Stalagmite records of hydroclimate in central California during Termination 1

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ABSTRACT

The last deglaciation is marked by large and abrupt hydroclimatic changes in the Great Basin and the American Southwest. However, comparatively little is known about how hydroclimate varied on the western side of the Sierra Nevada. We present new evidence for abrupt changes in precipitation amount in the central Sierra Nevada during the last deglaciation. Our new record from McLean's Cave overlaps with a previously published record from nearby Moaning Cave (Oster et al., 2009), and extends the record of hydroclimatic change in the region through Heinrich Stadial 1. McLean's Cave speleothem $\delta^{18}\text{O}$, $\delta^{13}\text{C}$, Mg/Ca, and Sr/Ca indicate a shift to drier conditions at the beginning of Heinrich Stadial 1 from 16.1 to 17.5 ka followed by wetter conditions at the end of Heinrich Stadial 1 when the majority of Great Basin lakes reached their deglacial highstands. During the last deglaciation, coincident shifts in the Moaning and McLean's Cave proxy records indicate drier conditions in the western Sierra Nevada during millennial-scale intervals of warming at high latitudes such as the Bölling and Alleröd, and wetter conditions during cooler intervals such as the Older and Younger Dryas. Thus, the Sierra Nevada speleothem records are consistent with other regional records that

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document increased winter rainfall during millennial-scale cold periods of the last deglaciation. However, regional differences exist in the hydroclimatic response to the Younger Dryas, with central Californian and interior southwestern sites indicating an increase in winter storms that is not apparent in southern California. Dynamic simulations of atmospheric circulation indicate that wetter conditions during the Younger Dryas relative to the Bölling at McLean's Cave may have resulted from a stronger storm track in the north Pacific at the latitude of northern California.

Introduction

The Late-Pleistocene to Holocene transition in the western United States was punctuated by rapid hydroclimatic changes that significantly influenced the geomorphology and ecology of this semi-arid region. Such changes include the growth of pluvial lakes to post-Last Glacial Maximum highstands (Munroe and Laabs, 2013), and changes in the amount and sources of rainfall as recorded in speleothem proxy records, alluvial fan aggradation, and pollen records from the California coast (Oster et al., 2009; Wagner et al., 2010; Asmerom et al., 2010; Lyle et al., 2012; Antinao and McDonald, 2013). Approximately synchronous shifts in several western US proxy records and time series of Greenland temperature indicate that robust teleconnections exist between the two regions. Likely mechanisms include changes in winter and summer insolation (Lachniet et al., 2014), the thickness and extent of continental ice sheets and sea ice (COHMAP Members, 1988; Hostetler and Benson, 1990; Oster et al., 2015), and variations in meridional temperature gradients (Asmerom et al., 2010) that in turn drove changes in atmospheric circulation and influenced the strength, position, and moisture sources of the winter westerly storm track in western North America.

Despite numerous and diverse proxy records of deglacial change in this region, questions remain about the spatial pattern of hydroclimatic change during this interval and its driving mechanisms. Disparities in the timing and sometimes the sense of hydroclimate change noted in various proxy records have led to multiple hypotheses about how the westerly storm track evolved from the Last Glacial Maximum to the Holocene, and the variable impacts of that evolution across the western United States (Hostetler and

Benson, 1990; Asmerom et al., 2010; Lyle et al., 2012). Given the presently marginal water resources in this region (Griffin and Anchukaltis, 2014), and the potentially dire though uncertain projections of hydroclimatic change expected with continued global warming (Seager and Vecchi, 2010; Diffenbaugh et al., 2013), understanding the pattern and causes of past hydroclimate change in the region is essential.

Stalagmites from caves on the western slope of the central Sierra Nevada provide proxy records of past precipitation and temperature changes during the last glacial cycle (Oster et al., 2009; 2010; 2014), including a multi-proxy record of precipitation change at decadal- to centennial-resolution from Moaning Cave over the range of 8.7 to 16.5 ka (Oster et al., 2009). Here, we present an improved chronology for the Moaning Cave record and a new multi-proxy record from a stalagmite from nearby McLean's Cave (ML-1) that grew from 12 to 19 ka. Sub-decadally to decadal-resolved stable isotope records from ML-1 reproduce centennial- to millennial-scale shifts evident in the lower-resolution Moaning Cave record and extend the central Sierra Nevada record to include Heinrich Stadial 1 (~14.5 to 18 ka). The two Sierra Nevada speleothem records indicate significant hydroclimate changes that are consistent with several other regional paleoclimate records (e.g. Wagner et al., 2010; Asmerom et al., 2010; Street et al., 2012) but display differences relative to others (e.g. Kirby et al., 2013). The hydrologic changes noted in the Sierra Nevada speleothem records are consistent with the results of recent models of changes in the strength of the westerly storm track on millennial timescales during the last deglaciation (Wong et al., *in revision*).

Site and Sample Description

Moaning Cave (MC) and McLean's Cave (ML) are developed in the same metamorphosed carbonate lens within the Calaveras Complex of the central Sierra Nevada foothills (Fig. 1). The cave entrances are located approximately 4 km apart. MC, situated at 520 m above sea level, is presently a tourist cave. The entrance to ML at 300 m above sea level lies below the historic level of the New Melones reservoir, though the entrance has recently reemerged in the present drought. Complete descriptions of both cave systems can be found in Oster et al., (2009) (for MC) and Oster et al., (2014) (for ML). Collection of the Moaning Cave stalagmite (MC3) has been described previously (Oster et al., 2009; 2010). The stalagmite from McLean's Cave used in this study (ML-1) is a 24 cm stalagmite that 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 dam construction.

The modern climate above both caves is characterized by cool, wet winters and warm, dry summers. On average, 90% of rainfall occurs between the months of October and April. These months are characterized by winter storms carrying moisture to the Sierra from the central and northern Pacific (Dettinger et al., 2004). However, storms that produce the most intense rainfall are often associated with atmospheric rivers, narrow bands of water vapor that can travel long distances in the lower atmosphere (Ralph and Dettinger, 2011). Such storms often tap vapor sources from the subtropical Pacific, and are frequently associated with flooding in southern and central California (Dettinger, 2011; Ralph and Dettinger, 2011).

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108 The tourist operation, which alters water flow within MC, and the perennial flooding of
109 ML, prevents monitoring of modern drip water chemistry and ventilation in these caves.
110 However, a study of the modern cave system was conducted over five years at a third
111 nearby cave, Black Chasm Cavern (BL) (Fig. 1b). BL is located 45 km to the northwest
112 of MC and ML and has a similar vertical, fracture-oriented geometry and should provide
113 a modern analogue to the response of central Sierra Nevada caves to environmental
114 change. The results of that study are detailed in Oster et al. (2012). We note here that BL
115 is ventilated during the cold winter months, during which drip rates also increase
116 substantially. Monthly measurements of rain and drip water $\delta^{18}\text{O}$ at Black Chasm Cavern
117 documented that rainwater $\delta^{18}\text{O}$ is significantly positively correlated with surface air
118 temperature, but shows no significant relationship with rainfall amount. Additionally we
119 found that changes in moisture source and trajectory do influence the $\delta^{18}\text{O}$ of
120 precipitation in the central Sierra Nevada. Atmospheric river storms that originate in the
121 subtropical Pacific have significantly less negative $\delta^{18}\text{O}$ values than North Pacific
122 sourced storms (-6.4 ± 2.5 ‰ versus -15.5 ± 0.4 ‰) (Oster et al., 2012). This finding is
123 consistent with other regional studies of precipitation and speleothem $\delta^{18}\text{O}$ variability,
124 including within California (Ersek et al., 2010; Berkelhammer et al., 2012; McCabe-
125 Glynn et al., 2013).

126

127 **Methods**

128 *Sample preparation and petrographic analysis*

A two cm-thick slab was cut from the face of one half of ML-1 using a diamond saw in order to obtain U-Th samples and a set of billets for preparation of thick sections. Eleven billets were cut in an overlapping offset pattern along the central growth axis in order to provide continuous sampling of the stalagmite for stable isotope ratios and trace element concentrations (Fig. S1). Thick sections (~200 μm) were studied petrographically under transmitted, plane-polarized, and epifluorescent (450-490 nm wavelength) light in order to identify banding and calcite fabrics and textures.

U-series dating

Ten samples for U-Th analysis were cut along visible stalagmite bands from mm-thick wafers of the McLean's Cave (ML-1) stalagmite. Four new samples were also taken from the Moaning Cave stalagmite (MC3) to improve age control on this proxy record at onset of the Bölling-Alleröd warm period. Analytical methods used for prior U-Th dating of MC3 are described in Oster et al. (2009). In the current study, all samples were sonicated in methanol and inspected under a binocular microscope prior to dissolution and fragments that contained visible detritus were discarded in order to minimize detrital Th. Samples were crushed, then completely dissolved in 7N HNO_3 and equilibrated with a mixed spike containing ^{229}Th , ^{233}U , and ^{236}U . The spike was calibrated using New Brunswick National Lab CRM145 uranium solution and solutions prepared from a 69 Ma U ore from Schwartzwald Mine, Colorado, USA that has been demonstrated to yield concordant U-Pb ages (Ludwig et al., 1985) and sample-to-sample agreement of $^{234}\text{U}/^{238}\text{U}$ and $^{230}\text{Th}/^{238}\text{U}$ ratios. U and Th were separated from sample matrix using two stages of HNO_3 -HCl cation exchange chemistry followed by reaction with HNO_3 and HClO_4 to remove any residual organic material. Procedural backgrounds for U and Th are

measured with each batch of samples. U and ^{232}Th procedural backgrounds were negligible relative to sample U and Th and ^{230}Th blanks averaged 0.00028 fmol, which we assigned an uncertainty of 50% for propagation into the final age.

Analyses were performed using a Thermo-Fisher NEPTUNE *Plus* multi-collector inductively coupled plasma mass spectrometer at the Berkeley Geochronology Center. U and Th fractions were analyzed simultaneously. Analyses were carried out in automated sequences, with each analysis consisting of a pre-analysis washout, a background measurement, a sample measurement, and a measurement of peak-tails. All measurements are normalized to simultaneous measurements of ^{238}U to cancel intensity fluctuations. Instrumental backgrounds at each U and Th mass, along with Faraday collector baselines and ion counter dark noise, were measured while aspirating the same dilute HNO_3 -HF solution used to introduce samples. Sample measurements were carried out in peak-jumping mode with all masses measured on Faraday collectors except ^{230}Th and ^{234}U , which were measured with an ion counter. Sample measurements incorporated within-run mass bias corrections using the known $^{233}\text{U}/^{236}\text{U}$ ratio of the spike. Faraday amplifier gains were calibrated weekly using the Neptune's constant current source and switching system. Ion counter yields (relative to Faraday collectors) were determined during each sample measurement by cycling ^{229}Th and ^{233}U between the ion counter and the L1 Faraday collector. Peak tails were monitored at seven half-mass positions from $m/e = 229.5$ to 241.5 while aspirating sample solution, and the sum of tails from all peaks was modeled at each U and Th mass position and subtracted from the measured intensities. Measured peak heights were corrected for Faraday baselines/ion counter dark

noise, instrumental backgrounds, Faraday gains, ion counter yields, mass fractionation, peak-tails, procedural blanks, and interfering spike isotopes. The external reproducibility of $^{234}\text{U}/^{238}\text{U}$ and $^{230}\text{Th}/^{238}\text{U}$ ratios of Schwartzwald Mine solutions measured during analytical sessions was better than 0.2% (2σ).

Activity ratios and ages were calculated using the half-lives of Jaffey et al. (1971) for ^{238}U , Holden (1989) for ^{232}Th , and Cheng et al. (2010) for ^{230}Th and ^{234}U . We note that slightly revised values for ^{230}Th and ^{234}U decay constants were presented by Cheng et al. (2013), however their effect on the ages reported here are minor and well within stated errors. Correction for U and Th from detritus was made assuming detritus with activity ratios of $(^{232}\text{Th}/^{238}\text{U}) = 1.2 \pm 0.5$, $(^{230}\text{Th}/^{238}\text{U}) = 1.0 \pm 0.1$, and $(^{234}\text{U}/^{238}\text{U}) = 1.0 \pm 0.1$ for McLean's Cave and $(^{232}\text{Th}/^{238}\text{U}) = 1.45 \pm 0.73$, $(^{230}\text{Th}/^{238}\text{U}) = 1.0 \pm 0.1$, and $(^{234}\text{U}/^{238}\text{U}) = 1.0 \pm 0.1$ for Moaning Cave. Ages and uncertainties were calculated with Isoplot 3.75 (Ludwig et al., 2010). Uncertainties of corrected ages are reported at the 2σ level and include measurement errors, procedural ^{230}Th blanks and uncertainties associated with detritus corrections but not decay constants. Data and ages are reported in (Table 1).

Age models were generated for both stalagmites using the StalAge algorithm of Scholz and Hoffmann (2011). Comparison of StalAge and other modeling approaches using both synthetic and natural data sets has demonstrated that it performs well in dealing with complexities such as outliers, age inversions, large and abrupt changes in growth rate, and hiatuses that are found in many speleothem time series constrained by $^{230}\text{Th}/\text{U}$ 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 Isotopes

Thick sections of the ML-1 stalagmite were microdrilled for stable isotope analysis using a Merchantek automated microdrilling system. Between 20-70 μg of powder was collected from systematically spaced locations along the growth axis on 8 of the thick sections. Samples were roasted at 375° C under vacuum for 30 minutes to remove organics and subsequently reacted in 105% phosphoric acid at 90° C in either a common acid bath on a GVI Optima Stable Isotope Ratio Mass Spectrometer (SIRMS) or a Gilson Multicarb Autosampler system (individual acid injection vials) interfaced with an Elementar Isoprime Mass Spectrometer housed in the UC Davis Stable Isotope Laboratory. CO₂ gas was analyzed in dual inlet mode and values were corrected using the Craig correction to account for the ¹⁷O contribution (Craig, 1957) and to an internal standard and reported relative to the Vienna Pee Dee Belemnite (VPDB). Both systems provide $\delta^{13}\text{C}$ precision of $\pm 0.04\text{‰}$ and $\delta^{18}\text{O}$ precision of $\pm 0.06\text{‰}$. Because the typical standard has significantly more positive carbon and oxygen isotope ratios than the ML1 samples, an additional internal standard, NBS 18, with more negative stable isotope ratios was analyzed as an unknown with each run to ensure accurate results. The average measured values for NBS 18 were -5.03‰ (SD 0.09‰) ($\delta^{13}\text{C}$) and -23.09‰ (SD 0.34‰) ($\delta^{18}\text{O}$) (n= 45) over the course of the measurement period. The accepted values of NBS 18 are -5.01‰ ($\delta^{13}\text{C}$) and -23.2‰ ($\delta^{18}\text{O}$).

Trace Elements

Trace element concentrations of Mg, Ba, and Sr were analyzed from eight thick sections using a Photon Machines 193 nm ArF UV excimer laser with a HelEx dual-volume laser ablation cell coupled to an Agilent 7700x quadrupole ICP-MS housed in the UC Davis Stable Isotope Laboratory. Continuous transects along each thick section were ablated adjacent to microdrill collection sites sampled for stable isotope analysis. Sample transects were preablated using a 20% laser power (4.45 J/cm^3), using a 10 by 175 μm rectangular slit and 20 $\mu\text{m/s}$ scan speed with a repetition rate of 10 Hz and sampled at 40% laser power using a 10 by 175 μm rectangular slit and 10 $\mu\text{m/s}$ scan speed with a repetition rate of 15 Hz and an energy of 4.45 J/cm^3 . Isotopes of ^{24}Mg , ^{25}Mg , ^{43}Ca , ^{44}Ca , ^{88}Sr , ^{137}Ba , and ^{138}Ba were measured. Multi-element synthetic glass standards NIST SRM 610 and 612 and MACS-3 (synthetic aragonite) were analyzed at the beginning and end of each transect. Only the NIST glass standards were used for elemental quantification despite matrix differences given that Jochum and others (2012) demonstrated that mass-load matrix effects are not significant for refractory lithophile elements, including Mg, Sr and Ba and that repeated analysis of a pressed synthetic calcium carbonate standard, MACS-3, yielded very noisy data. Following background subtraction, concentrations were derived using Iolite software (Hellstrom et al. 2008). The average uncertainty (2σ) of replicate ($n=14$) analysis of NIST 612 is 0.63% for ^{43}Ca , 0.95% for ^{24}Mg , 1.53 for ^{25}Mg , 0.75% for ^{88}Sr , and 0.65% for ^{138}Ba .

Data Treatment and Statistical Tests

Given the density and varying sampling resolutions of proxy data, it was necessary to smooth the proxy datasets (in particular Mg/Ca, and Sr/Ca) to assist with visual

comparison between proxies. As the trace element data were non-normally distributed, data were first log-transformed, and then Gaussian kernel density estimations were employed on the log-transformed data (Rehfeld et al., 2011). Kernel smoothing bandwidths were selected using the `dpill` function in the `Kernsmooth` package in R (Wand, 2014).

In order to compute Spearman's rank order correlation coefficients (r_s) and p values, data were detrended using the `RSEIS` package in R. For determining correlations between trace element and stable isotope time-series, all data were interpolated to a 10-year time-step (the median time-step in the lower resolution stable isotope times-series).

Results

U-series

Analytical results and calculated ages for ML-1 and new MC3 samples are presented in Table 1. All errors are 2σ . Measured U concentrations range from 27 to 70 ppb in ML-1 and 176 to 193 ppb for the new MC3 samples. Dates from both stalagmites preserve stratigraphic order within uncertainties. The new `StalAge` model for MC3 shows good agreement with the previous linearly interpolated age model (Oster et al., 2009) (Fig. 2). The ML-1 stalagmite precipitated from 19.4 to 11.9 ka, spanning the last deglaciation, including Heinrich Stadial 1 and largely overlapping with the MC3 record (16.7 to 8.8 ka).

Macroscopic and Microscopic Fabrics

Speleothem fabrics provide clues into drip water properties and degassing conditions of a cave that complement geochemical proxy records as well as inform the degree to which

the presence of impurities influenced stalagmite growth rates (Frisia and Borsata, 2010; Frisia 2015). Therefore, calcite fabrics of stalagmite ML-1 were evaluated and are reported here. Stalagmite ML-1 consists of low Mg calcite that is macroscopically banded in its lower half passing upward into a very poorly banded portion that contains mm- to cm-scale vuggy porosity (Fig. S1) and is overall made up of columnar calcite. The combined macroscopic features and microscopic calcite fabrics define three distinct segments (Fig. S1): (1) Segment 1, composed of the interval between 24.5 and 14.5 cm and corresponding to the period of 16.5 to 19.4 ka, (2) Segment 2, which spans the interval between 14.5 and 10.8 cm and corresponds to the period between 16.5 and 14.5 ka, and (3) Segment 3, defining the interval between 10.8 cm and the top of the stalagmite (0 cm) and corresponding to the period from 14.5 to the end of the record.

Segment 1 consists of macroscopic, alternating thin bands (0.5 mm to 0.7 cm) of dense translucent calcite and white inclusion-rich calcite with the former composing up to 70% of this interval (Fig. S1). Segment 2 is similarly composed of alternating white inclusion-rich and dense translucent bands, but individual bands are overall thicker (from 2 mm to 2 cm) and white inclusion-rich bands predominate. Microscopically for both segments, dense translucent calcite consists of inclusion-poor crystals, which are oriented perpendicular to the growth surface, exhibit straight compromise boundaries with no discernible intercrystalline porosity (Fig. 3a; white bands in 3c), and continue across different band types. This fabric is referred to as columnar compact (Frisia and Borsato, 2010). Intercalated white, inclusion-rich bands consist of acicular (elongated) crystals delineated by inclusions or pores along growth boundaries; inclusions are air- and/or

fluid-filled (Fig. 3b, darker bands in 3c). Crystals in some bands exhibit variable orientation relative to the growth surface creating a ‘feathery’ texture (Fig. 3e). This fabric is referred to as columnar open (Frisia and Borsato, 2010). Under epifluorescence, white inclusion-rich bands fluoresce green to bright yellow indicating the presence of variable amounts of colloidal organic matter (cf: McGarry and Baker, 2000; van Beynen et al., 2001; Smailer and White, 2013), whereas dense translucent bands are non-fluorescent or are mottled dully fluorescent (Fig. 3d).

Segment 3 is texturally distinct from the other two segments given the loss of well-developed banding, overall inclusion-rich calcite, and intervals with mm-to-cm-scale vuggy porosity (Figs. S1). Brightly fluorescent inclusion-rich bands, interpreted to record the presence of colloidal organic matter, are most prevalent in this interval. Calcite fabrics in this interval range from columnar open to elongated columnar with well-developed acute rhomboidal terminations. Inclusion-free and non- fluorescent and clear calcite forms overgrowths on some crystals, in particular those that partially occlude vuggy porosity (Fig. 3f). This calcite also overgrows crystal terminations at the top of the stalagmite.

Stable isotopes

The stable isotope time-series developed for the ML1 stalagmite offer sub- to multi-decadal resolution based on U-Th constrained calcite growth rates calculated using the StalAge age model. The $\delta^{18}\text{O}$ values fluctuate between -10.4‰ and -6.5 ‰, whereas $\delta^{13}\text{C}$ values vary systematically with clearly defined rises and falls between -11.29 ‰ and -

3.52 ‰ (Fig. 4). ML1 $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values exhibit strong positive down-axis correlation ($r_s = 0.63$, $p < 0.001$; Table 2), which could indicate kinetically driven non-equilibrium isotopic fractionation (Hendy, 1971). Nonequilibrium fractionation can occur during rapid CO_2 degassing and calcite precipitation driven by decreased drip rates, increased evaporation, and/or changing cave $p\text{CO}_2$, (Mickler et al., 2004; 2006; Dreybrodt and Scholz, 2011, Deininger et al., 2012). However, rapid precipitation of calcite can result in equilibrium $\delta^{18}\text{O}$ values (Feng et al., 2012), and covariation between $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ in stalagmites has been documented by some to record equilibrium isotopic fractionation (Romanov et al., 2008) and result from environmental processes that influence both isotopic compositions (Genty et al., 2003; Fleitmann et al., 2004; Frappier et al., 2007; Oster et al. 2009; 2014).

A “Hendy test” (1971) is typically used to rule out non-equilibrium fractionation during stalagmite precipitation by sampling along a single growth layer in both directions from the growth axis. The criteria for passing a Hendy test and thus confidently dismissing nonequilibrium fractionation includes relatively constant $\delta^{18}\text{O}$ values lacking progressive enrichment and a lack of correlation between $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ along a given growth layer. Samples were collected from ML1 along both a dense translucent band and a white inclusion-rich band. The $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values of the dense translucent band vary from the growth axis values by up to ± 0.55 ‰ and ± 0.42 ‰ respectively (Table 3) and do not exhibit progressive $\delta^{18}\text{O}$ enrichment or significant covariation between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values ($r_s = 0.43$, $p = 0.4$). The $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values of the white inclusion-rich band exhibit more variation from growth axis values of up to ± 2.18 ‰ and ± 1.25 ‰,

respectively (Table 4), but similarly do not exhibit $\delta^{18}\text{O}$ enrichment or significant covariation between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values ($r_s = -0.43$, $p = 0.3$). While stable isotopic values do vary significantly from on-axis values in the white inclusion-rich bands, the lack of along-band $\delta^{18}\text{O}$ enrichment, and lack of covariation between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values suggest minimal influence by non-equilibrium isotopic fractionation on measured isotopic compositions obtained along the growth axis. Furthermore, Romanov and others (2008) and Dorale and Liu (2009) suggested that the Hendy test is sometimes not a valid indicator of equilibrium isotopic fractionation as precipitation along the flanks of speleothems may be kinetically controlled despite equilibrium conditions during growth along the stalagmite axis. Kinetic effects on C and O isotopic fractionation, however, cannot be completely ruled out for the white inclusion-rich bands given the observed variation in $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values and their inclusion-rich, elongate columnar textures, which may record precipitation from rapid degassing of highly supersaturated drip waters.

Trace elements

Concentrations of Mg, Sr, and Ba are given as ratios with Ca. Sr/Ca and Ba/Ca display a strong positive correlation ($r_s = 0.75$, $p \ll 0.001$), while Mg/Ca displays a moderate positive correlation with both Sr/Ca, and Ba/Ca ($r_s = 0.38$, $p \ll 0.001$, and $r_s = 0.32$, $p \ll 0.001$, respectively). Mg/Ca and Ba/Ca show significant weak to moderate positive correlations with $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ (Mg/Ca: $r_s = 0.21$, $p \ll 0.001$; $r_s = 0.37$, $p \ll 0.001$, Ba/Ca: $r_s = 0.26$, $p \ll 0.001$, $r_s = 0.29$, $p \ll 0.001$, respectively) (Table 2). Sr/Ca shows significant positive correlation with $\delta^{18}\text{O}$ ($r_s = 0.26$, $p \ll 0.001$), and especially with $\delta^{13}\text{C}$ ($r_s = 0.48$, $p \ll 0.001$). Trace elements further show a similar long-term pattern to $\delta^{13}\text{C}$,

increasing from the beginning of the record until ~ 16.5 ka, then displaying an overall decreasing trend until ~ 14 ka (Fig. 4). Shorter, multi-centennial shifts are superimposed on these long-term trends.

Discussion

Interpretation of McLean's Cave proxy records

The ML-1 isotopic and trace element records display long-term trends and abrupt shifts across the end of the last glacial period and the deglaciation that show distinct similarities with regional and global climate changes at this time (Fig. 4). Both the ML-1 $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ records display rapid centennial-scale oscillations. For ML-1, long-term, millennial-scale changes in $\delta^{18}\text{O}$ of $\sim 1\text{-}2\text{‰}$, and larger $\delta^{13}\text{C}$ fluctuations are similar in timing to millennial-scale changes in the lower resolution Moaning Cave (MC3) record (Fig. 4), albeit with some offset in isotopic values discussed below.

Four years of observations of stable isotopes in rain and drip water from nearby Black Chasm Cavern reveal a significant positive correlation between rain $\delta^{18}\text{O}$ and surface air temperature, but show no relationship between rain $\delta^{18}\text{O}$ and rainfall amount (Oster et al., 2012). Additionally, moisture source and trajectory influence rain $\delta^{18}\text{O}$, with storms sourced in the North Pacific having significantly lower $\delta^{18}\text{O}$ values than storms that originate in the subtropical Pacific ($-15.5 \pm 0.4\text{‰}$ vs. $-6.4 \pm 2.5\text{‰}$) (Oster et al., 2012). Changes in rainwater $\delta^{18}\text{O}$ are transferred to cave drip water within ~ 3 weeks but with some mixing with stored water in the epikarst. Thus, while the isotopic signature of individual storms is muted in the cave drip water signature, changes occurring over

weeks to years should be transferred to drip waters and recorded in speleothem carbonate (Oster et al., 2012).

As McLean's Cave is located in the same climate regime and receives the same storm systems as Black Chasm, and is also a vertically oriented cave developed along fractures in a small marble unit, we anticipate that the behavior of drip-water isotopic signatures should be similar between the two systems on the timescales that are relevant for the ML-1 record. Thus, we interpret changes in ML-1 $\delta^{18}\text{O}$ to reflect changes in temperature and potentially also sustained changes in moisture source across the last deglaciation. The overall long-term patterns of change in $\delta^{18}\text{O}$ suggest a temperature minimum centered around 18.5 ka (Fig 4 – all values corrected for changes in ice-volume) that increases until ~16.5 ka. The ~2‰ increase observed over this interval suggests a maximum temperature increase of 3 to 4°C using the mid-latitude relationship of 0.5-0.7‰ per °C (Rozanski et al., 1993), though it is possible that some of the speleothem $\delta^{18}\text{O}$ increase resulted from a larger proportion of subtropical moisture. Following 16.5 ka, $\delta^{18}\text{O}$ values decrease, reaching ~1‰ lower values by 16 ka, and subsequently recover by 15ka. $\delta^{18}\text{O}$ values once again decrease at ~14.5 ka followed by a moderate rise between 13.5 and 12.3 ka. Similar trends are evident in the MC3 $\delta^{18}\text{O}$ record, but a more pronounced shift to higher $\delta^{18}\text{O}$ values occurs at 13.5 ka

There is no evidence for a significant proportion of C_4 flora in this region during the last glacial and deglaciation (Cole, 1983; Thompson et al., 1993, Davis, 1999). Carbon isotope values in ML-1 should therefore reflect a combination of changes in soil

respiration and prior calcite precipitation (PCP). Both of these processes shift speleothem $\delta^{13}\text{C}$ in the same direction in response to changes in water availability above the cave. Increased rainfall should lead to increased soil respiration and decreased PCP, causing a negative shift in the speleothem $\delta^{13}\text{C}$ record, and vice versa, as has been noted in drip waters and speleothem records from the central Sierra Nevada (Oster et al., 2009; 2012; 2014). Significant positive correlations between ML-1 $\delta^{13}\text{C}$ and Sr/Ca and Mg/Ca support the interpretation that $\delta^{13}\text{C}$ at least in part reflects changes in PCP (Fairchild et al., 2000).

Overall, there is a negative trend throughout the ML-1 $\delta^{13}\text{C}$ record, upon which are superimposed three shorter-term asymmetric cycles of gradually rising $\delta^{13}\text{C}$ values followed by more rapid declines. In general, these shorter-term trends coincide with stadials and interstadials with values rising from minimum values in a given stadial through the remainder of the stadial to maximum values in the subsequent interstadial. $\delta^{13}\text{C}$ values decrease rapidly to new minimum values in the early portions of the following stadial interval (Fig. 4). Highest $\delta^{13}\text{C}$ values, suggesting the overall driest conditions and/or potentially the least soil respiration, occur during the late glacial, with an upward trend from ~ 18.5 to 16 ka. The peak in $\delta^{13}\text{C}$, and potentially the driest interval in the ML-1 record, occurs between ~ 17.5 and 16.1 ka during the first part of Heinrich Stadial 1 (HS1). A prominent shift toward substantially more negative values occurs during the later half of HS1 beginning around 16.1 ka. Based on pluvial lake records from New Mexico, Broecker et al., (2009) divided HS1 into an earlier dry phase (the Big Dry), and a later wet phase (the Big Wet), and we see some evidence of these changes in the ML-1 record as well. Increases in both ML-1 and MC3 $\delta^{13}\text{C}$ occur beginning ~ 15.5

ka and 13.5 consistent with drier conditions and/or less soil respiration during warm periods (Bölling, Alleröd). Decreases in $\delta^{13}\text{C}$ occur at 14.5 ka and 12.6 ka, consistent with wetter conditions and/or increased soil respiration during cooler periods (Older and Younger Dryas).

The records of Mg/Ca and Sr/Ca for the ML-1 stalagmite display an overall similar long-term trend to the $\delta^{13}\text{C}$ record (Fig. 4), although differences in magnitudes of increases and decreases and timing of changes exist. These ratios in stalagmites can vary due to changes in PCP with increased Sr/Ca and Mg/Ca indicating enhanced PCP due preferential removal of Ca relative to Sr and Mg upstream of the stalagmite growth site (Fairchild et al., 2000). Variations in PCP can arise due to changes in seepage water residence time in the epikarst, with longer residence times promoting enhanced CO_2 degassing and PCP, and shorter residence times promoting reduced degassing and PCP (Fairchild et al., 2000; Johnson et al., 2006; McDonald et al., 2007). In this sense, inferred changes in PCP can be related to changes in water supply to a cave. However, changes in cave ventilation can also influence the degree of CO_2 degassing and calcite precipitation along the seepage water flow path that can obscure the interpretation of PCP proxies as reflecting paleo-aridity (Banner et al., 2007; Matthey et al., 2010; Kowalczyk and Froelich, 2010; Wong et al., 2011). Additionally, in dolomite-rich host-rocks, changes in the degree of dolomite dissolution can influence Mg/Ca and Sr/Ca values of drip waters and speleothems (Fairchild et al., 2000; Tremaine and Froelich, 2013). Variations in the amount of dolomite dissolution have been variably interpreted to reflect changes in climate. Incongruent dissolution of dolomite and calcite given solubility differences leads

to seepage waters first becoming supersaturated with calcite before they reach dolomite supersaturation (Roberts et al., 1998). Thus, slower seepage water flow and longer residence times due to drier conditions could lead to increased dolomite dissolution, raising the seepage water Mg/Ca ratio and possibly decreasing Sr/Ca as dolomite typically contains less Sr than calcite (Fairchild et al., 2000; Hellstrom and McCulloch, 2000).

Drip water Mg/Ca and Sr/Ca at Black Chasm Cavern respond to variations in PCP that are controlled both by changes in cave air ventilation and water supply (Oster et al., 2012). However, the dominant control on PCP during the seasons of speleothem growth (autumn-winter-spring) at Black Chasm is water supply, and Mg/Ca and Sr/Ca ratios of drip water have been shown to record changes in water supply to the cave (Oster et al., 2012). Similar controls should apply to McLean's Cave. However, the marble unit that contains McLean's Cave has a higher dolomite component than the Black Chasm host rock (Bowen, 1973), which may also influence drip water and speleothem Mg/Ca and Sr/Ca ratios. Variable dissolution of dolomite in the host rock appears to have influenced the Mg/Ca and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in a stalagmite from McLean's Cave that grew during the last glacial period (55 to 67 ka), with increased dolomite dissolution occurring during drier interstadials (Oster et al., 2014). Importantly, host rock calcite at McLean's Cave has a slightly higher $\delta^{13}\text{C}$ value than host rock dolomite (2.22‰ vs. $1.63\text{‰} \pm 0.03$) (Oster et al., 2014), suggesting that correspondence between stalagmite $\delta^{13}\text{C}$ and Mg/Ca in the ML-1 deglacial record (Fig. 4, Table 2) does not primarily reflect the influence of

dolomite dissolution on $\delta^{13}\text{C}$. Rather it suggests these two parameters independently respond, at least in part, to PCP, which would lead to a positive co-variation.

In order to further investigate and disentangle the relative influences of dolomite dissolution vs. PCP on the ML-1 trace element records through time, we divided the record into seven temporal intervals: the late glacial (LG, older than 17.5 ka), Big Dry (BD, 16.1-17.5 ka), the Big Wet (BW, 15-16.3 ka) after Broecker and Putnam (2012), the Bölling (BO, 14.5 – 15 ka), the Older Dryas (OD, 13.5-14.5 ka), the Alleröd (AL, 12.7-13.5 ka), and the Younger Dryas (YD, <12.7 ka). In dolomite-rich host rocks such as the McLean's Cave host-rock, speleothem Mg/Ca vs. Sr/Ca should follow linear relationships if PCP is the dominant factor, where the degree of dolomite dissolution sets the slope of the line, with steeper slopes arising from smaller contributions of dolomite to the initial seepage water (Tremaine and Froelich, 2013). A highly significant linear correlation ($p < 0.001$) between Mg/Ca and Sr/Ca across the entire record suggests that differential PCP plays a large role in controlling trace element concentration in ML-1. The relatively low observed coefficient of determination ($R^2 = 0.17$) for the ML-1 data reflects varying dolomite contribution across the record that have a greater influence on Mg/Ca over Sr/Ca.

Variation in the density of points along a common line of best fit during each time interval (Fig. 5) indicates that PCP was more prevalent during some time periods than others. For example, points are spread widely along the line of best fit during the LG and BW, suggesting variable amounts of PCP during those intervals. By comparison, during

the BD, the locus of points is shifted positively along the line of best fit, suggesting that PCP was at a maximum during this interval. Distribution of the trace element points is similar between the AL and BO and suggests variable degrees of PCP. However, AL trace elements appear to fall into two distinct clumps possibly reflecting distinct climate (wetter vs. drier) or hydrologic regimes, with wetter conditions (less PCP) during the early AL. The apparent negative slopes during the OD and YD suggest that trace element variations in these intervals are more directly controlled by changes in dolomite contribution with little modification by PCP, especially during the OD. A shift to less dolomite contribution, under shorter flow residence times, would lead to lower Mg/Ca, but potentially higher Sr/Ca (Fairchild et al., 2000, Hellstrom and McCulloch, 2000; Tremaine and Froelich, 2013). PCP would also be limited under shorter flow residence times, limiting the enhancement of both Mg/Ca and Sr/Ca, consistent with clustering toward lowest values during the OD. Like the AL, OD trace element values appear to be partitioned into two distinct clusters, also potentially indicating a shift in climate or hydrologic regime within this interval. As a whole, the Mg/Ca and Sr/Ca time-series (Fig. 4) and cross-plots (Fig. 5) support observations from the $\delta^{13}\text{C}$ record. The highest degree of PCP, and thus potentially the driest conditions above McLean's Cave, occurred during the Big Dry at the beginning of HS1 (BD in Figs. 4 and 5). This interval was followed by an overall trend to wetter conditions reflected by decreasing trace element ratios from 16.3 to 13.5 ka, punctuated by drying during the AL. With the exception of the very beginning of the BO, Mg/Ca and Sr/Ca show a decreasing trend during the BO while $\delta^{13}\text{C}$ is increasing. This could reflect a stronger influence of decreased soil respiration on

the $\delta^{13}\text{C}$ signal due to warmer temperatures inferred from stalagmite $\delta^{18}\text{O}$ during this interval (Fig. 4).

Interpretations of the ML-1 stable isotope and trace element records are supported by variations in stalagmite crystal fabric. The predominance of inclusion-poor, columnar compact calcite fabric in the interval of the stalagmite, which we describe as Segment 1 (Figs. 3 and 4), suggests precipitation during the LG and BD under relatively constant and low drip rates and enhanced degassing likely due to well-ventilated cave conditions (Frisia et al., 2000; Frisia and Borsato, 2010; Boch et al., 2011; Frisia, 2015). The alternation of bands of this fabric with bands of columnar open calcite in Segment 1 are consistent with the stable isotope and trace element data that suggest variable, but overall enhanced PCP, especially during the BD (Fig. 3 and 4). Notably, the slowest speleothem growth rates are not observed during the BD when proxies indicate potentially the driest conditions above the cave, indicating that growth rate alone cannot be used as a proxy for water supply, and underscoring the importance of using a multi-proxy approach that includes crystal fabric interpretations to deduce past climate conditions.

Segment 2 was deposited during the BW and BO (16.3 to 14.5 ka) and is characterized by a predominance of white bands of inclusion-rich calcite that displays a columnar open fabric. Such fabrics form beneath thicker fluid films that lead to generally less efficient degassing (Frisia et al., 2000; Frisia and Borsato, 2010; Belli et al., 2013; Frisia, 2015), which corresponds well with the overall slower growth rate of Segment 2 (Fig. 4). Columnar open fabrics have been associated with strong seasonal contrasts in drip rate

(Boch et al., 2011; Frisia, 2015), which possibly became more pronounced in McLean's Cave beginning with the BO as climate became seasonally warmer (higher $\delta^{18}\text{O}$) and potentially drier (increasing $\delta^{13}\text{C}$) (Figs. 3-5). This hypothesis is further supported by an increase in the thickness of white inclusion-rich bands upward in this interval as well as the development of up to mm-scale vugs, aligned with banding, in the upper half of Segment 2.

Segment 3, which was deposited during the OD, AL, and early YD (14.4 to 12 ka), is characterized by less distinctly banded columnar open and elongated columnar crystal fabrics. These fabrics, which are indicative of precipitation under relatively fast drip rates and low calcite supersaturation states, are consistent with the intervals of higher growth rates in this segment (OD in Fig. 4). Intervals within Segment 3 that are dominated by elongated columnar fabric may record increased drip-water Mg/Ca and/or an increase in the influx of organic colloidal compounds (Frisia, 2015). The loss of banding and increased vugginess, inferred to record rapid degassing, in the AL portion of Segment 3 further supports the drier climate independently suggested by the rising Mg/Ca ratios and $\delta^{13}\text{C}$ values.

Taken together the ML-1 proxy records and stalagmite fabrics point to slow but steady stalagmite growth during the LG and BD (19-16.3 ka), with increased PCP during the BD (16.3-17.5ka), likely due to drier conditions. The ML-1 $\delta^{18}\text{O}$ record also indicates a warming trend with a possible shift to more subtropical moisture throughout this interval. The onset of the BW (16.3-14.5ka) is characterized by a shift to colder conditions and less PCP and potentially also increased soil respiration, suggestive of a wetter

environment. Warming returned leading into the BO (14.5 ka), with calcite fabrics suggesting seasonally variable flow and variable amounts of PCP. A shift to cooler, wetter climate above McLean's Cave marks the onset of the OD (13.5-14.5 ka), with overall less PCP, faster speleothem growth rates and increased influx of organic colloids. PCP increases again modestly during the AL (12.7-13.5 ka), and decreases into the beginning of the YD (12.7 ka). On the whole, the records suggest a progressively wetter climate from the BD onward, with episodic, millennial-scale dry intervals largely coincident with interstadial events. Soil respiration likely also increased in the later part of the record (OD, AL, YD), as suggested by the more negative $\delta^{13}\text{C}$ and the increased inclusion of organic colloidal compounds in the stalagmite. Only the very beginning of the YD is captured by the ML-1 stalagmite, but the MC3 record suggests overall wetter conditions persisted in the region until ~10 ka (Oster et al., 2009).

Comparison of the McLean's and Moaning Cave Records

Long-term, millennial-scale changes in ML-1 $\delta^{18}\text{O}$ of ~ 1-2 ‰ are similar in timing to millennial-scale changes in the lower resolution Moaning Cave (MC3) record (Fig. 4). Overall ML-1 $\delta^{18}\text{O}$ values are about 1‰ lower than coincident MC3 $\delta^{18}\text{O}$ values. Likewise, millennial-scale changes in the ML-1 $\delta^{13}\text{C}$ record are similar to those in MC3. With the exception of the interval from ~14.5 to 13 ka, $\delta^{13}\text{C}$ values overlap between the two records. It is possible that there is a small temporal offset between these two records, as the MC3 record lacks U/Th ages in that specific interval. Additionally, while both cave records preserve a rapid decrease in $\delta^{13}\text{C}$ at 14.5 ka, the magnitude of the decrease is ~ 2‰ larger in the ML-1 record. These differences in the isotopic ratios (for $\delta^{18}\text{O}$) and

magnitude of some shifts (for $\delta^{13}\text{C}$) might occur due to differences in the time-scales of water flow into these two caves. Although the area above McLean's Cave is presently quarried, and we cannot assess the overlying soil, there is evidence from $^{87}\text{Sr}/^{86}\text{Sr}$ measurements of an older McLean's Cave stalagmite that the volcanic soil that is developed above Moaning Cave (Oster et al., 2009), was not present above McLean's Cave during the last glacial period (Oster et al., 2014). Thus, these differences in the stable isotope records from the two caves might result from thinner soils above McLean's Cave that allowed more rapid infiltration into the host rock, with the potential for less modification of seepage water $\delta^{18}\text{O}$ by evaporation. Likewise, more rapid infiltration through the epikarst at McLean's Cave would lead to less modification of $\delta^{13}\text{C}$ by PCP during relatively wet intervals possibly accounting for the $\sim 2\text{‰}$ offset between MC3 and ML-1 $\delta^{13}\text{C}$ between 14.5 and 13 ka. Regardless of the differences in water transmission from surface to cave in these two locations, the replication of isotopic variations between these stalagmites indicates that what is being recorded are changes in climate rather than local variations in the karst systems.

Regional climate change during the last deglaciation

The transition out of the last glacial period coincided with significant hydroclimatic changes in the western United States (US) as indicated by proxy records from lakes (e.g. McGee et al., 2012; Broecker and Putnam, 2012; Munroe and Laabs, 2013; Benson et al., 2013; Ibarra et al., 2014; Reheis, 2014) and speleothems (Fig. 6) (Polyak et al., 2004; 2012; Oster et al., 2009; Wagner et al., 2010; Asmerom et al., 2010; Feng et al., 2014). The most distinct feature of Great Basin lake records is widespread lake expansion

following the discharge of icebergs in the North Atlantic during HS1 (Fig. 6; Munroe and Laabs, 2013; Benson et al., 2013; Ibarra et al., 2014). There is some evidence that the climate of HS1 in western North America can be divided into two distinct phases. Desiccation at Lake Estancia in New Mexico during the first part of HS1 lead Broecker et al., (2009) to call this interval the “Big Dry”. However, evidence for drying at this time is not consistent across the region (Fig. 6). Dry conditions are suggested by a paired speleothem $\delta^{13}\text{C}$ and $\delta^{234}\text{U}$ record from Fort Stanton Cave in New Mexico (Polyak et al., 2012), and proxy records from Lake Elsinore and Swamp Lake in California (Kirby et al., 2013; Street et al., 2012), but are not apparent from Great Basin Lake levels (Bacon et al., 2006; Broecker and Putnam, 2012; McGee et al., 2012; Munroe and Laabs, 2013).

The later part of HS1 when many lakes, including Lake Franklin, Lake Bonneville, and Lake Surprise reached highstands, has been described as the wettest period in the Great Basin in the past 30,000 years, and termed the “Big Wet” interval (BW) (Broecker and Putnam, 2012; Munroe and Laabs, 2013; Ibarra et al., 2014). Some proxy records indicate a transition to wetter conditions in the middle of HS1 (Fig. 6). The Fort Stanton speleothem $\delta^{234}\text{U}$ record shows such a transition around 16.1 ka (Polyak et al., 2012). The Lake Elsinore leafwax $\delta^2\text{H}$ record suggests a brief shift to colder/wetter conditions (Kirby et al., 2013), and cold and wet conditions are also noted at Swamp Lake on the west side of the Sierra Nevada during the later part of HS1 (Street et al., 2012). Consistent with these records, our ML-1 record shows evidence of a two-phased HS1 with drier conditions above McLean’s Cave between 16.1 and 17.5 ka – during the Big Dry – and a shift toward wetter conditions at 16.1 ka (Figs. 4-6).

633

634 Oxygen isotope records from Cave of the Bells and Fort Stanton Caves in Arizona and
635 New Mexico, respectively, in the southwestern US display large and abrupt increases at
636 the onset of the Bölling interstadial (BO), which are interpreted as decreases in the
637 proportion of winter, relative to summer, moisture to that region (Fig. 6) (Asmerom et al.,
638 2010; Wagner et al., 2010). However, the Fort Stanton $\delta^{234}\text{U}$ record indicates wet
639 conditions persisted through this interval in New Mexico (Polyak et al., 2012). Together,
640 these records suggest enhanced summer moisture to the southwest, which is consistent
641 with a Texas speleothem record suggesting greater moisture transport from the Gulf of
642 Mexico during this time (Feng et al., 2014). The Lake Elsinore leafwax $\delta^2\text{H}$ record points
643 to warmer or drier conditions, or an increase in subtropical moisture across this interval
644 (Kirby et al., 2013). Great Basin lakes show evidence of desiccation at this time (Fig. 6),
645 however lake levels are also influenced by evaporation, which likely increased with
646 rising summer insolation at this time (McGee et al., 2012; Munroe and Laabs, 2013). The
647 transition out of HS1 into the (BO) in the ML-1 record is marked by increased $\delta^{18}\text{O}$,
648 which is consistent with warming and/or more subtropical moisture, and a substantial
649 increase in $\delta^{13}\text{C}$ suggesting increased PCP and/or reduced soil respiration driven by
650 decreased effective moisture (Figs. 4, 6). Calcite crystal fabrics are indicative of highly
651 variable drip rates, and trace elements indicate variable amounts of PCP. Together, the
652 ML-1 proxies indicate a more seasonally variable, overall drier climate during the BO.

653

654 Following the Bölling interstadial, the ML-1 and MC3 records document millennial-scale
655 shifts in temperature and/or moisture source, and precipitation amount that are noted to

656 varying degrees in other regional records (Fig. 6). The Cave of the Bells record
657 documents a relatively small increase in winter precipitation at 14 ka (Wagner et al.,
658 2010), coincident with the increase in effective moisture during the OD inferred from the
659 ML-1 record. Two shifts to lower $\delta^{18}\text{O}$ values are also observed during this interval in the
660 Fort Stanton, NM record, but they are small and are not sustained (Asmerom et al., 2010).
661 Both Cave of the Bells and ML-1 show shifts to drier conditions and less winter
662 precipitation coincident with the Alleröd interstadial (Fig. 6). The Fort Stanton $\delta^{234}\text{U}$
663 record suggests progressive drying from 14 ka until 12 ka (Polyak et al., 2012). Although
664 the Lake Elsinore leafwax $\delta^2\text{H}$ record does not suggest millennial-scale variability during
665 this interval (Fig. 6), the higher resolution sediment records from Lake Elsinore do
666 indicate shifts to wetter conditions and a deeper lake around 14 ka (Kirby et al., 2013),
667 coincident with the inferred increase in effective moisture and winter precipitation noted
668 in the ML-1, MC3, and Cave of the Bells proxy records.

669
670 At the onset of the Younger Dryas, the Cave of the Bells and Fort Stanton records
671 indicate an increase in the proportion of winter precipitation (Wagner et al., 2010,
672 Asmerom et al., 2010), while the Leviathan and Oregon Caves records suggest a shift to
673 colder temperatures (Fig. 6) (Vacco et al., 2005; Lachniet et al., 2014). The Fort Stanton
674 U isotope record, however, suggests continued drying throughout the YD interval
675 (Polyak et al., 2012), while a study of speleothem growth rates in three other New
676 Mexico caves has previously suggested wet conditions from 12.5 to ~10.5 ka (Polyak et
677 al., 2004). Lake records from the Sierra Nevada suggest cold and wet conditions at the
678 height of the YD (Mensing et al., 2001; Porinchu et al., 2003; MacDonald et al., 2008;

Street et al., 2012), though there is variable evidence for warming and drying later in the YD (MacDonald et al., 2008; Street et al., 2012). The ML-1, and in particular the MC3 records which extend through the YD interval also suggest wetter conditions in the central Sierra Nevada. However, there is no evidence of a wet YD or an increase in winter precipitation at Lake Elsinore (Kirby et al., 2013), or at Owens Lake in the eastern Sierra Nevada (Bacon et al., 2006).

Mechanisms for precipitation change in western North America

Collectively, paleoclimate proxy records from the western United States demonstrate significant and abrupt hydroclimatic changes during the last deglaciation that were sometimes, but not always, regionally consistent. The hydroclimatic changes observed in proxy records from the western US likely reflect changes in the winter westerly storm track across the last deglaciation. This interval represents one of significant changes in boundary conditions, as summer insolation increased to its early Holocene maximum, and the Laurentide ice sheet waned. As such, the most important controls on the strength and position of the westerly storm track, which delivers the majority of winter precipitation to this region, likely also varied across this interval. At the Last Glacial Maximum, climate models simulate a splitting of the jet stream and a southward displacement of the westerly storm track due to the presence of the ice sheet (COHMAP Members, 1988; Bromwich et al., 2004; Kim et al., 2008). Recent proxy comparisons with the Paleoclimate Model Intercomparison Project ensembles suggests that the high pressure that develops over the ice sheet steers the storm track in a NW-SE trend across the western US, producing a pattern of increased precipitation in the southwest, and

decreased precipitation in the north (Oster et al., 2015). It has been proposed that the storm track stepped northward with the shrinking ice sheet, filling pluvial lakes to their maximum on its northward march (Hostetler and Benson, 1990; Negrini et al., 2002). However, this hypothesis has been questioned, as pluvial lake highstands are not zonally synchronous during the deglaciation (Lyle et al., 2012). Furthermore, evidence for millennial-scale wet and dry periods in speleothem records call for a different mechanism to alter the storm track, as changes in the ice sheet height cannot account for all abrupt changes in rainfall inferred from the proxy records (Oster et al., 2009; Wagner et al., 2010; Asmerom et al., 2010).

In particular, the rise of Great Basin lakes to deglacial high-stands during HS1 (Munroe and Laabs, 2013; Putnam, 2015) is suggestive of significant and rapid teleconnections between winter rainfall in the western US and conditions in the North Atlantic. Proxy records from the North Atlantic region indicate that the influx of icebergs and melt water associated with Heinrich Event 1 lead to weakened overturning circulation and an expansion of winter sea ice (Denton et al., 2005; 2010). Climate models connect this winter sea ice build up with a southward shift of the ITCZ (Chiang and Bitz, 2005); a strengthening of the northern Hemisphere Hadley cell and boreal subtropical jet that lead to enhanced rainfall in western North America (Chiang et al., 2014). Collectively, the McLean's Cave proxy records demonstrate an abrupt shift to wetter conditions during the BW/HS1, consistent with this scenario.

Overturning circulation in the North Atlantic recovered during the BO, and the ITCZ stepped northward (Denton et al., 2010). However, an influx of melt water during the YD again weakened overturning circulation in the North Atlantic and has been associated with a southward shift of the ITCZ (Peterson et al., 2000; Chiang and Bitz, 2005; Denton et al., 2010). A dynamical comparison of simulated BO and YD climates over the Pacific (Wong et al., in revision) finds that perturbation of the atmospheric state during the BO relative to the YD results in an enhanced Aleutian Low and stronger mid-level atmospheric high pressure ridge off the North American coast (Fig. 7a). This atmospheric configuration is conducive to greater poleward meridional transport, and results in the northward diversion of the Pacific winter storm track (Sewall and Sloan, 2004; Otto-Bliesner et al., 2006; Singarayer et al., 2006; Wang and Schubert, 2014; Liu et al., 2007). Model results further suggest that the Pacific winter storm track is less intense during the BO relative to YD, as opposed to being solely offset or shifted (Fig. 7b). Differences in the atmospheric state between the BO and YD may reflect a combination of higher ice sheet elevations (Otto-Bliesner et al., 2006) and/or diminished sea ice cover (Sewall and Sloan, 2004; Singarayer et al., 2006; Budikova, 2009).

A stronger winter westerly storm track during the YD relative to the BO would explain many of the observations noted in western United States paleoclimate records. The decrease in speleothem $\delta^{18}\text{O}$ at Cave of the Bells and Fort Stanton are consistent with this scenario as is the shift to wetter conditions above Moaning and McLean's Cave. The Fort Stanton $\delta^{234}\text{U}$ record, however, suggests that this increase in winter storm proportion did not lead to significantly wetter conditions above the cave (Polyak et al., 2012), and proxy

records from Lake Elsinore further suggest that no increase in northerly sourced precipitation occurred during the YD at that site (Kirby et al., 2013). This indicates that increased winter storm activity associated with the YD might not have extended far enough to the south to influence Lake Elsinore. This inference is supported by the modeled increase in YD over BO storminess over the Pacific, which is larger off of the northern California coast and the Pacific Northwest than off the southern California coast (Fig. 7b). That the increase in winter storms is noted in the Cave of the Bells and Fort Stanton records at similar latitude to Lake Elsinore, though perhaps was not strong enough to elicit substantial changes infiltration amount, might have arisen due to a NW-SE trending storm track, similar to what is hypothesized for the LGM (Oster et al., 2015). This spatial trend in the LGM storm track arises due to steering by high-pressure cells over the Laurentide Ice Sheet and the Pacific Ocean. Though the Laurentide Ice Sheet was comparably smaller at the YD (Tarasov et al. 2012), it is possible that weaker anticyclonic tendencies over the ice sheet (COHMAP, 1988) continued to contribute to NW-SE steering of the storm track that brought its influence to sites in the interior southwest of North America.

Conclusion

The McLean's and Moaning Cave speleothem proxy records indicate substantial changes in the amount, and perhaps the source, of precipitation during the last deglaciation. The Sierra Nevada speleothem records are consistent with other paleoclimate records from the western US that document changes in regional hydroclimate on millennial timescales during the last deglaciation (Asmerom et al., 2010; Wagner et al., 2010; Polyak et al., 2012; Kirby et al., 2013). Collectively, these records point to increased effective moisture

due to enhanced winter precipitation during BW/HS1 and the OD and drier conditions during the AL and BO. The Sierra Nevada speleothem records also show evidence of late glacial drying coincident with the Big Dry event noted in some Great Basin lake records (Broecker and Putnam, 2012).

Regional differences exist in the hydroclimatic response to the YD, with central Californian and interior southwestern sites indicating an increase in northern-sourced storms that is not apparent in southern California. Comparison of modeled warm/interstadial (BO) and cold/stadial (YD) climates in this region indicate a strengthened YD storm track off the northern California coast that would have contributed to wetter conditions at the YD (Wong et al., in revision). Steering of this storm track by high pressure over the Laurentide Ice sheet might have directed some of these storms into the interior southwest.

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Figure Captions

Fig. 1: A) Map of paleoclimate records discussed in the text: OC= Oregon Caves (Vacco et al., 2005; LS= Lake Surprise (Ibarra et al., 2014); LL: Lake Lahontan (); LF= Lake Franklin (Munroe and Laabs, 2013); LB= Lake Bonneville (McGee et al., 2012); MC=Moaning Cave; ML= McLean's Cave; SL= Swamp Lake (Street et al., 2012); LV= Leviathan (Lachniet et al., 2014); LEL= Lake Elsinore (Kirby et al., 2013); COB = Cave of the Bells (Wagner et al., 2010); LE= Lake Estancia (Broecker and Putnam, 2012); FS= Fort Stanton Cave (Asmerom et al., 2010). B) Location of McLean's Cave (ML; 38°4.20'N, 120°25.20W; elevation of 300 m above sea level) in the western foothills of the Sierra Nevada 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.

Fig. 2: U/Th ages vs. depth and age models for the Moaning Cave stalagmite MC3 (top) and the McLean's Cave stalagmite ML1 (bottom). StalAge age models shown in black with error envelopes in gray. The original linearly interpolated MC3 age model (Oster et al., 2009) is shown as a dashed line in the upper panel.

Fig. 3: Photomicrographs of calcite fabrics in ML1: a) Large columnar crystals (crossed-polarized light) in a white inclusion rich band from Segment 1 (thick section 6). b) Variably oriented columnar crystals of Segment 2; compromise boundaries are irregular and sub-crystals exhibit scalenohedron terminations (arrow) (thick section 3). c)

Stained inclusion-rich couplet from Segment 2 (plane-polarized light) (thick section 7). d) Stained inclusion-rich couplet (Segment 2) under epi!uorescence; note brighter color of stained bands and scalloped termination of moderately bright, inclusion-rich band (arrow) (thick section 7). e) Columnar calcite (plane-polarized) in Segment 2 composed of variably oriented subcrystals creating a 'feathery' texture within individual inclusion-rich band (thick section 3). f) Lack of banding in Segment 3 due to variable orientation of irregular crystals; note stepped scalenohedral terminations of sub-crystals and inclusion-free rhombohedral overgrowths (arrow) on sub-crystals growing into vugs (thick section 11). For references to thick section numbers, please see Figure S1 in the supporting online material.

Fig. 4: ML and MC proxy records. From top to bottom: MC (purple) and ML (red) $\delta^{18}\text{O}$ records and U-Th ages. Black line is kernel smoother through ML $\delta^{18}\text{O}$. MC and ML $\delta^{18}\text{O}$ values have been corrected for changes in sea level. MC (green) and ML (blue) $\delta^{13}\text{C}$ records. ML-1 Mg/Ca (orange) and Sr/Ca (green) with associated kernel smoothers (black lines). Data have been log-transformed. ML-1 $\delta^{13}\text{C}$ (blue) and growth rate (black) with associated 2σ errors. Dashed vertical lines designate segments defined by different calcite fabrics (see text). Gray and white bands highlight cold/wet and warm/dry intervals, respectively, as described in the text. More PCP, less soil respiration, and more dolomite dissolution can each indicate a drier climate, while less PCP, more soil respiration, and less dolomite dissolution can each indicate a wetter climate.

Fig. 5: Plots of Mg/Ca vs. Sr/Ca in the ML-1 stalagmite for different intervals (see text).

Shading shows number of points in each area of the plots, with blue shades indicated increased numbers. Black lines represent the line of best fit for the entire data set for reference. Lower right panel shows the entire dataset. Solid red arrow shows vector of increasing PCP. Dashed lines show same PCP vector but for different amounts of dolomite dissolution with steeper slopes indicating less input from dolomite (after Tremaine and Froelich, 2013).

Fig. 6: Regional paleoclimate record comparison. From top to bottom: Oregon Caves speleothem $\delta^{18}\text{O}$ (Vacco et al., 2005); MC (purple) and ML (red) $\delta^{18}\text{O}$; MC (green) and ML (blue) $\delta^{13}\text{C}$; Lake level as % maximum for Lake Franklin (green line; Munroe and Laabs (2013), Lake Lahontan (black line; Benson et al., 1995; Adams et al., 2008), and Lake Bonneville (gray line; Oviatt et al., 1992; McGee et al., 2012); Lake Elsinore leafwax $\delta^2\text{H}$ (Kirby et al., 2013); Leviathan speleothem $\delta^{18}\text{O}$ (Lachniet et al., 2014); Cave of the Bells speleothem $\delta^{18}\text{O}$ (Wagner et al., 2010); Fort Stanton speleothem $\delta^{18}\text{O}$ (Asmerom et al., 2010); and U isotope ratios (Polyak et al., 2012). Dashed lines separate time intervals described in the text with gray shading designating the cold/wet intervals noted in the ML-1 record.

Fig. 7: A. Differences in fraction of winter sea ice extent (shaded) and winter sea level pressure (hPa; contours) simulated for the Younger Dryas (YD) cool event and Bölling (BO) warm event. B. Differences in the Pacific storm track simulated during the two intervals shown as percent difference in modeled winter low-level (850 hPa) eddy kinetic

- 1149 energy (a measure of storminess; shading) and winter precipitation (mm/d; contours).
- 1150 Differences are calculated as YD minus BO.

Table's

Table 1

Sample Name	Sample Wt. (mg)	U (ppb)	^{232}Th (ppb)	$(^{230}\text{Th}/^{232}\text{Th})$	$(^{232}\text{Th}/^{238}\text{U})$	$\pm$ (%)	$(^{230}\text{Th}/^{238}\text{U})$	$\pm$ (%)	$(^{234}\text{U}/^{238}\text{U})$	$\pm$ (%)	Uncorrected Age, Error (ka)	Corrected Age, Error (ka)	Initial $(^{234}\text{U}/^{238}\text{U})$
McLean's Cave													
ML1-5	199.7	43.80	1.643	10.8	0.01206	0.29	0.12999	0.74	1.100	0.46	13.82 $\pm$ 0.13	12.70 $\pm$ 0.52	1.104 $\pm$ 0.0054
ML1-11	246.3	35.40	0.998	13.3	0.00919	0.40	0.1207	0.87	1.100	0.24	12.85 $\pm$ 0.12	11.90 $\pm$ 0.40	1.104 $\pm$ 0.0028
ML1-9	265.3	38.99	0.568	12.4	0.01108	0.30	0.13613	0.61	1.104	0.47	14.43 $\pm$ 0.12	13.41 $\pm$ 0.48	1.109 $\pm$ 0.0056
ML1-2a	269.9	33.62	0.746	25.6	0.00533	0.22	0.13559	0.36	1.096	0.37	14.49 $\pm$ 0.08	13.93 $\pm$ 0.24	1.101 $\pm$ 0.0042
ML1-8b	255.4	27.00	1.007	13.7	0.01064	0.20	0.14415	0.63	1.100	0.54	15.44 $\pm$ 0.14	14.41 $\pm$ 0.47	1.105 $\pm$ 0.0063
ML1-8c	213.3	53.36	5.275	129	0.00114	0.30	0.14669	0.56	1.099	0.46	15.69 $\pm$ 0.12	15.49 $\pm$ 0.13	1.104 $\pm$ 0.0053
ML1-bL	270.8	47.75	0.608	23.5	0.00693	0.44	0.16220	0.59	1.100	0.37	17.44 $\pm$ 0.13	16.78 $\pm$ 0.32	1.106 $\pm$ 0.0044
ML1-7	229.7	43.60	0.056	35.8	0.00466	0.34	0.16592	0.49	1.098	0.73	17.94 $\pm$ 0.17	17.44 $\pm$ 0.26	1.103 $\pm$ 0.0084
ML1-6	239.3	70.19	0.388	384	0.00045	1.10	0.17423	0.47	1.094	0.36	18.95 $\pm$ 0.12	18.83 $\pm$ 0.12	1.099 $\pm$ 0.0041
ML1-1a	250.4	63.44	4.819	23.6	0.00761	0.22	0.17896	0.18	1.095	0.38	19.51 $\pm$ 0.09	18.79 $\pm$ 0.33	1.100 $\pm$ 0.0044
Moaning Cave													
MC-13-16	182.7	192.8	6.071	13.8	0.01027	0.20	0.14200	0.45	1.085	0.14	15.27 $\pm$ 0.08	14.56 $\pm$ 0.74	1.089 $\pm$ 0.0022
MC-14C-1	128.1	175.9	1.145	68.5	0.00213	0.27	0.14553	0.66	1.108	0.21	15.32 $\pm$ 0.12	15.18 $\pm$ 0.19	1.113 $\pm$ 0.0025
MC-14C-4	270.0	186.5	0.970	85.1	0.00169	0.41	0.1443	1.09	1.107	0.21	15.20 $\pm$ 0.18	15.09 $\pm$ 0.22	1.111 $\pm$ 0.0024
MC-14C-7	132.8	177.9	1.818	43.8	0.00333	0.19	0.14588	0.34	1.104	0.20	15.43 $\pm$ 0.07	15.20 $\pm$ 0.24	1.108 $\pm$ 0.0024

All isotope ratios are activity ratios. Uncertainties are given at 2 standard deviations. Uncorrected ages are calculated from measured ratios. Corrected ages were corrected for initial ^{230}Th using $(^{232}\text{Th}/^{238}\text{U}) = 1.21 \pm 0.50$, $(^{230}\text{Th}/^{238}\text{U}) = 1.0 \pm 0.1$, and $(^{234}\text{U}/^{238}\text{U}) = 1.0 \pm 0.1$ for McLean's Cave and $(^{232}\text{Th}/^{238}\text{U}) = 1.45 \pm 0.73$, $(^{230}\text{Th}/^{238}\text{U}) = 1.0 \pm 0.1$, and $(^{234}\text{U}/^{238}\text{U}) = 1.0 \pm 0.1$ for Moaning Cave. Decay constants are those of Jaffey (1971) for ^{238}U and Cheng et al. (2000) for ^{230}Th and ^{234}U . Initial $(^{234}\text{U}/^{238}\text{U})$ is back-calculated from the measured ratio and the corrected ^{230}Th age.

Table 2: Spearmans rank order correlation coefficients (r_s values). Proxies have been interpolated where necessary (see text) and linear trends have been removed

	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$	Mg/Ca	Sr/Ca	Ba/Ca
$\delta^{18}\text{O}$		0.63**	0.21**	0.26**	0.26**
$\delta^{13}\text{C}$	0.63**		0.37**	0.48**	0.29**
Mg/Ca	0.21**	0.37**		0.38**	0.32**
Sr/Ca	0.26**	0.48**	0.38**		0.75**
Ba/Ca	0.26**	0.29**	0.32**	0.75**	

**p<0.001

Table 3: Results of Hendy Test. Values taken from left and right of the growth axis (axis values also reported) in dense translucent and white inclusion-rich bands

Dense Translucent Band				White Inclusion-rich band			
		$\delta^{13}\text{C}$ (‰)	$\delta^{18}\text{O}$ (‰)			$\delta^{13}\text{C}$ (‰)	$\delta^{18}\text{O}$ (‰)
<i>On Axis</i>	ML1-6.90	-7.34	-8.44	<i>On Axis</i>	ML1-7.79	-5.74	-8.10
<i>Left of Axis</i>	ML1-6HT1	-7.50	-8.23	<i>Left of Axis</i>	ML1-7HW1	-5.66	-8.31
	ML1-6HT2	-7.13	-8.61		ML1-7HW2	-5.91	-8.31
	ML1-6HT3	-7.56	-8.02		ML1-7HW3	-5.78	-6.91
<i>Right of Axis</i>	ML1-2HT1	-7.77	-8.74	<i>Right of Axis</i>	ML1-3HW1	-7.92	-7.71
	ML1-2HT2	-7.89	-8.75		ML1-3HW2	-6.82	-7.81
					ML1-3HW3	-6.03	-6.85
Average		-7.53	-8.47	Average		-6.27	-7.71
(st. dev)		0.29	0.33	(st. dev)		0.83	0.61

Figure's

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7