mg concentrations are independently controlled by changes in hydroclimate. Taken together, variations in ML-2 reflectance, growth rate, δ13C, trace elements, and 87Sr/86Sr indicate that stadials and MIS 4 more broadly were characterized by wetter climates with reduced PCP, less dolomite dissolution

<table>
<thead>
<tr>
<th>ML-δ18O</th>
<th>ML-δ13C</th>
<th>ML-reflectance</th>
<th>Hulu-δ18O</th>
<th>NGRIP-δ18O</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.51**</td>
<td>0.51**</td>
<td>0.53**</td>
<td>-0.54**</td>
<td>0.35*</td>
</tr>
<tr>
<td>0.46*</td>
<td>0.46*</td>
<td>0.46*</td>
<td>-0.74**</td>
<td>0.69*</td>
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<tr>
<td>-0.52**</td>
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<tr>
<td>0.16</td>
<td>0.16</td>
<td>0.16</td>
<td>0.69**</td>
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** p < 0.001.
* p < 0.05.
and steady water supply to the stalagmite. Conversely, interstadials, and MIS 3 more broadly were characterized by drier climates with increased PCP and dolomite dissolution and episodic stalagmite growth that resulted in overall lower growth rates. Oxygen isotope ratios in speleothems have often been interpreted to reflect changes in the δ18O of precipitation related to variations in rainfall amount (the “amount effect” of Dansgaard, 1964) (e.g. Wang et al., 2001, 2007; Kanner et al., 2012). However, recent studies have indicated that the source and trajectory of moisture reaching a cave site can have a greater influence on the δ18O ultimately recorded by speleothem calcite (e.g. Pausata et al., 2011; McCabe-Glynn et al., 2013). We monitored rain and drip water δ18O at Black Chasm Cavern (BL, Fig. 1B), a tourist cave located 45 km northwest of McLean’s Cave and found that rainfall water δ18O is significantly positively correlated with surface air temperature, but shows no significant relationship with rainfall amount (Oster et al., 2012). Additionally we found that changes in moisture source and trajectory do influence the δ18O of precipitation in the central Sierra Nevada. Storms that originate in the subtropical Pacific have significantly less negative δ18O values than North Pacific sourced storms (δ18O ≈ 6.4 ± 2.5‰ versus ~15.5 ± 0.4‰) (Oster et al., 2012). This finding is consistent with other regional studies of precipitation and speleothem δ18O variability (Berkelhammer et al., 2012; McCabe-Glynn et al. 2013). Given the observed relationships between modern climate conditions and precipitation δ18O, the δ18O record of the ML-2 stalagmite likely records changes in surface air temperature and variations in the contribution of subtropical water vapor to the central Sierra Nevada. However, speleothem δ18O can also be influenced by temperature changes at the site of calcite precipitation. Although flooding of the cave precludes measurements of modern internal air temperature variations, temperature has been monitored in other nearby caves. Annual temperature variation within Black Chasm Cavern (BL) is ~1.5 °C (Oster et al., 2012). Temperature variations in Kaweah, Clough, and Soldiers Caves, four wild caves similar in geometry to McLean’s Cave (single entrance, vertically oriented) located approximately 240 km to the southwest in Sequoia National Park at elevations of 480, 1260, and 1220 m asl respectively ranged from 0.7 to 1.2 °C over a 2 yr measurement period (2002–2004). If the temperature range of 1.5 °C observed in BL is taken as a likely maximum for temperature variability within McLean’s Cave, temperature change would only account for a 0.3% change in the δ18O of speleothem calcite (using the equilibrium fractionation relationship of Kim and O’Neil, 1997). We therefore interpret the ML-2 oxygen isotope values to primarily reflect the drip water δ18O, and thus the δ18O of precipitation falling above the cave, rather than temperature variability at the site of speleothem deposition (McDermott, 2004).

In addition to surface air temperature and moisture source, the δ18O of precipitation falling above McLean’s Cave during the last glacial cycle was also influenced by changes in continental ice volume that altered the δ18O of source waters (Dansgaard, 1964; Rozanski et al., 1993). Records from Pacific benthic foraminifera indicate that δ18O, decreased by ~0.3‰ across the MIS 4/3 transition (Lisiecki and Raymo, 2005). Such a shift may have slightly attenuated the amplitude of variability recorded by ML-2, as δ18O in this record increases across this transition. The range of variability across stadial–interstadial transitions in ML-2 (0.7 to 2.5‰) would suggest maximum surface air temperature changes of ~1 to 5 °C, using a mid-latitude relationship of 0.5–0.7‰ per °C change for the δ18O of precipitation proposed by Rozanski et al. (1993), if no other effects were influencing precipitation δ18O. As described, however, δ18O variability could also reflect shifts in vapor source to the central Sierra Nevada. It is possible that the proportion of subtropical moisture vapor reaching McLean’s Cave increased during interstadials and contributed to increased stalagmite δ18O during these warmer times. A sustained increase in the frequency of subtropical atmospheric river storms that presently deliver higher δ18O precipitation to California is predicted to occur in future global-warming scenarios (Dettinger, 2011). If a similar increase in subtropical storms occurred during warmer periods of the past, this would contribute to increased drip water and speleothem δ18O in central Sierra Nevada caves across the MIS 4/3 transition.

McLean’s Cave record and regional climate at the MIS 4/3 transition

Our interpretation of the ML-2 stalagmite record may be compared with existing western North American paleoclimatic records of the last glacial period. Stable isotope and trace element records from Owens Lake sediments in east-central California (Li et al., 2004) and a speleothem record from Carlsbad Caverns, New Mexico (Brook et al., 2006) both suggest that wetter conditions prevailed in the southwestern United States during MIS 4 relative to MIS 3, consistent with our findings. The Devils Hole calcite exhibits overall lower δ18O during MIS 4, with a local minimum occurring at ~631 ka (±0.64‰), within dating errors of a period of reduced δ18O, δ13C, and reflectance in the ML-2 stalagmite record. The decrease in the Devils Hole δ18O record is interpreted to reflect changes in regional δ18Orecord due to decreased SST off the California coast during MIS 4 (Winograd et al., 2006).

Speleothem records document wetter winters during the stadials of MIS 3 in the southwestern US (Asmerom et al., 2010; Wagner et al., 2010). While these stalagmite records begin near the onset of GIS 14, and thus do not overlap with the ML-2 record, our record indicates that the association of interstadials with drier conditions in the western US continued into early MIS 3 and MIS 4, a glacial period but with smaller northern Northern Hemisphere ice sheets relative to MIS 2 (Siddall et al., 2008). In contrast, stable isotope records from four Great Basin lakes suggest that D–O stadials were associated with drier conditions in this region between 27 and 50.5 ka (Benson et al., 2003). However, the chronologies for these lake records are complex, based on radiocarbon measurements on lake carbonates and correlation of ash layers and paleomagnetic features between basins (Benson et al., 2003), and some of these correlations may be problematic (Zimmerman et al., 2006). ML-2 proxy records are consistent with variations in foraminiferal assemblages in the Santa Barbara Basin during MIS 3 that suggest stadials off the California coast were associated with decreased SST and increased upwelling, which are attributed to increased northerly winds, a strengthened Aleutian Low, and a southward displaced North Pacific high-pressure cell (Hendy and Kennett, 2000). Strengthening of the Aleutian Low, which is associated with cyclones that pass along the jet stream, would also have led to wetter conditions in the central Sierra Nevada during stadials of MIS 3 and MIS 4.

Paleoclimate records from western North America show some disagreement as to the response of local hydroclimates to Heinrich Events. Pollen from Bear Lake on the Utah–Idaho border suggests Heinrich Events were characterized by cooler and wetter conditions during MIS 2–4 (Jiménez-Moreno et al., 2007), and rock flour abundance in Owens Lake suggests glacial advances occurred in the Sierra Nevada in association with Heinrich Events, including H6 (Bischof and Cummins, 2001). Many Great Basin lakes reached their most recent highstands during H1 (Munroe and Laabs, 2013), and oxygen and uranium isotope records from Lake Bonneville carbonates suggest wetter conditions prevailed during H2 (McGee et al., 2012). However, δ18O and TIC records from a Lake Bonneville core indicate drier conditions prevailed during H1–H4 (Benson et al., 2011). Shifts in the ML-2 proxy records during H6 are consistent with others that indicate that wetter conditions prevailed in western North America during Heinrich Events of MIS 2–4.

McLean’s Cave record in a global climate context

Comparison of the McLean’s Cave proxy records and the Greenland ice core (Fig. 5) suggests generally drier conditions during Greenland Interstadials (GIS 15–18) and wetter conditions during stadials in the Sierra Nevada. At the multi-centennial scale (250 yr time bins), ML-2 reflectance and δ13C display significant correlations with NGRIP
The Hulu Cave record of East Asian Monsoon variability (Wang et al., 2001; Liu et al., 2010) or Nevada during Greenland interstadials occurred contemporaneously with an enhanced East Asian (Wang et al., 2001; Liu et al., 2010) or Indian Monsoon (Pausata et al., 2011), and increased river runoff to the Cariaco Basin (Peterson et al., 2000), as well as decreased precipitation in northeastern and southern Brazil (Wang et al., 2004, 2007). The transition from H6 to GIS 17 was also associated with a marked temperature increase in western Europe as documented by the δ13C record of a Villars Cave stalagmite (Fig. SE) (Genty et al., 2003). In contrast, wetter conditions in the central Sierra Nevada during stadials were associated with decreased Asian Monsoon intensity, decreased river runoff to the Cariaco Basin, and increased precipitation in northeastern Brazil. Such changes in precipitation in South American and Asian monsoon systems can be explained by shifts in the position of the ITZC in both the Atlantic and the Pacific basins that were coincident with Greenland temperature changes (Wang et al., 2004, 2007; Peterson and Haug, 2006; Clement and Peterson, 2008; Kageyama et al., 2010). Climate models indicate that southward shifts in the ITZC can be initiated by freshwater input to the North Atlantic leading to slowing of meridional overturning circulation (Vellinga and Wood, 2002; Zhang and Delworth, 2005). A weakening of Atlantic thermohaline circulation has also been shown to lead to a stronger Aleutian low-pressure cell and increased precipitation in western North America (Okumura et al., 2009), as is suggested by the Santa Barbara Basin record for MIS 3 stadials (Hendy and Kennett, 2000). In particular, Heinrich Events provide evidence of freshwater inputs to the North Atlantic through the discharge of icebergs, and thus this mechanism could explain the precipitation increase inferred from the ML-2 proxy records during H6. Climate models generally do not successfully link the climatic shifts that occur during D–O cycles to changes in thermohaline circulation and, in particular, models have difficulty reproducing the abrupt warming events noted in Greenland temperature records (Clement and Peterson, 2008; Kageyama et al., 2010). In part, this may be attributable to differences in boundary conditions between MIS 3 and the pre-industrial and LGM climates traditionally used in freshwater forcing experiments (Van Meerbeeck et al., 2009). Variations in sea-ice extent may occur on short timescales and could generate the abrupt changes associated with D–O cycles (Li et al., 2005). Models suggest that, under modern boundary conditions, reductions in Arctic sea-ice extent can cause northward-shifted storm tracks and a drier western North America (Sewall and Sloan, 2004; Singarayer et al., 2006). It is possible that changes in sea-ice extent in the Arctic and North Pacific may have contributed to the abrupt shifts in precipitation amount inferred for the central Sierra Nevada during MIS 3 and 4. The inferred decrease in Sierra Nevada precipitation across the MIS 4/3 boundary is also approximately coeval with an increase in eustatic sea level noted in coral and deep-sea records between 60 and 57 ka (Siddall et al., 2008). This suggests that this precipitation decrease may have been associated with a decrease in the size and extent of the Laurentide Ice Sheet that, similar to predictions for Termination I (COHMAP Members, 1988; Bromwich et al., 2004; Kim et al., 2008), may have caused a northward shift in the mean position of the polar jet stream across the MIS 4/3 transition.

Conclusions

High-resolution, U-series dated records of multiple proxies from a stalagmite from McLean’s Cave, central California demonstrate that climate there became warmer and drier during the Greenland interstadials associated with D–O events of MIS 4 and early MIS 3 (GISs 15–18), and colder and wetter during Greenland stadials, including Heinrich Event 6. These results are consistent with those from a stalagmite from nearby Moaning Cave, which records a similar relationship between high northern latitude temperatures and Sierra Nevada precipitation between 16.5 and 8.7 ka (Oster et al., 2009), across a similar transition out of glacial conditions. Like the Moaning Cave record, the McLean’s Cave records highlight complex interactions between soil respiration, PCP, and water–rock interactions and emphasize the importance of multi-proxy development of speleothem records, especially in semi-arid environments. Despite their complexity, the McLean’s Cave stalagmite proxy records provide a reliably dated, continuous, relatively high-resolution snapshot of precipitation variability in western North America during an interval when few such climate records have been developed.

Comparison with global MIS 3 and 4 climate records indicates that the changes in precipitation documented by ML-2 are consistent with migration of the ITZC and variations in the westerly storm track synchronized with warming in the high northern latitudes. The McLean’s Cave record lends support to a growing body of evidence supporting links between high latitude warming and reduced precipitation in western North America.

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