TRACKING REGIONAL AND GLOBAL TELECONNECTIONS RECORDED BY WESTERN NORTH
AMERICAN SPELEOTHEM RECORDS,

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Abstract

Speleothem proxy records are useful for interrogating past climates in the low and mid-latitudes given their ability to provide continuous, high-resolution, and long-lived records that can be dated with high precision. Several speleothem oxygen isotope records from western North America have recently been developed that highlight the importance of this archive in documenting past changes in atmospheric circulation. Taken individually, these records hint at teleconnections between western North American hydroclimate and climate changes in the high northern latitudes and tropics. However, there has been no systematic investigation of global climate teleconnections to this region that draws upon the body of North American speleothem records as a whole. Here we review the dominant controls on precipitation oxygen isotopes across the region, and conduct statistical comparisons and network visualizations of high-resolution speleothem oxygen isotope records from western North America to investigate the regional response to pronounced climate changes of the last deglaciation and to determine the pattern of global teleconnections to this region. We find that most western North American speleothem oxygen isotope records demonstrate a robust and consistent response to the events of the last deglaciation, despite differing controls on the oxygen isotope ratio of precipitation across the region. One record that receives a strong influence from the Gulf of Mexico exhibits a contrasting pattern in oxygen isotopes relative to most of the other records, which are dominated by westerly storms generated in the Pacific. During the studied interval, major shifts in Western North American speleothem records appear broadly synchronous or at least within the typical uncertainty of age models. We also find strong statistical linkages between

western North American speleothem records and speleothem records of Asian monsoon variability and other records from regions directly influenced by movement of the Intertropical Convergence Zone, demonstrating that nearly synchronous changes in atmospheric circulation measurably altered precipitation dynamics in these regions during the last deglaciation. Further multi-proxy investigations are needed to determine how such changes in atmospheric circulation recorded by speleothem oxygen isotope ratios influenced precipitation amounts in drought-prone western North America.

1. Introduction

Episodic droughts, and long-term shifts in hydroclimate are defining features in the climate history of semi-arid western North America (WNA) (Reheis et al., 2014; Oster et al. 2015). Since 2013, an ongoing multi-year drought in the western United States has had observable impacts on ecosystems (Bell et al., 2014; Creeden et al., 2014), wildfire occurrence (Dennison et al., 2014), and even uplift (Borsa et al., 2014). Given predictions for future severe drying and snowpack decreases in the region (e.g. Seager et al., 2007; Seager and Vecchi, 2010; Diffenbaugh et al., 2013), the potential link between the present drought and global warming is actively being investigated through analysis of historical and recent paleoclimate records (Griffin and Anchukaitis, 2014; Funk et al., 2014; Swain et al., 2014; Wang and Schubert, 2014; Mao et al., 2015; Seager et al., 2014; Diffenbaugh et al., 2015). Yet, it remains unclear whether the effects of global warming on precipitation amount are already being felt, or when they might be expected, largely due to natural variability in the ocean and atmosphere (Seager and Vecchi, 2010; Seager and Hoerling, 2014). Predictions for future drought are further hindered by complex relationships among sea surface temperature (SST) variations, changes in atmospheric circulation, and precipitation. These uncertainties are compounded by inconsistencies among climate models in reproducing these relationships (Seager et al., 2014).

Terrestrial paleoclimate proxy records provide windows into past periods of changing climate that can help to develop context and understanding for present conditions. In Western North America (WNA), there is a dynamic history of hydroclimatic change across the last glacial-interglacial transition documented by proxy records from lake sediments (e.g. Hostetler and Benson, 1990; Mensing, 2001;

Munroe and Laabs, 2013; Kirby et al., 2013) and pollen analysis (e.g. Worona and Whitlock, 1995; Whitlock and Bartlein, 1997; Wigand and Rhode, 2002). Comparisons between this network of WNA paleoclimate observations and paleoclimate model simulations have yielded important insights regarding the driving mechanisms behind WNA precipitation variability (e.g. COHMAP et al., 1988; Diffenbaugh et al., 2006; Harrison et al., 2013; Oster et al., 2015). Over the past few decades, proxy records from speleothems have proven particularly useful for interrogating past climates in the low and mid-latitudes due to their ability to provide continuous, high-resolution, and long-lived records that can be dated with high precision (Fairchild and Baker, 2012; Wong and Breecker, 2015), and several new speleothem paleoclimate records for western North America (WNA) have recently been developed (Fig. 1; Table 1). Specifically, many new speleothem oxygen isotope ($\delta^{18}\text{O}$) records cover all or part of the transition out of the last glacial period (~20 ka to present) at centennial resolution or better and are thus particularly well suited to document the response of WNA hydroclimate to both abrupt and longer term climate change and highlight important global mechanistic linkages that influence WNA precipitation variability. These records have pointed to important relationships between precipitation variability in WNA and changes in global climate including Northern Hemisphere temperatures, the strength of Atlantic Meridional Overturning Circulation (AMOC) and monsoon strength that occurred across the last deglaciation. While these relationships have been examined in the context of interpreting individual proxy records and how they reflect hemispheric and global teleconnections, there has been no attempt at a systematic investigation of the global climate linkages elucidated by the body of WNA speleothem records as a whole. Here we review recent thinking regarding WNA climate during the last deglaciation deduced from speleothem $\delta^{18}\text{O}$ records, present statistical analyses and network visualizations that highlight robust regional patterns and global teleconnections with WNA climate, and investigate the influence of chronologic uncertainty in deducing these patterns. Our aim with this analysis is to integrate available, well dated, high-resolution global speleothem records in order to place quantitative constraints on climate teleconnections to WNA across the last deglaciation. We conclude by summarizing open questions and

active areas of further investigation where speleothem records can be particularly useful in revealing the driving mechanisms of WNA hydroclimatic processes.

2. Modern Controls on WNA Precipitation

Modern precipitation patterns in WNA are variable and influenced by complex regional topography that generates large gradients in precipitation and temperature (Redmond and Koch, 1991; Palecki et al., 2005). With the exception of the north Pacific coast, most of the region is arid or semi-arid, receiving the majority of precipitation during seasonally specific rainfall events (Redmond and Koch, 1991). Much of WNA experiences a winter-dominated precipitation regime when moisture is advected zonally over the continent from the central and northern Pacific by winter cyclones derived from the region south of the Aleutian Low and carried by the winter storm track (Antinao and McDonald, 2013). Moisture from these winter storms can penetrate deeply into WNA and is the dominant source for precipitation and groundwater recharge from the west coast into the Great Basin (Friedman et al., 2002). Occasionally, extra-tropical cyclones that pull moisture directly from the central or eastern Tropical Pacific influence WNA. These systems can sometimes develop atmospheric rivers, narrow filaments of concentrated near-surface water vapor that can cause intense rainfall and are often associated with flooding along the Pacific coast (Zhu and Newell, 1998; Dettinger, 2011).

The southeastern portion of WNA also receives summer precipitation from the North American Monsoon (NAM). Heating of the Mojave and northern Sonoran deserts during the summer generates a thermal low that pulls tropical moisture from the Gulf of Mexico, Caribbean Sea, Gulf of California, and the eastern Pacific into parts of northern Mexico, southeastern California, Arizona, New Mexico, and the southern Rockies, with the dominant moisture source varying from west to east across the region (Higgins et al., 1997; Adams and Comrie, 1997; Antinao and McDonald, 2013; Metcalfe et al., 2015). The Pacific and Gulf of California are the dominant source of monsoon moisture for the Mojave and Sonoran Deserts and the southern Colorado Plateau (Higgins et al., 1997; Antinao and McDonald, 2013). To the east, the Gulf of Mexico and Caribbean provide the primary moisture source for eastern and north-central Mexico and

contribute upper-level moisture to the southwestern United States. The Great Plains low-level jet also carries moisture through eastern Mexico and into the Great Plains (Metcalf et al., 2015). Most monsoon precipitation occurs as small isolated thunderstorms. In general, the proportion of annual precipitation contributed by monsoon rains increases from west to east across the region (Hereford et al., 2006; Antinao and McDonald, 2013).

Precipitation patterns in WNA are strongly influenced by large-scale patterns of ocean-atmosphere interaction in the tropical and northern Pacific such as the El Niño-Southern Oscillation (ENSO) and the Pacific Decadal Oscillation (PDO). Comparisons of historical records of precipitation variability in WNA have revealed a dipole or seesaw pattern, with the Pacific Northwest and desert southwest displaying opposing relationships to ENSO indices (Redmond and Koch, 1991; Cayan et al., 1999). Cool season (October-March) precipitation displays a lagged relationship behind the summer and fall Southern Oscillation Index (SOI), with positive SOI (La Niña conditions) corresponding with reduced precipitation in the desert southwest and increased precipitation in the Pacific Northwest and vice versa (Redmond and Koch, 1991). This dipole pivots on a transition zone between 40 and 42°N (Wise, 2010). The PDO (Mantua and Hare, 1997) is thought to modulate the SOI-precipitation relationship on decadal timescales and does appear to modify the shape of the dipole transition zone through time (Wise, 2010).

Speleothem records covering the last glacial-Holocene transition in WNA with centennial or finer temporal resolution (see below) spread from 42.09 to 29.88°N and 120.4 to 98.6°W with Oregon Caves (OC) representing the furthest site to the northwest and Cave Without a Name (CWN) representing the furthest site to the southeast (Fig. 1). Among this group, sites to the northwest in Oregon, California, and the Great Basin (OC, Moaning Cave [MC], McLean's Cave [ML], Leviathan Cave [LV], and Pinnacle Cave [PC]) receive the majority of precipitation and groundwater infiltration from winter cyclones (Oster et al., 2009; 2015b; Vacco et al., 2005; Ersek et al., 2013; Lachniet et al., 2011; 2014). Sites further to the southwest (Cave of the Bells [COB], Fort Stanton [FS], and Pink Panther [PP]) are influenced by the NAM, and receive a measureable proportion of precipitation during the summer. Roughly half of annual

precipitation at COB is derived from the summer monsoon (Wagner et al., 2010), while more than half of annual precipitation at FS comes from monsoon rains (Asmerom et al., 2010). Site CWN in central Texas receives precipitation primarily from rainfall events originating from the Gulf of Mexico (Feng et al., 2014) with occasional influence from tropical cyclones traveling either from the Pacific or the Gulf of Mexico (Pape et al., 2010). All of these cave sites are located in the region that experiences a positive correlation between cool season rainfall and SOI with the exception of OC, which is in the southernmost portion of the region that experiences the opposite correlation, and MC and ML which lie in the transition zone (Wise et al., 2010).

3. Controls on Speleothem $\delta^{18}\text{O}$ variability in WNA

The overall stability of cave environments on long timescales allows speleothems to record variations in the $\delta^{18}\text{O}$ of cave drip waters, which are often closely related to the $\delta^{18}\text{O}$ of rainfall above a cave. Such records of oxygen isotope variability in speleothems worldwide have provided powerful tools for the investigation of global climate teleconnections, highlighting links between precipitation in the low and mid-latitudes and insolation, ocean circulation, and ocean-atmosphere oscillations on seasonal to orbital timescales (e.g. Cheng et al., 2009; 2012; Meckler et al., 2012; Myers et al., 2015). Comprehensive reviews of the general processes governing $\delta^{18}\text{O}$ variability in speleothems have been recently produced by McDermott (2004) and Lachniet (2009). Here we provide a brief summary of recent work on precipitation and cave drip water $\delta^{18}\text{O}$ in WNA specifically that supports interpretations of WNA speleothem $\delta^{18}\text{O}$ records.

Studies of $\delta^{18}\text{O}$ and stable hydrogen ($\delta^2\text{H}$) isotope ratios in precipitation, surface water, and groundwater in WNA have provided important insights into moisture transport variability that have proven essential for understanding water resources in this semi-arid region. Through dense sampling and analysis of $\delta^2\text{H}$ in spring and stream water along a W-E transect through northern California and southern Oregon and northern Nevada, Ingraham and Taylor (1986) documented a balance in the control on water isotope ratios between Rayleigh distillation of water vapor and recycling of terrestrial water through

evapotranspiration. While Rayleigh distillation is important near the coast, recycling of meteoric water becomes increasingly important as weather systems move inland (Ingraham and Taylor, 1986; Wassenaar et al., 2011; Winnick et al., 2014). In general, the range of annual variability in precipitation $\delta^{18}\text{O}$ and $\delta^2\text{H}$ is small along the coast and increases inland (Vachon et al., 2010). Along the west coast, observations of precipitation on the seasonal to event scale point to a seasonal signal in water isotope ratios with summer rain characterized by higher isotope ratios and winter rain characterized by lower isotope ratios (Vachon et al., 2010; Ersek et al., 2010; Oster et al., 2012). This seasonal signal has also been observed in cave drip waters in near-coastal environments (Oster et al., 2012; Beddows et al., 2015). Analyses of rain and snow isotopes along the coast from central to southern California indicate that moisture source plays a dominant role in determining precipitation isotopic ratios, with subtropical and tropical Pacific sourced moisture leading to precipitation with higher isotope ratios and mid-latitude and north Pacific sourced moisture leading to lower isotope ratios (Friedman et al., 2002; Berkelhammer et al., 2012). However, Ersek et al. (2010) found that surface air temperature at the time of collection is more closely related to precipitation $\delta^{18}\text{O}$ than moisture source in southwestern Oregon. In the central Sierra Nevada foothills of California, Oster et al. (2012) determined that both temperature and moisture source influence rainwater $\delta^{18}\text{O}$, but observed that the longer-lived isotopic signals from temperature changes were more likely to be transmitted to cave drip waters over the event-scale moisture source signals. Temperature and moisture source have also been shown to be strong influences on precipitation isotopes further inland in the Great Basin (Friedman et al., 2002; Lachniet et al., 2014). Buenning et al., (2012; 2013) used isotope-enabled modeling to determine the controlling factors on seasonal and interannual precipitation isotope signals observed on the west coast. By performing sensitivity experiments with the IsoGSM model, they determined that variations in droplet condensation height due to seasonal changes in the polar jet are the primary driver of the observed seasonal signal in precipitation isotope ratios. However, how these seasonal-scale processes such as variation in condensation height would vary over long timescales and be transmitted to cave drip waters and recorded by speleothems is unclear.

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178 Moving inland into southwestern WNA the balance of precipitation from isotopically distinct winter
179 westerly and summer monsoonal precipitation sources becomes a more important control on the isotopic
180 composition of precipitation. However, comparison of the isotopic signatures of precipitation and spring
181 waters in the Spring Mountains, southern Great Basin, Nevada, suggests that intense but short summer
182 storms contribute 10% or less of annual recharge (Winograd et al., 1998). In Arizona and New Mexico,
183 monitoring of cave drip water and modern calcite at both COB and FS, indicates that intense summer
184 evaporation prevents infiltration of monsoon rains, also suggesting that the majority of infiltration that
185 would presently lead to speleothem growth is derived from winter storms (Asmerom et al., 2010; Wagner
186 et al., 2010). In central Texas, assessment of $\delta^{18}\text{O}$ in rainwater determined that precipitation amount (the
187 amount effect) is an important control on rainfall $\delta^{18}\text{O}$ values during the summer but not the winter and
188 documents that the lowest rainwater $\delta^{18}\text{O}$ values are associated with tropical cyclones (Pape et al., 2010).

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190 Interpretations of observed variability in WNA speleothem $\delta^{18}\text{O}$ over decadal to glacial-interglacial
191 timescales are in part based on these seasonal to interannual observations of precipitation, surface and
192 ground water, and drip water $\delta^{18}\text{O}$. Near the coast, $\delta^{18}\text{O}$ records from OC are interpreted to reflect
193 changes in surface air temperatures that influenced rainwater $\delta^{18}\text{O}$ (Vacco et al., 2005; Ersek et al., 2012).
194 For central Sierra Nevada speleothems from MC and ML, temperature changes are also the thought to be
195 an important control on speleothem $\delta^{18}\text{O}$, however, the potential influence of moisture source variability
196 is acknowledged (Oster et al., 2009; 2014; 2015b). Slightly further south in the Sierra Nevada, a modern
197 (last ~1200 years) speleothem $\delta^{18}\text{O}$ record from Crystal Cave is interpreted to reflect changes in moisture
198 source that are ultimately driven by Pacific sea surface temperature changes that influence storm track
199 trajectories (McCabe-Glynn et al., 2013). Moving further inland, speleothem $\delta^{18}\text{O}$ from LV, PC, and
200 Lehman Cave in the Great Basin are interpreted to reflect a combination of changes in temperature and

rainout history of winter storm systems that are closely aligned with northern hemisphere summer insolation on orbital timescales (Lachniet et al., 2004). In the southwest, speleothem $\delta^{18}\text{O}$ records from COB, FS, and PP are interpreted to reflect the balance of contribution of winter (relatively depleted in ^{18}O) versus summer (relatively enriched in ^{18}O) precipitation, with variations mainly attributed to changing inputs of winter precipitation from the Pacific (Wagner et al., 2010; Asmerom et al., 2010). However, the potential influence of the amount effect in lowering precipitation $\delta^{18}\text{O}$ in both seasons is acknowledged in the mostly Holocene record from PP (Asmerom et al., 2008). In central Texas, the $\delta^{18}\text{O}$ record from CWN is thought to reflect changing $\delta^{18}\text{O}$ values of the primary moisture source region, the Gulf of Mexico, due to variable inputs of melt water from the Laurentide Ice Sheet (Feng et al., 2014). Feng et al., (2014) go on to suggest expansion of the influence of Gulf of Mexico moisture might play a more important role in determining speleothem $\delta^{18}\text{O}$ at COB, and especially FS than changing amounts of winter precipitation from the Pacific during the last deglaciation.

In the interpretation of each of these speleothem records, care must be taken to account for possible influences of changing cave air temperatures and kinetic effects on speleothem $\delta^{18}\text{O}$. Additionally, when records cover periods of changing global ice volume such as the last deglaciation, speleothem $\delta^{18}\text{O}$ values should be corrected to account for this effect on the $\delta^{18}\text{O}$ value of ocean source waters (McDermott 2004; Lachniet, 2009). Concern for potential changes in cave air temperature, which has the opposite effect on the $\delta^{18}\text{O}$ of calcite precipitating in equilibrium than the temperature influence on the $\delta^{18}\text{O}$ of water during condensation (Kim and O'Neil, 1997), is often mitigated by choosing speleothems from restricted sites far removed from cave entrances where temperature variations should be minimized. The potential influence of changing cave temperatures on the speleothem $\delta^{18}\text{O}$ record can be estimated based on cave monitoring data or other coeval temperature-sensitive proxy records (e.g. Oster et al., 2014). However, recently, it has also been proposed that $\delta^{18}\text{O}$ records from speleothems growing in well-ventilated cave sites where

cave air temperature varies closely with surface air temperature might provide a resource for quantitative paleo-temperature reconstructions (Feng et al., 2014b). The potential influence of kinetic effects during calcite precipitation can obscure climate signatures in speleothem stable isotope records. Such effects are monitored by conducting Hendy tests, which include measurements of oxygen and carbon isotope variability along growth bands (Hendy, 1971), replication tests that involve the development of $\delta^{18}\text{O}$ records from multiple coeval speleothems from the same cave (Dorale and Liu, 2009), and comparison of modern calcite/drip water pairs and/or farmed calcite during cave monitoring (e.g. Mickler, 2004; Tremaine et al., 2011). In cases where these tests reveal that speleothem $\delta^{18}\text{O}$ is not a reliable climate proxy due to kinetic influences, other proxies such as trace element variability are often utilized to assess past climates (e.g. Steponaitis et al., 2015). Even when strong influence of kinetic effects can be accounted for, speleothem $\delta^{18}\text{O}$ values can still be influenced by processes that might modify or smooth climate signals such as evaporation in the soil zone or mixing of vadose waters in the epikarst (Lachniet, 2009), and practices such as cave monitoring and geochemical modeling can assist in determining the relative importance of these processes at a given site (e.g. Bradley et al., 2010; Moerman et al., 2014).

In the following discussion, we statistically compare a network of well-dated, high-resolution records of speleothem $\delta^{18}\text{O}$ variability in WNA and globally and utilize these results and the authors' published interpretations to investigate regional and global teleconnections across the last deglaciation. In our analysis, we focus on the multi-centennial to millennial-scale variability of the last deglaciation, which encompasses events such as the Younger Dryas, the Bølling-Allerød, and Heinrich Stadial 1. Variability at this temporal scale is often clearly recorded in WNA and global records with event durations that fall outside typical dating uncertainties for most speleothem records, allowing the assessment of synchronicity in regional and global records and of the influence of chronological uncertainty on event correlation. Our goal is to provide a comprehensive and systematic view of precipitation $\delta^{18}\text{O}$ variability and

teleconnections that moves beyond visual and qualitative comparison to elucidate robust regional and global patterns.

4. Methods

4.1 Record selection

We began with the compilation of global speleothem $\delta^{18}\text{O}$ records since the Last Glacial Maximum (LGM) published by Shah et al., (2012) that is based on the speleothem paleoclimatology data set hosted by the NOAA National Climate Data Center (NCDC). We updated this dataset with published speleothem $\delta^{18}\text{O}$ records for WNA that were archived on the NCDC website after December 2011. Our compilation includes all WNA records of vadose zone speleothem $\delta^{18}\text{O}$ variability published before November 2015 (Fig. 1; Table 1). We then selected records from this compilation that cover at least 3,000 years of the transition between the LGM and Holocene (9 to 22 ka) at 200-year or finer temporal resolution in order to provide sufficient data to evaluate response to multi-centennial to millennial scale changes (Fig. 2). Records in the Shah et al., (2012) compilation were previously standardized to the calBP timescale where present is 1950, and we standardized additional records to match this timescale. We did not include the recently published Lehman Cave record of Steponaitis et al., (2015), as the authors determined the stalagmite $\delta^{18}\text{O}$ record was likely kinetically modified.

4.2 Statistical comparisons

Time series analysis of irregularly sampled records such as speleothems is challenging as most conventional time series analysis techniques are designed for continuous, regularly sampled records. Linear interpolation is commonly used to convert irregular time series to regular time series, however this approach leads to bias (Rehfeld et al. 2011). Here, we follow an approach developed by Rehfeld et al. (2011) and Rehfeld and Kurths (2014) which uses Gaussian-kernel based smoothing to generate regularly sampled estimates that follow the pattern of change observed in the original irregular time series (Fig. 2). Pearson correlation of these Gaussian smoothed records, termed Gaussian-kernel-based cross-correlation (gXCF), allows positive or negative correlations between time series to be identified. We used the

MATLAB toolbox, NESToolbox (Rehfeld and Kurths, 2014) to calculate gXCF estimates for each pairwise comparison of WNA speleothem records. We also compared WNA records to other key deglacial global climate records including the NGRIP Greenland ice core (Vinther et al., 2006; Rasmussen et al., 2006) and West Antarctic Ice Sheet $\delta^{18}\text{O}$ record (WAIS Members, 2015), and the Cariaco Basin record of reflectance (Peterson et al., 2000). In addition, we compared these records and the WNA speleothem records to other deglacial speleothem $\delta^{18}\text{O}$ records from around the world. We screened our updated global speleothem record database for records that cover at least 3,000 years of the last deglaciation and exhibit 200-year or finer temporal resolution, which resulted in the selection of fifteen speleothem $\delta^{18}\text{O}$ records from Asia, the Middle East, South America, Indonesia, northern Australia, and New Zealand (Fig. 3; Table 2). All records were trimmed to the focal interval of the last deglaciation (9-22ka) prior to analysis.

We applied linear detrending to the raw data series prior to gXCF analysis by taking the residuals from a linear function $y=a+bt$ fitted to the original data series. This was done to remove correlations between records that simply reflect long-term change in the climate system from the LGM to the Holocene. Pairs of these detrended data-series were compared using the similarity function in NESToolbox using the gXCF option with bandwidth selection following the recommended $h=0.25$ for the common sampling interval on the rescaled time-axis (Rehfeld and Kurths, 2014) which is the default setting within NESToolbox. Time series data are centralized and standardized within NESToolbox prior to gXCF calculation. Significance of gXCF values was evaluated by comparison with independent AR(1) autocorrelated but mutually uncorrelated surrogate time series generated using NESToolbox. Correlation values that fall outside the 5% and 95% quantiles drawn from 2000 replicate surrogate testing are accepted as significant (Rehfeld and Kurths, 2014).

Network graphs were generated for visual comparison of gXCF correlations using the qgraph package in R (Epskamp et al., 2015). Node placement on these graphs is determined geographically or using the “spring” layout in qgraph that uses the Fruchterman-Reingold algorithm to generate a force directed layout based on linkage strength.

4.3 Accounting for age uncertainty in proxy records

Although $^{230}\text{Th}/\text{U}$ dating of speleothems is capable of producing paleoclimate proxy records with precisely dated chronologies, each $^{230}\text{Th}/\text{U}$ age possesses some amount of uncertainty, and this uncertainty is transferred to the proxy time series. Given this uncertainty, it can be challenging to evaluate the robustness of correlations, leads, and lags between paleoclimate records, particularly at centennial or finer scales. Thus, we followed two approaches to account for age uncertainty and investigate its influence on correlations between WNA speleothem $\delta^{18}\text{O}$ records:

4.3.1 Lagged comparisons

Each of the pairwise comparisons described previously was assessed at zero lag between the two records. However, we also computed cross-correlation functions on the smoothed records across a range of lag values chosen to span the maximum age uncertainty within the overlap of the two compared records using the gXCF similarity function in NESToolbox. We then compared the strength and significance of maximum and minimum gXCF values and noted the lags at which these values occurred to determine the strongest correlation within age uncertainty. Significance testing of these lagged gXCF estimates was performed following the procedure given above, however the significance threshold was adjusted for multiple testing by dividing the stated alpha level (0.05) by the number of lags for which gXCF was estimated within the maximum age uncertainty window.

4.3.2 Comparisons with multiple age model realizations

We followed the approach of Rehfeld and Kurths (2014) to further evaluate the influence of age uncertainty on statistical comparisons between WNA speleothem records. To do this, we re-evaluated the age models of WNA records using the COPRA algorithm (Breitenbach et al., 2012) that uses a Monte

Carlo approach to generate an ensemble of potential age models before determining the most appropriate age model for a given time-series. We were able to generate an ensemble of age models for the cave records for which authors provided the depth below stalagmite surface for both $^{230}\text{Th}/\text{U}$ and $\delta^{18}\text{O}$ measurements either within the NCDC database or in the online data repositories for given articles. Thus, we were capable of generating multiple age model realizations using COPRA for COB, PP, CWN, MC, and ML. For each of these records, we input the age, age uncertainty and proxy depth data, and used the pchip interpolation method in COPRA to generate 1000 age model realizations for each record. We then conducted the statistical tests outlined previously to compare each realization to the published time series for the other WNA records, analyzed the resulting distribution of computed gXCF values, and compared this outcome to the lagged and unlagged values calculated from the published age models for each record.

5. Results

5.1 Pairwise comparisons of WNA records at zero-lag

The results of the pairwise comparisons between WNA speleothem records are given in Table 3. Figure 4 depicts the relationships among WNA records in two different ways. Figure 4A shows the records in a geographic arrangement where each cave location is represented by a circle, or node. The lines connecting nodes represent correlations that are statistically significant outside the 95% confidence intervals, where red lines represent negative correlations, and blue lines represent positive correlations and the strength of the correlation is shown both through the thickness of the line and the depth of color. Figure 4B displays the same correlations but the nodes are rearranged so that those with more solid connections to other records are placed at the center of the arrangement, and those with fewer connections are placed on the periphery.

The majority of significant correlations among WNA records at zero-lag are positive (blue lines in Fig. 4). These significant positive correlations range from strong (e.g., FS vs. OC, $\text{gXCF} = 1$) to moderate (e.g., LV vs. OC, $\text{gXCF} = 0.45$) (all confidence intervals are reported in Table 1 and shown graphically in the

Supplemental Material). Only CWN consistently exhibits moderate to strong negative correlations with other WNA records, including with OC ($gXCF = -0.92$), LV ($gXCF = -1$), and COB ($gXCF = -0.79$). However, CWN also displays a significant moderate positive correlation with ML ($gXCF = 0.44$). ML also displays a weak negative correlation with PP ($gXCF = -0.32$) and a positive correlation with its neighbor MC ($gXCF = 0.37$). When arranged by centrality, a key group of five speleothem records demonstrating the strongest correlations emerges. These include COB, LV, OC, FS, and CWN (Fig. 4B). The remaining records (MC, PP, ML, and PC) surround this cluster, linked by weak to moderate correlations with each other and with records within the strongly correlating cluster, with the exception of PC, which displays no significant correlations with other records. The ranges of overlap of PP with FS, CWN and COB were each less than 1000 years and therefore deemed too small to be tested (Table 1).

5.2 Lagged comparisons

The lagged comparison results indicate that, for the majority of comparisons between the five strongly interlinked records (OC, FS, COB, LV, and CWN), the strongest correlations occur within ~ 100 years of zero lag (FS vs. LV, FS vs. OC, LV vs. CWN, LV vs. OC, COB vs. OC) or improve only mildly with lagging (OC vs. CWN). Correlations between COB vs. FS and COB vs. CWN both improve when COB is lagged forward in time (within the maximum observed age uncertainty) relative to the other records. The positive correlation between COB and LV is not significant in the lagged analysis despite displaying a significant positive correlation at zero lag. This results from the more conservative confidence intervals in the lagged analysis that are corrected for multiple testing. However, examination of the cross correlation function (Supplemental Material) suggests that the positive correlation between these two records is stable across multiple lags within age uncertainty, suggesting the relationship is not likely to be a spurious product of repeated hypothesis testing. For the more weakly linked records (MC, PP, ML, and PC), a few additional significant correlations emerge if one record is lagged within the window of maximum age uncertainty (Fig. 4C, D). Generally, these records have the larger age uncertainties among the WNA records (± 450 to 1000 years in places), and many maximum correlations fall within these larger age uncertainty windows. Even when lagged within uncertainty, however, PC and MC exhibit

correlations only with ML. ML displays lagged negative correlations with LV, FS and COB when ML is lagged by ~300 to ~500 years. The negative relationship between PP and ML observed at zero-lag is strengthened when ML is lagged nearly 1000 years ahead, which is just within the maximum age uncertainty of these two records.

5.3 Age model realizations

Statistical comparisons between WNA time-series with multiple age-model realizations computed using the COPRA algorithms result in distributions of gXCF that generally agree with those computed for the lag zero and lagged comparisons based on published age models. Two illustrative examples are shown in Figure 5 which displays histograms of gXCF values computed for comparisons of the COB COPRA realizations with the published FS record (Fig. 5A, B) and the published MC record (Fig. 5C, D). The comparisons of FS with the COB COPRA realizations produce a majority of gXCF values in line with the gXCF value computed at zero lag (Fig. 5B). Comparison between the published MC record, which has larger age uncertainties, and the COB COPRA realizations, produces a bimodal distribution, with one maximum centered near the gXCF value computed at zero lag, which is weakly positive but not significant (Fig. 5D). The other maximum falls at a higher gXCF value (0.6), outside of the 95 % confidence interval, indicating that there may be a significant positive correlation between these two records that is detected under certain age model parameters but is otherwise obscured by uncertainties in the age models. Computed gXCF values for comparisons with COPRA age model realizations of records with larger uncertainties, such as PP and ML, tend to cluster around 0.

5.4 Global record comparisons

For the comparison of WNA records against global paleoclimate records, we focused on global connections to COB, FS, and CWN as these three records display strong and consistent correlations with most other WNA speleothem records, cover the majority of the deglacial period, and display strong signals for the transition into the Bölling-Alleröd and into and out of the Younger Dryas (Fig. 2). We

included CWN despite the large hiatuses in that record primarily because of the consistent, moderate to strong negative correlations it displays with most other WNA records.

Of these three WNA records, COB displays moderate to strong negative correlations with speleothem records from East Asia (Hulu, Dongge, and Yamen Caves in China) as well as Moomi Cave in Oman, Soreq Cave in Israel, and Nettlebed Cave in New Zealand (Fig. 6A). COB shows moderate to strong positive correlations with NGRIP, and Ballgown Cave in the southern Hemisphere tropics. FS shows similar relationships with global records as observed with COB, but also shows negative correlations with Timta and Sanbao Caves in Asia and the Cariaco Basin sediment reflectance record and a positive correlation with El Condor Cave in Peru (Fig. 6B). Consistent with the negative relationships observed with most North American records, CWN generally shows the opposite relationship to what is observed with COB and FS, displaying moderate to strong positive correlations with Asian monsoon records (Timta, Hulu, Dongge, Yamen, Sanbao) and with Moomi Cave and the Cariaco Basin (Fig. 6C). CWN also displays moderate to strong negative correlations with Liang Luar and Ballgown Caves and NGRIP.

The network diagram of these global records reveals a tight cluster of positively correlated Asian Monsoon speleothem records (Fig. 7A; Supplemental Material), the Cariaco Basin, Moomi Cave in the Indian Ocean. FS, COB, NGRIP, and Ballgown Cave are also closely linked with this tight cluster of monsoon records via moderate to strong *negative* correlations, while CWN is linked via a positive correlation. South American records (El Condor, Botuvera, and Santiago Caves) plot together, and are linked to the Asian monsoon group through the Cariaco and NGIP records. The Gunung Buda record from Borneo plots with these records due to a strong negative correlation with Botuvera Cave. The Soreq, Nettlebed, Liang Luar, and Sofular speleothem records are the least tightly linked to the global ensemble. The Antarctic ice core record, WAIS, is primarily linked through weak but significant positive correlations with multiple Asian monsoon records. Lagging one record relative to another within their shared maximum age uncertainty allows records to be tuned, leading to stronger overall correlations (Supplemental Material). However, the relatively large age uncertainties on some of the global records

(>1000 years in some instances) can lead to the appearance of contradictory relationships between lagged records, especially in this deglacial interval where many significant events (e.g. HS1, YD) occur at the millennial scale.

6. Discussion

6.1 Relationships among WNA records

The network visualization of correlations between WNA records (Fig. 4A,B) reveals strong relationships among records across the last deglaciation, with the strongest correlations evident between five records: FS, COB, LV, OC, and CWN. These five records possess some common characteristics that explain these strong correlations. These records generally have smaller age uncertainties, with maximum 2σ errors on $^{230}\text{Th}/\text{U}$ dates ranging from 100 to 400 years in the overlapping portions of records used for correlation calculations (Fig. 2). These records also cover the transition into the Bölling warm period (BO) (~14.5 ka), and/or the transitions in and out of the Younger Dryas cold period (YD) (~12.9-11.7 ka) and display relatively large $\delta^{18}\text{O}$ changes across these transitions. By contrast, the speleothem $\delta^{18}\text{O}$ records that are less well correlated with other WNA records (PP, PC, MC, ML) tend to have larger maximum age uncertainties (typically 400-500 years, or 1000 years for PP). Additionally, the PC and PP records do not cover the BO or YD transitions. In the Sierra Nevada records, ML covers the BO and the onset of the YD, but displays noisier changes in $\delta^{18}\text{O}$ across significant climate shifts than is observed in the five most tightly correlated records. MC also does not demonstrate sustained shifts in $\delta^{18}\text{O}$ associated with the BO as are shown in the COB and FS records, but rather shows centennial-scale oscillations following a positive $\delta^{18}\text{O}$ shift at the BO onset. Possibly, relatively fast transit times for seepage waters from the surface in the small Sierra Nevada caves (Oster et al., 2015b) and/or more frequent shifts in precipitation source generate noisier proxy records that make statistical comparisons with other records more challenging. Thus, small age uncertainty, temporal coverage of some or all of the major events of the

deglaciation, and the presence of smooth, measureable shifts in $\delta^{18}\text{O}$ across these events are key to producing strong correlations between WNA records.

In general, allowing lags to vary within age errors strengthens correlations between records that are already significant at zero lag. This suggests that climate events recorded by the five most tightly correlated records are tunable within age uncertainty and that these events occurred essentially synchronously between records. Understanding correlations with those records with larger age uncertainties (MC, ML, PP) is more challenging, as these records can exhibit both positive *and* negative significant correlations with other records at large lag values that fall within uncertainty (Supplemental Material). Additionally, lagging can account for consistent offsets, but cannot capture variable offsets that may occur due to changing age uncertainties along the timespan of a record that may result in squeezing or stretching of an age model. The COPRA age model realizations for records with larger age uncertainties speak to this, as the majority of comparisons lead to widely ranging gXCF values that peak near zero (Supplemental Material). This further highlights the need for good age control in evaluating mechanistic links between records.

The onset of the BO warm period is covered by three of the five tightly correlated records: FS, COB, and CWN. MC and ML also cover this interval. FS and COB are positively correlated with each other. COB is negatively correlated with CWN, and FS and CWN also display a negative correlation that is just below the significance threshold (Table 3). In general, both MC and ML display much smaller changes in $\delta^{18}\text{O}$ across the BO transition than FS and COB (~1.5‰ for MC vs. ~3‰ for COB and FS), and these changes are manifest in MC and ML as increases leading up to the BO, followed by decreases within the BO interval (Fig. 2). Each of these locations receives the majority of precipitation from winter storms originating from the Pacific. The MC and ML records, which are positively correlated with each other, are interpreted to reflect increasing temperatures and/or a shift to a more subtropical moisture source at the BO onset (Oster et al., 2009; 2015b). The FS and COB records are interpreted to reflect a decrease in

the proportion of winter precipitation (Asmerom et al., 2010; Wagner et al., 2010), and/or an increase in moisture from the Gulf of Mexico (Feng et al., 2014) that is sustained for the BO interval, with up to half of the $\delta^{18}\text{O}$ ascribed to temperature change (Asmerom et al., 2010).

There are a greater number of records that cover the YD cold event, including four of the five most tightly correlated records: COB, FS, LV and OC. CWN covers the onset of the YD before a hiatus that occurs during the height of the event (Fig 2). MC also covers the YD, and ML covers the YD onset. Most WNA speleothems record pronounced $\delta^{18}\text{O}$ decreases during the YD. In the OC and LV records, this has been interpreted as a decrease in temperature (Vacco et al., 2005; Lachniet et al., 2014), and temperature may have contributed to the decrease noted in the MC record (Oster et al., 2009). In the FS and COB records, this is interpreted as an increase in the proportion of moisture from winter westerly storms (Asmerom et al., 2010; Wagner et al., 2010). In contrast, CWN records a pronounced increase at the beginning of the YD that is interpreted as changes in the $\delta^{18}\text{O}$ of Gulf of Mexico source waters (Feng et al., 2014). Larger isotopic shifts are again observed for inland (LV, COB, FS) versus coastal sites (OC, MC, ML) for the YD.

Comparatively few WNA speleothem records provide good coverage of Heinrich Stadial 1 (HS1, ~14.5-18 ka). This is unfortunate, as many Great Basin lakes reached their maximum extent during this interval, suggesting it was the wettest portion of the last glacial period in WNA (Broecker and Putnam, 2012; Munroe and Laabs, 2013; Ibarra et al., 2014; Reheis et al., 2014). Of WNA speleothem records, only COB, FS, ML, and CWN provide good coverage, while PC and MC cover only part of the interval (Fig. 2). Many of the records that do cover this interval display significant correlations with each other (ML, CWN, FS and PC when lagged), due to prominent shifts in many speleothem $\delta^{18}\text{O}$ records at the HS1-BO transition. Most of these records (CWN, FS, ML) display a shift in $\delta^{18}\text{O}$ values at the beginning of HS1,

501 followed by positive or negative $\delta^{18}\text{O}$ excursions within the interval. In contrast, COB shows only a slight
502 shift in $\delta^{18}\text{O}$ at the beginning of HS1 followed by minimal variability thereafter.

503
504 Together, the moderate to strong positive correlations for the majority of WNA speleothem $\delta^{18}\text{O}$ records
505 covering the BO and YD, and the strengthening of these correlations when records are allowed to lag
506 within age uncertainty, indicate a robust regional response to the events of the last deglaciation, despite
507 somewhat different mechanistic interpretations of speleothem $\delta^{18}\text{O}$ among records. The notable exception
508 to this is CWN, which displays strong negative correlations with the majority of other records, and is
509 located furthest to the south and east of the WNA records. Isotope-enabled modeling of interannual
510 precipitation variability in WNA (Buenning et al., 2013) may help to reconcile the variable interpretations
511 of speleothem $\delta^{18}\text{O}$ in this region. Although modeling determined that interannual variations in
512 precipitation $\delta^{18}\text{O}$ are largely controlled by droplet condensation height and further amplified by post-
513 condensation processes, variations in moisture source and temperature were also found to be important in
514 determining precipitation $\delta^{18}\text{O}$ to variable degrees across the region (Buenning et al., 2013). Variations in
515 the proportion of subtropical versus mid-latitude moisture advection were important in determining
516 precipitation $\delta^{18}\text{O}$ along the west coast, with increasing influence further north in the Pacific Northwest.
517 Further inland, the effect of temperature on isotopic rainout increases in importance (Buenning et al.,
518 2013). These modeling results suggest that coastal sites such as OC, MC, and ML should be more
519 sensitive to the proportion of mid-latitude versus subtropical moisture in winter storms, whereas an inland
520 site, such as LV should be more sensitive to the influence of temperature. This is consistent with
521 interpretations of speleothem records from MC, ML, and LV (Oster et al., 2015; Lachniet et al., 2014),
522 but suggests that moisture source, rather than temperature, may have a greater influence on the OC record
523 than previously estimated (Vacca et al., 2005; Ersek et al., 2012). This increased sensitivity to changing
524 moisture source from the Pacific may provide a partial explanation for the noisier nature of the Sierra

Nevada speleothem $\delta^{18}\text{O}$ records compared to inland records where source signals may be modified or smoothed by distillation that occurs during vapor transport. The model does not predict a strong correlation between Pacific-sourced mid-latitude vs. subtropical moisture or temperature and precipitation $\delta^{18}\text{O}$ for sites in Arizona or New Mexico, and model results were not evaluated for most of Texas (Buenning et al., 2013). Further isotope-enabled modeling will be necessary to validate interpretations of speleothem records from the southwest and evaluate the role of moisture from the Gulf of Mexico in determining precipitation $\delta^{18}\text{O}$. Isotope-enabled paleoclimate models may also assist in investigating the influence of vapor condensation height and post-condensation processes on precipitation and speleothem $\delta^{18}\text{O}$ variations on longer timescales.

Thus, despite these somewhat differing controls on $\delta^{18}\text{O}$, WNA speleothem records point to consistent changes across the region during the last deglaciation. The BO was characterized by a reduction in mid-latitude sourced winter westerly storms along the west coast and potentially an incursion of Gulf of Mexico moisture in the southwest. The YD was characterized by a resurgence of mid-latitude sourced winter westerly storms along the coast, decreased temperatures leading to increased isotopic rainout as storms traveled inland, and a potential reduction of Gulf of Mexico vapor in the southwest. Given that speleothem $\delta^{18}\text{O}$ in WNA primarily reflects changes in atmospheric circulation, additional proxies must be used in order to determine what these changes meant for rainfall amounts. For example, analysis of carbon isotopes and trace elements in OC, MC, and ML point to wetter conditions along the coast during the YD (Oster et al., 2015; Ersek et al., 2012). Trace element (Mg/Ca) and $\delta^{13}\text{C}$ records from Lehman Caves in Nevada also suggests a wet YD (Steponaitis et al., 2015), and this record, together with MC, indicates these wet conditions lasted well into the early Holocene (Oster et al., 2009). Faster growth rates at COB may also point to a wetter YD in Arizona (Wagner et al., 2010; Feng et al., 2014), however speleothem growth rate may also be influenced by changes in soil to cave pCO_2 gradients in addition to water supply (Banner et al., 2007). Records of speleothem $\delta^{234}\text{U}$ and $\delta^{13}\text{C}$ from FS suggest a relatively

dry YD in New Mexico (Polyak et al., 2012), a result that is supported by slower growth rates for FS speleothems, but conflicts with faster growth rates noted in three other New Mexico caves (Polyak et al., 2004). A dry YD is also indicated in Texas by a growth hiatus in the CWN record (Feng et al., 2014). Conversely, shifts in $\delta^{13}\text{C}$, trace elements, and Sr isotopes at MC indicate drier conditions along the coast during the BO. A small increase in COB growth rate during the BO suggests conditions that may have been slightly wetter than previously. Maximum growth rates at FS and CWN during the BO potentially indicate wetter conditions accompanied the incursion of Gulf moisture into the southwest at this time (Feng et al., 2014). Wetter conditions are also supported by relatively low values of $\delta^{234}\text{U}$ and $\delta^{13}\text{C}$ at FS that began prior to the BO onset during Heinrich Stadial 1, but persisted until ~14 ka (Polyak et al., 2012). Therefore, speleothem multi-proxy records suggest relatively wet conditions during the YD and dry conditions during the BO for all but the most southeastern WNA records (CWN, FS) that were likely most influenced by Gulf of Mexico moisture. Likewise, the $\delta^{13}\text{C}$ record from ML and the $\delta^{234}\text{U}$ record from FS also point to wetter conditions in WNA during HS1 (Oster et al., 2015b; Polyak et al., 2012). Including additional proxies in future WNA speleothem studies will assist in reconstructing moisture availability across this region and determining how changes in rainfall amount relate to the changes in moisture source predicted by $\delta^{18}\text{O}$ records.

6.2 Global teleconnections with WNA

The three WNA records that were compared with other global records (COB, FS, CWN) also display strong and consistent global correlations (Fig. 6). The most robust relationship to emerge from this comparison is a strong negative correlation between COB and FS and speleothem records from the Indian and East Asian monsoon regions (Figs. 6 and 7). FS also displays a negative correlation with the Cariaco Basin record (Peterson et al., 2000), and both display positive correlations with Ballgown Cave in northern Australia (Denniston et al., 2013) and NGRIP (Rasmussen et al., 2006; Vinther et al., 2006). Conversely, strong positive correlations emerge between CWN and Asian monsoon region caves. CWN

also displays a positive correlation with the Cariaco Basin record and negative correlations with Liang Luar Cave (Ayliffe et al., 2013), Ballgown Cave and NGRIP. These relationships point to strong teleconnections between WNA, the Asian monsoon region, and areas influenced by movement of the Intertropical Convergence Zone (ITCZ). COB and FS are positively correlated with records in the southern region of ITCZ influence (where a southward shift leads to a precipitation increase), while CWN is positively correlated with records in the northern region (where a northward shift leads to a precipitation increase).

The strong correlations between WNA records and records of paleoclimate change from monsoon and ITCZ-influenced regions at zero lag suggest that variations in atmospheric circulation in the subtropics and mid-latitudes occurred synchronously during the last deglaciation, or at least with leads and lags smaller than the age precision typically achieved by U-series dating methods. These changes led to measureable shifts in precipitation $\delta^8\text{O}$ on either side of the Pacific. Synchronous shifts in precipitation in the ITCZ region and in precipitation $\delta^8\text{O}$ in the Asian monsoon region and WNA are consistent with climate models that document the response of atmospheric circulation to cooling in the North Atlantic (Okumura et al., 2009; Chiang et al., 2014; Liu et al., 2014). For example, during the YD, increased meltwater flux led to cooling in the North Atlantic, and increased sea ice extent (Chiang and Bitz, 2005). This altered the interhemispheric thermal gradient, and caused a southward shift of the ITCZ (Vellinga and Wood, 2002; Zhang and Dellworth, 2005; Cheng et al., 2007; Chiang and Friedman, 2012), decreasing precipitation in the region surrounding the Cariaco Basin (Peterson et al., 2000). The southward shift of the ITCZ also led to a southward shift of the Australian-Indonesian monsoon, which caused more negative precipitation $\delta^8\text{O}$ in Indonesia (Ayliffe et al., 2013), and northern Australia (Denniston et al., 2013) by increasing the proportion of rainfall depleted in ^{18}O that is supplied by the monsoon annually. The southward ITCZ shift decreased precipitation over Socotra Island off the coast of Yemen (Shakun et al., 2007), weakened the monsoon over southeast Asia (Sinha et al., 2005), and

weakened southerly monsoon winds over China, reducing rainfall in northern China, and leading to precipitation with less negative $\delta^{18}\text{O}$ over the cave sites of southern China (Liu et al., 2014).

The deglacial changes in precipitation $\delta^{18}\text{O}$ in WNA that occurred synchronously with shifts in precipitation surrounding the ITCZ and Asian monsoon circulation are consistent with the scenario proposed by Chiang et al., (2014) in which the southward shift of the ITCZ in response to Northern Hemisphere cooling leads to a strengthening of the northern Hadley cell (Lindzen and Hou, 1988) and subsequently the winter northern subtropical jet. The stronger winter jet would cause increased precipitation in WNA, conditions suggested by multi-proxy (non- $\delta^{18}\text{O}$) WNA speleothem records for the YD. However, this scenario is also consistent with the changes noted by WNA $\delta^{18}\text{O}$ records, which indicate increased winter storms characterized by mid-latitude Pacific moisture. Decreased temperatures during the YD would also cause enhanced isotopic rainout as storms moved inland, as is suggested by the LV record (Lachniet et al., 2014). The positive correlations between COB and FS and the NGRIP ice core record (Fig. 6) also support these teleconnections between temperature in the high northern latitudes and precipitation in the tropics and mid-latitudes.

The strong correlations between the CWN record and other global records that oppose those observed for COB and FS indicate that different processes govern precipitation $\delta^{18}\text{O}$ at this site. In particular, CWN displays a positive correlation with the Cariaco Basin, a site that experiences increased rainfall when the ITCZ is shifted northward, and negative correlations with Liang Luar and Ballgown Caves that experience decreased rainfall when the ITCZ is shifted northward. These relationships predict increased CWN $\delta^{18}\text{O}$ at times, such as the YD, when Cariaco Basin sediment reflectance is also increased, which corresponds with periods of decreased rainfall in the surrounding region and vice versa (Peterson et al., 2000; Peterson and Haug, 2006). CWN receives the majority of its precipitation from the Gulf of Mexico, and the CWN speleothem $\delta^{18}\text{O}$ is interpreted to reflect changes in the isotopic signature of Gulf of

Mexico source waters (Feng et al., 2014). During the last deglaciation, which is the only time period covered by the CWN record, changes in the Gulf of Mexico $\delta^{18}\text{O}$ signature occurred due to variable influx of isotopically depleted meltwater from the Laurentide Ice Sheet via the Mississippi River (Poore et al., 2003; Flower et al., 2004). This caused decreased precipitation $\delta^{18}\text{O}$ during the BO at CWN (Feng et al., 2014). This mechanism of CWN $\delta^{18}\text{O}$ variability is limited to periods of glacial melting. A more direct mechanistic link between movement of the ITCZ and precipitation $\delta^{18}\text{O}$ at CWN is possible for periods not characterized by large influxes of meltwater. A deglacial to Holocene record of foraminiferal $\delta^{18}\text{O}$ from the western Gulf of Mexico suggests that deeper penetration of the Loop Current when the ITCZ is shifted northward increases transport of Caribbean surface waters into the Gulf of Mexico (Poore et al., 2003). This mechanism would bring Caribbean waters characterized by lower $\delta^{18}\text{O}$ due to increased regional precipitation into the Gulf of Mexico during warm periods when the ITCZ is shifted northward, providing a source of lower $\delta^{18}\text{O}$ precipitation for CWN. This mechanism could lead to a correlation between Texas speleothem $\delta^{18}\text{O}$ and proxies of ITCZ movement on millennial timescales. However, a Holocene speleothem $\delta^{18}\text{O}$ record from a different cave in central Texas shows no correspondence with Gulf of Mexico sediment records (Wong et al., 2015). The development of longer-lived speleothem $\delta^{18}\text{O}$ from CWN for the last glacial period would assist in determining if the strong relationships between central Texas precipitation and the ITCZ and Asian monsoon regions persist, as the correlations presented here suggest.

7. Conclusions and future directions

Our analysis demonstrates that speleothem records from WNA document a robust regional response to the events of the last deglaciation, in particular the BO warming and YD cooling events. We have shown that this response is largely consistent across the region, despite differing controls on precipitation $\delta^{18}\text{O}$ across WNA. Comparison with global records further documents strong teleconnections between WNA and the

tropics and mid-latitudes, suggesting that nearly synchronous changes in atmospheric circulation measurably altered precipitation dynamics during the last deglaciation. However, large age uncertainties and noisier proxy variability in some records make evaluation of mechanistic relationships more challenging.

This analysis of WNA speleothem $\delta^8\text{O}$ also highlights important avenues of future research. In particular, isotope-enabled paleoclimate modeling might help to reconcile the various interpretations of speleothem $\delta^8\text{O}$ for WNA with each other and with observations and interpretations of precipitation $\delta^8\text{O}$ on seasonal to interannual timescales (e.g. Berkelhammer et al., 2012; Ersek et al., 2012; Oster et al., 2012; Buening et al., 2012; 2013). Further development of multi-proxy speleothem records (e.g. Oster et al., 2009; 2015; Polyak et al., 2012; Ersek et al., 2013; Steponaitis et al., 2015), including proxies that are potentially sensitive to precipitation amount such as trace elements, C, Sr, and U isotopes, will assist in reconstructing how rainfall varied in this region and how the changes in atmospheric circulation suggested by the $\delta^8\text{O}$ records influenced precipitation amount. Development of such records across HS1 will be especially useful in verifying the regional hydroclimate response to HS1, which was potentially the wettest interval of the last deglaciation in WNA, and further investigating teleconnections between WNA and ITCZ and monsoon regions (Chiang et al., 2014; Putnam 2015). Lastly, the records included in our analysis were of multi-decadal to centennial resolution, which allowed us to investigate multi-centennial to millennial-scale variability across the last deglaciation. However, important controls on precipitation in WNA exist at the interannual to multi-decadal timeframe. Ocean-atmosphere oscillations such as the El Niño Southern Oscillation (ENSO), the Pacific Decadal Oscillation (PDO), and the Atlantic Multi-decadal Oscillation (AMO) have been shown to influence patterns of precipitation change to variable degrees across WNA in the modern and over the last millennium (e.g. Wise 2010; Wise and Dannenberg, 2014). Development of multi-proxy speleothem records from WNA that are of sufficient resolution and with sufficient age control to document variations on multi-annual to multi-decadal

timescales will allow investigation of the stationarity of these patterns through time and under different boundary conditions.

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Figure Captions

Fig. 1: Location map of published WNA speleothem records. Diamonds represent sites included in this analysis: OC = Oregon Caves (Vacca et al., 2005; Ersek et al., 2012); MC = Moaning Cave (Oster et al., 2009; 2015b); ML = McLean’s Cave (Oster et al., 2015b); LV = Leviathan Cave (Lachniet et al., 2014); PC = Pinnacle Cave (Lachniet et al., 2014); COB = Cave of the Bells (Wagner et al., 2010); FS = Fort Stanton Cave (Asmerom et al., 2010; Polyak et al., 2012); PP = Pink Panther Cave (Asmerom et al., 2007); CWN = Cave without a Name (Feng et al., 2014). Circles represent sites of records outside of or

with insufficient overlap with the deglacial period: Minnetonka Cave (Lundeen et al., 2013); Reeds Cave (Serefidin et al., 2004); Goshute Cave (Denniston et al., 2007); Lehman Caves (Steponaitis et al., 2015; Cross et al., 2015); Crystal Cave (McCabe-Glynn et al., 2013); Carlsbad Caverns (Railsback et al., 2015; Brook et al., 2006); Natural Bridge Cavern (Wong et al., 2015).

Fig. 2: Time series plots for all WNA records included in this analysis. Black dots show measured $\delta^{18}\text{O}$ data points for each published record. Red lines show kernel smoothers for each record. Published U-series ages with 2σ uncertainties are shown on bottom of plots. Time intervals for the Younger Dryas (YD), Bölling-Alleröd (BO) and Heinrich Stadial 1 (HS1) are shown by dashed lines for comparison. See text or Table 1 for record acronyms.

Fig. 3: As for Figure 2 but for global records used in this analysis.

Fig. 4: Network plots for WNA records. Blue lines indicate positive correlation and red lines indicate negative correlation between records. Correlation strength increases with width of line and depth of color. **A)** gXCF correlations are shown at zero temporal lag between records with sites arranged geographically. **B)** As in **A** but sites are arranged by number and strength of connections between records. **C** and **D)** As in **A** and **B** but plotting the strongest gXCF correlations that occur when records are lagged within the maximum age uncertainty (see text).

Fig. 5: A) Time-series plots of 1000 COPRA age model realizations of the COB record, each colored line represents a different age model realization, and the FS record on the published age model. Time series have been de-trended following procedure given in text. **B)** Histogram of gXCF values for each COB COPRA realization and the FS record. **C** and **D)**, as in **A** and **B** but for COB COPRA realizations and the MC record on the published age model. In **B)**, **D)**, green lines show the gXCF value for the records on the published age models for lag zero, blue lines and pink lines show the maximum and minimum gXCF values respectively when the records are lagged within age uncertainty. Red lines show 5% and 95% confidence intervals on lag zero correlations based on 2000 simulated autocorrelated but mutually uncorrelated surrogate records. Confidence intervals on lagged correlations are corrected to account for multiple comparisons and are given in Table 3. Results of comparisons for all WNA records for which COPRA realizations were computed are given in the Supplemental Material.

Fig. 6: Maps showing correlations between WNA records and global speleothem records, the NGRIP and WAIS ice core records, and the Cariaco Basin sediment reflectance record. Color and shading of lines as in Fig. 4. All correlations are computed at zero lag. Records for which no significant correlation exists with the given WNA record are shown only by black circles.

Fig. 7: Network plots for all global records at zero lag. All significant correlations are shown. Line color and shading as in Figs. 4 and 6. Records are arranged by number and strength of correlations as in Fig. 4 **B** and **D**.

Table 1: WNA speleothem records covering the last deglaciation used in statistical comparisons.

Site Name	Acronym	State	Latitude	Longitude	Record Start (ka)	Record End (ka)	Citation
Cave of the Bells	COB	AZ	31.75	-110.75	11.484	21.914	Wagner et al., (2010)
Fort Stanton Cave	FS	NM	33.30	-105.30	11.336	21.957	Polyak et al., (2012); Asmerom et al., (2010)
Leviathan Cave	LV	NV	37.79	-115.82	9.050	13.300	Lachniet et al., (2014)
Pinnacle Cave	PC	NV	36.62	-115.49	15.600	20.000	Lachniet et al., (2014)
Pink Panther Cave	PP	NM	32.08	-105.17	9.001	12.087	Asmerom et al., (2007)
Moaning Cave	MC	CA	38.07	-120.47	9.090	15.870	Oster et al., (2009); Oster et al., (2015)
McLean's Cave	ML	CA	38.07	-120.42	10.970	19.391	Oster et al., (2015)
Oregon Caves	OC	OR	42.10	-123.41	9.170	13.320	Vacco et al., (2005)
Cave Without a Name	CWN	TX	29.88	-98.62	9.546	19.264	Feng et al., (2014a)

Table 2: Global records covering the last deglaciation used in statistical comparisons

Site Name	Acronym	Country	Latitude	Longitude	Record Start (ka)	Record End (ka)	Citation
Sofular Cave	Sofular	Turkey	41.42	31.93	9.003	20.093	Fleitmann et al., (2009)
Hulu Cave	Hulu	China	32.50	119.16	10.290	21.964	Wang et al., (2001)
Sanbao Cave	Sanbao	China	31.67	110.43	9.010	19.210	Wang et al., (2008); Dong et al., (2010)
Soreq Cave	Soreq	Israel	31.45	35.03	9.140	21.890	Bar-Matthews et al., (2003)
Timta Cave	Timta	India	29.84	82.03	11.660	15.215	Sinha et al., (2005)
Yamen Cave	Yamen	China	25.48	107.90	9.068	16.139	Yang et al., (2010)
Dongge Cave	Dongge	China	25.28	108.08	9.016	15.810	Dykoski et al., (2005); Yuan et al., (2004)
Moomi Cave	Moomi	Yemen	12.50	54.00	11.086	21.943	Shakun et al., (2007)
Gunung Buda National Park	GB	Borneo	4.03	114.80	9.035	21.963	Partin et al., (2007)
Santiago Cave	Santiago	Ecuador	-3.02	-78.13	9.066	21.920	Mosblech et al., (2012)
El Condor Cave	ELC	Peru	-5.93	-77.30	9.160	21.810	Cheng et al., (2013)
Liang Luar Cave	LL	Indonesia	-8.53	120.43	9.927	21.987	Ayliffe et al., (2013)
Ballgown Cave	BG	Australia	-17.33	124.08	9.016	21.993	Denniston et al., (2013)
Botuvera Cave	Botuvera	Brazil	-27.22	-49.16	9.071	21.912	Cruz et al., (2005)
Nettlebed Cave	Nettlebed	New Zealand	-41.25	172.92	9.034	21.850	Hellstrom et al., (1998)
NGRIP $\delta^{18}O$	NGRIP	Greenland	75.1	-42.3			Vinther et al., (2006); Rasmussen et al., (2006)
Cariaco Basin reflectance	CrC	Venezuela	10.7	-65.2			Peterson et al., (2000)
WAIS $\delta^{18}O$	WAIS	Antarctica	-79.47	-112.085			WAIS Members (2015)

Table 3: Timing of overlap, age uncertainty, and gXCF values for pairwise comparisons of WNA speleothem records.

Record 1	Record 2	Overlap start (ka)	Overlap end (ka)	Overlap	Max age error w/in overlap (kyr)	gXCF*	Lower CI	Upper CI	abs. max gXCF in error*	lag (yrs)	Adj. Lower CI	Adj. Upper CI
COB	CWN	11.48	19.26	7.78	0.29	-0.79	-0.61	0.62	-1	284	-0.87	0.91
COB	FS	11.48	21.91	10.43	0.23	0.61	-0.40	0.40	0.71	209	-0.56	0.56
COB	LV	11.48	13.30	1.82	0.21	0.72	-0.69	0.69	0.73	-157	-0.74	-0.76
COB	MC	11.48	15.87	4.39	0.53	0.17	-0.46	0.44	0.43	-177	-0.69	0.74
COB	ML	11.48	19.39	7.91	0.52	-0.05	-0.21	0.20	-0.40	516	-0.38	0.37
COB	OC	11.48	13.32	1.84	0.29	0.54	-0.43	0.41	0.77	128	-0.54	0.52
COB	PC	15.60	20.00	4.40	0.45	-0.1	-0.45	0.42	-0.19	-146	-0.62	0.65
COB	PP	11.48	12.09	0.60	NA	NA	NA	NA	NA	NA	NA	NA
CWN	MC	9.55	15.87	6.32	0.53	0.623	-0.68	0.69	0.77	474	-1	0.95
CWN	PC	15.60	19.26	3.66	0.44	0.06	-0.95	0.84	0.27	-213	-1	1
CWN	PP	9.55	10.09	0.54	NA	NA	NA	NA	NA	NA	NA	NA
FS	CWN	11.34	19.26	7.93	0.29	-0.63	-0.65	0.66	-0.72	-94	-0.86	0.86
FS	LV	11.34	13.30	1.96	0.11	0.89	-0.51	0.54	0.98	-61	-0.52	0.53
FS	MC	11.34	15.87	4.53	0.53	-0.56	-0.72	0.72	-0.82	-134	-1	1
FS	ML	11.34	19.39	8.05	0.52	-0.28	-0.28	0.27	-0.62	430	-0.43	0.41
FS	OC	11.34	13.32	1.98	0.29	1	-0.34	0.36	1	7	-0.51	0.47
FS	PC	15.60	20.00	4.40	0.45	0.09	-0.47	0.45	0.23	-446	-0.61	0.62
FS	PP	11.34	12.09	0.75	NA	NA	NA	NA	NA	NA	NA	NA
LV	CWN	9.55	13.30	3.75	0.40	-1	-0.69	0.66	-1	75	-0.97	1
LV	MC	9.09	13.30	4.21	0.45	0.15	-0.40	0.37	0.25	235	-0.56	0.56
LV	ML	10.97	13.30	2.33	0.52	-0.03	-0.40	0.41	-0.77	380	-0.68	0.65
LV	OC	9.17	13.30	4.13	0.29	0.45	-0.29	0.30	0.53	-15	-0.38	0.37
LV	PP	9.05	12.09	3.04	1.06	0.04	-0.29	0.28	0.41	235	-0.44	0.48
MC	PP	9.09	12.09	3.00	1.06	-0.06	-0.31	0.29	-0.43	766	-0.59	0.56
ML	CWN	10.97	19.26	8.29	0.52	0.44	-0.25	0.25	0.65	-308	-0.57	0.47
ML	MC	10.97	15.87	4.90	0.53	0.37	-0.37	0.36	0.73	64	-0.56	0.54
ML	OC	10.97	13.32	2.35	0.52	-0.15	0.45	0.43	-0.59	216	-0.65	0.65
ML	PC	15.60	19.39	3.79	0.44	0.12	-0.24	0.25	0.4	63	-0.32	0.32
ML	PP	10.97	12.09	1.12	1.06	-0.32	-0.29	0.27	-0.65	953	-0.56	0.52
OC	CWN	9.55	13.32	3.77	0.40	-0.92	-0.49	0.47	-1	217	-0.73	0.73
OC	MC	9.17	13.32	4.15	0.45	0.04	-0.32	0.34	0.38	400	-0.56	0.56
OC	PP	9.17	12.09	2.92	1.06	-0.14	-0.30	0.29	0.36	-430	-0.47	0.5

*bolded values are significant outside the 95% confidence intervals

Figure 1

A gXCF detrended

C gXCF detrended in error

B gXCF detrended

D gXCF detrended in error

Figure 5

Figure 6

Figure 7