1 Mid-Holocene drying of the U.S. Great Basin recorded in Nevada speleothems

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16 **Keywords:** Great Basin; Holocene; speleothems; U-Th dating; paleoclimate

17 Abstract

18 Lake level records point to dramatic changes in Great Basin water balance over
19 the last 25 ka, but the timing and pace of Holocene drying in the region remains poorly
20 documented. Here we present stable isotope and trace metal data from two Lehman
21 Caves, NV speleothems that provide a well-dated record of latest Pleistocene to mid-
22 Holocene hydroclimate in the U.S. Great Basin. Together the stalagmites span the
23 interval between 16.4 ka and 3.8 ka, with a hiatus from 15.0 ka to 12.7 ka. Mg/Ca and
24 $\delta^{13}\text{C}$ covary throughout the records, consistent with control by the extent of degassing
25 and prior calcite precipitation (PCP); measurements of modern cave and soil waters
26 support PCP as the primary control on drip-water trace-element composition. We
27 therefore interpret Mg/Ca and $\delta^{13}\text{C}$ as reflecting infiltration rates, with higher values
28 corresponding to drier periods. Both Mg/Ca and $\delta^{13}\text{C}$ indicate a wet period at the
29 beginning of the record (12.7-8.2 ka) followed by pronounced drying after 8.2 ka. This
30 mid-Holocene drying is consistent with records from around the western United States,

including a new compilation of Great Basin lake-level records. The strong temporal correspondence with the collapse of the Laurentide ice sheet over Hudson Bay suggests that this drying may have been triggered by northward movement of the winter storm track as a result of ice sheet retreat. However, we cannot rule out an alternative hypothesis that wet early Holocene conditions are related to equatorial Pacific sea-surface temperature. Regardless, our results suggest that Great Basin water balance in the early Holocene was driven by factors other than orbital changes.

1. Introduction

The Great Basin is a large internally drained region in the western United States that covers large parts of Nevada, Utah, California, and Oregon (Figure 1). Modern climate over much of the Great Basin is arid, with most of its sub-basins unable to sustain permanent lakes; however, the spectacular paleoshorelines and lake deposits in the Great Basin have long been recognized as evidence of dramatic hydrologic changes in the past. As such, the Great Basin has fascinated geologists since the late 19th century, when G. K. Gilbert and I.C. Russell began to unravel the histories of the region's massive paleo-lakes. More recently, developments in the application of radiocarbon, and later, U-series, dating methods have yielded improved chronologies of hydrologic change from the Great Basin.

Despite years of research, well-dated Holocene records of hydrological change from the Great Basin remain sparse. Most existing records of past Great Basin hydrology utilize either shoreline and sediment deposits from closed-basin lakes or biological archives like packrat middens. Although they offer valuable information, these types of records do not always provide the temporal resolution necessary to make inferences about

mechanisms of climate change. In addition, lake deposits commonly record wetter conditions during the last glacial period but offer incomplete records of drier conditions during the Holocene.

In recent decades, high-precision U-Th dating of speleothems, combined with trace element and stable isotope measurements, has allowed for the development of detailed chronologies of climate change. To date, there are few published speleothem records from in and around the Great Basin (Asmerom et al., 2007; Denniston et al., 2007; Lachniet et al., 2014; Lundeen et al., 2013; Oster et al., 2009; Polyak et al., 2004; Shakun et al., 2011; Wagner et al., 2010), and only a small number of these offer substantial coverage of the Holocene. Well-dated terrestrial records from this region are necessary to better understand the response of Great Basin hydroclimate to changing boundary conditions over the late Quaternary and to assess the representation of regional precipitation patterns in general circulation models simulating past climates.

This study presents geochemical (Mg/Ca and Sr/Ca) and stable isotope ($\delta^{18}\text{O}$, $\delta^{13}\text{C}$) records spanning much of the deglaciation and Holocene from two speleothems from Lehman Caves, Nevada. We interpret these data – and in particular, the Mg/Ca and $\delta^{13}\text{C}$ records – as primarily reflecting infiltration rates above the cave, and we present evidence that local infiltration rates are well correlated with water balance changes over a large portion of the Great Basin during the early to mid Holocene. These records provide important constraints on the potential drivers of relatively wet early Holocene conditions and of mid-Holocene drying in the Great Basin.

2. Regional Setting

2.1 *Lehman Caves*

Lehman Caves is situated on the east flank of the southern Snake Range on the western margin of the Bonneville Basin, at 39°00'20"N, 114°13'13"W and 2130 m elevation (Figure 1). Average annual precipitation above the cave is approximately 33 cm/year (National Park Service). Seasonal recharge in Lehman Caves is dominated by winter precipitation, as evidenced by seasonal changes in drip rates and cave pool levels; dripwater response time is 1-4 weeks (Ben Roberts, National Park Service, personal communication). The cave is situated within a local topographic high in the Pole Canyon limestone such that the great majority of water entering the cave is from infiltration directly above the cave, not from infiltration or run-off from the higher elevations of the Snake Range. Most of the cave network is situated between 30 and 60 m from the surface (National Park Service). HOBO data loggers (Onset Computer Corporation, Bourne, MA) placed in the cave in 2009-2010 indicate that air temperature and relative humidity in the cave remain approximately constant year round, at 11.0°C and approximately 100%, respectively.

The Bonneville Basin enclosed a very large (~55,000 km²) lake during the Last Glacial Maximum (LGM) and early deglaciation that lay just to the east of the cave site (Figure 1), reflecting significantly more positive water balance in the region at these times. The rise of Lake Bonneville leading into the LGM has been suggested by a number of studies to reflect the southward displacement of the mean winter storm track by the Laurentide and Cordilleran ice sheets (Antevs, 1952; Bromwich et al., 2004; COHMAP Members, 1988), although Lyle et al. (2012) used coastal precipitation records to suggest that post-LGM precipitation entered the Great Basin from the tropical Pacific.

Superimposed on this response to ice sheet topography, the basin experienced its wettest conditions during ice-rafting events in the North Atlantic, in particular Heinrich events 1 and 2 (McGee et al. 2012; Munroe and Laabs 2013; Oviatt 1997). Although the Bonneville Basin is well studied, relatively little is known about the precise timing of hydrological changes in the Great Basin during the latest Pleistocene and early-to-mid-Holocene. Lake levels dropped considerably around 15 ka, approximately at the time of the Bølling/Allerød warming in the Northern Hemisphere (Godsey et al., 2011; McGee et al., 2012; Oviatt et al., 1992). The work of Murchison (1989) and Oviatt et al. (2005) on lacustrine deposits indicates a modest rise of the lake known as the Gilbert highstand between ~12.9 and 11.2 ka, a time which is roughly correlative with the Younger Dryas cold event in the Northern Hemisphere (12.9-11.7 ka; Rasmussen et al. 2006). Other studies from the Bonneville Basin, reviewed in section 5.4 below, document the drying of the basin during the early- to mid-Holocene, but the timing and drivers of this drying remain unclear (Madsen et al., 2001; Patrickson et al., 2010).

3. Materials and Methods

3.1 Sample collection

Two Lehman Cave stalagmites, WR11 and CDR3, were analyzed for this study (Figure 2). WR11 was collected from the West Room of the cave, located approximately 50 m below the surface (National Park Service); it originally precipitated on a piece of flowstone that had been broken during cave development over the past century. CDR3 had been broken during previous cave vandalism and was collected from the part of the

cave known as the Civil Defense Room (Figure 2) that is used for storage of broken stalagmites; the original growth location is thus unknown. The stalagmites were cut, polished and rinsed in deionized water before being sampled.

In May 2013 and January 2014, small (10 to 30 mL) acid-cleaned HDPE bottles were used to collect drips from soda straws from locations throughout the cave. Water was also collected from standing pools on the cave floor. In addition, soil was sampled over 10-cm intervals at depths ranging from 0-50 cm from three pits dug above the cave in May 2013.

3.2 U-Th dating of speleothems

Dating samples weighing 20-100 mg were drilled from the speleothems using a vertical mill. The resulting powders were weighed, spiked with ^{229}Th - ^{233}U - ^{236}U tracer, and dissolved. Following the methods of Edwards et al. (1987), U and Th were removed from solution by co-precipitation with Fe oxyhydroxides, redissolved, and eluted separately through 0.5 mL bed volume columns packed with BioRad AG1-X8 resin. A total procedural blank was included in each set of chemistry (5-10 samples).

U and Th fractions prepared at MIT were measured either on Brown University's Thermo Scientific Neptune Plus multicollector ICP-MS or on the Thermo Scientific Neptune at Woods Hole Oceanographic Institution (WHOI). Samples prepared at the University of Minnesota were analyzed on the Thermo Scientific Neptune at the University of Minnesota. In all locations, samples were introduced using a CETAC Aridus II desolvating nebulizer intake system and a 100 $\mu\text{L}/\text{min}$ PFA nebulizer. For

analyses at Brown University and WHOI, ^{234}U and ^{230}Th were measured on the secondary electron multiplier (SEM) and all other isotopes (^{233}U , ^{235}U , ^{236}U , ^{238}U , ^{229}Th , and ^{232}Th) were measured on the Faraday cups. The retarding potential quadrupole (RPQ) energy filter was not used on the SEM, as we found it reduced the stability of the Faraday-SEM relative yield. Each U sample was bracketed with a 5 ng/g solution of the CRM112a U isotopic standard to monitor SEM yield. Tailing for both samples and standards was estimated using measurement of half-masses and mass 237 immediately following each on-peak measurement. Each Th analysis was bracketed by an in-house ^{229}Th - ^{230}Th - ^{232}Th standard used to monitor mass bias and SEM yield; this standard was calibrated by bracketing with IRMM 3636a ^{233}U - ^{236}U solution. 2% HNO_3 solution blanks were run bracketing each sample and standard to determine background signal. At the University of Minnesota, samples were analyzed using a peak-jumping routine on the axial SEM following the methods of Shen et al. (2002).

Total procedural blanks for various sample sets were less than 0.07 fg ^{230}Th , 3 pg ^{232}Th , 0.03 fg ^{234}U , and 10 pg ^{238}U . After making corrections for background, blank, mass bias, tailing, and SEM yield, U-Th ages were calculated using the decay constants determined by Jaffey et al. (1971) for ^{238}U and Cheng et al. (2013) for ^{234}U and ^{230}Th . Corrections for initial ^{230}Th assumed an initial $^{230}\text{Th}/^{232}\text{Th}$ atomic ratio of $4.4 \pm 2.2 \times 10^{-6}$ (Yardley, 1986). Uncertainties from measurements, spike calibration, procedural blanks, SEM yield drift and tailing were propagated to determine the uncertainties reported in Supplementary Table 1. Reported ages do not include uncertainties on U and Th half-lives.

3.3 Speleothem major and trace element measurements

Major and trace elements were measured by laser ablation ICPMS on a single collector Thermo Element2 ICPMS at Woods Hole Oceanographic Institution (WHOI) using a New Wave Research UP 193 nm excimer laser system. Major and trace elements were measured along the primary growth axis, using previously drilled stable isotope holes for spatial reference. Analyses were taken at 1 mm spacing (i.e. next to every second stable isotope hole) using a spot size of 100 microns and an integration time of 60 seconds. The first 20 seconds of data were discarded.

Internal calibration was established by normalization to ^{48}Ca , assuming constant calcium content in the sample. Final results reflect corrections for blank intensities and machine drift monitored by external calibration to a solid carbonate standard (USGS MACS-3) with well-characterized compositions (Jochum et al., 2012). Reproducibility was checked through sampling of 23 previously analyzed points along the same horizontal growth plane. Of those measurements, 18 agreed within 5% and the remaining 5 agreed within 15%.

3.4 Stable isotope analyses

In WR11, carbonate powders were drilled in a vertical milling machine at 0.5 mm spacing using a digital tachometer readout to ensure regular sample spacing. In CDR3, powders were hand drilled at 0.5 mm spacing. The powders were dissolved in dehydrated phosphoric acid at 70 °C in a KIEL-III automated carbonate preparation device and analyzed with a Finnigan MAT 252 gas ratio mass spectrometer at the University of Arizona. 2σ uncertainty is approximately $\pm 0.22\text{‰}$ for $\delta^{18}\text{O}$ values and $\pm 0.16\text{‰}$ for $\delta^{13}\text{C}$ values.

3.5 Cave water and soil analyses

Cave waters were diluted by a factor of 500 with 0.5M ultra-clean nitric acid, spiked with Sc and In at concentrations of 1 ng/g to monitor yield, and filtered through 0.45 μm PTFE syringe filters to remove any solids. In order to assess dissolved element ratios in soil pore waters, dry soil samples were rinsed following the procedure of Oster et al. (2012). Approximately 30 g of dry soil were combined with 30 mL of 18.2 M Ω de-ionized water in centrifuge tubes for 24 hours. The mixture was then centrifuged, and approximately 1 mL of this water was diluted by a factor of 100 with 0.5M ultra-clean nitric acid, then spiked with Sc and In and filtered in the same manner as the cave waters.

All waters were analyzed for Mg, Sc, Ca, Sr, and In on a VG Elemental PlasmaQuad 2+ quadrupole ICP-MS at MIT. Bracketing standards with 1:100 ratios of all other elements to Ca were run after every five samples to monitor the relative yield of each element. Uncertainties were estimated by repeat measurements of samples and measurements of multiple Mg and Sr isotopes; analytical uncertainties for ratios are <2%, and reproducibility averaged better than 5%. Procedural blanks were prepared with each set of waters and were negligible. Oxygen and hydrogen stable isotope ratios of waters were determined on a Picarro L2130-i Analyzer. Results are reported relative to the VSMOW standard and have a reproducibility of better than 0.1‰ for $\delta^{18}\text{O}$ and 0.4‰ for δD .

3.6 Lake-level compilation

A lake-level database was compiled using published data from 23 locations around the Great Basin. In this database, low lake level was defined as any lake level

more than 0.5 standard deviations below mean lake level. Lake locations and references are included in Supplementary Table 6.

4. Results

4.1 Sample description

WR11 is a translucent speleothem marked by light-colored calcite deposited on a dark brown flowstone, with only light banding visible to the naked eye in some parts of the speleothem (Figure 2). The base of the lighter part of WR11 is morphologically a flowstone and is not included in this study. Flowstone transitions into a speleothem morphology 10 mm from the base of the light-colored deposit. As banding in the lower part of this speleothem is not visible to the naked eye or in an ordinary light microscope, the morphology of WR11 was determined using a Zeiss 710 confocal microscope at the Whitehead Institute at MIT.

CDR3 is a fragment of a larger speleothem. It is more opaque and lighter in color than WR11, with distinct banding visible to the naked eye and in confocal imagery. A depositional hiatus is marked by a transition from opaque white calcite to more translucent, yellow calcite 11 mm above the base (Figure 2).

4.2 U-Th dating of speleothems

We obtained twelve ages within WR11 and nine within CDR3 (Supplementary Table 1), all of which were in stratigraphic order. In WR11, U concentrations were between 40 and 300 ng/g, and in CDR3, U concentrations were between 160 and 300 ng/g. Several ages in each stalagmite were replicated by resampling the stalagmite to

determine the reproducibility of MIT protocols and to test for offsets between UMN and MIT (Supplementary Figure 1); replicates are denoted in Supplementary Table 1 by a lowercase letter following the sample number (i.e. WR11-7a). Replicates showed only small offsets (40-200 years) and did not indicate systematic differences between UMN and MIT.

Growth rates vary substantially in both CDR3 and WR11 (Figure 3). In CDR3, between 16 ka and 15 ka, growth rate is approximately 8 mm/kyr. Growth ceases at 15.0 ka and resumes around 12.6 ka, after which time the speleothem grows at about 30 mm/kyr until the record ends at 10.2 ka. The onset of stalagmite (as opposed to flowstone) deposition in WR11 occurs at 11.5 ka. Between 11.5 and 10.4 ka, the growth rate of WR11 is 20 mm/kyr, comparable with the high 30 mm/kyr growth rate in CDR3. After 10.4 ka, the growth rate of WR11 decreases to about 7 mm/kyr.

4.3 Elemental and stable isotope composition of speleothems

In WR11, $\delta^{18}\text{O}$ values range between -12.6 and -10.0 ‰ and $\delta^{13}\text{C}$ values range between -5.6 and -0.9 ‰ (Figure 4, Supplementary Table 2); Mg/Ca ratios range from 2.1 to 6.1 mmol/mol and Sr/Ca ratios range from 0.087 to 0.16 mmol/mol (Figure 4, Supplementary Table 3). In CDR3, $\delta^{18}\text{O}$ values range between -13.2 and -10.0 ‰ and $\delta^{13}\text{C}$ values range between -7.2 and -2.7 ‰ (Figure 4, Supplementary Table 2); Mg/Ca ratios range from 1.3 to 2.6 mmol/mol and Sr/Ca ratios range from 0.083 to 0.14 mmol/mol (Figure 4, Supplementary Table 3).

4.4 Elemental and stable isotope composition of cave and soil water

Mg/Ca of drip and pool waters ranges between approximately 0.06 and 2.0 mol/mol; Sr/Ca is between approximately 0.40 and 6.3 mmol/mol (Supplementary Table 4). Mg/Ca and Sr/Ca ratios in cave waters covary (Figure 5). Soil washes return ratios similar to the lowest ratios measured in cave waters, with Mg/Ca ranging from 0.053 to 0.14 mol/mol and Sr/Ca from 0.72 to 2.1 mmol/mol. Ratios generally increase with soil depth (Supplementary Table 4). Cave water $\delta^{18}\text{O}$ values range from -13.2 to -10.0 ‰; δD values range from -101.8 to -82.4 ‰ (Supplementary Table 4). $\delta^{18}\text{O}$ values of cave drip and pool waters sampled in May 2013 average -11.9 ± 0.1 ‰ VSMOW (1 standard error of the mean; n=18) after exclusion of one sample taken closest to the natural entrance with a $\delta^{18}\text{O}$ value of -10.0 ‰. This average value is similar to the isotopic composition of winter precipitation above the caves (Bryan Hamilton, National Park Service, unpublished data).

5. Discussion

5.1 Interpretation of elemental records

High pCO_2 in soil waters favors the dissolution of carbonate minerals by fluids moving down into the epikarst, via the following reaction:

In the cave, low pCO_2 conditions reverse this chemical reaction; CO_2 is released and the fluid begins to precipitate calcite. The behavior of trace elements such as Mg and Sr in these reactions is described by the distribution coefficients D_{Mg} and D_{Sr} , which are defined as:

274
$$D_x = (X/Ca)_{\text{calcite}} / (X/Ca)_{\text{fluid}}$$

275 where X= Mg, Sr. D_{Mg} and D_{Sr} are both <1 (Day and Henderson, 2013), which means,
276 assuming closed-system behavior, that host fluids will become enriched in Mg and Sr as
277 calcite precipitation proceeds. Where mixing is negligible, this enrichment follows simple
278 Rayleigh distillation.

279 In drier conditions, longer fluid residence times in the epikarst and slower drip
280 rates from stalactites allow for substantial degassing of CO₂ and calcite precipitation
281 before waters reach stalagmites, increasing Mg/Ca and Sr/Ca ratios in fluids from which
282 speleothems eventually precipitate “downstream” of the epikarst. Conversely, in wetter
283 conditions with higher epikarst recharge rates, less calcite is precipitated in the epikarst,
284 producing fluids with correspondingly lower Mg/Ca and Sr/Ca ratios. This process is
285 known as prior calcite precipitation (PCP) and is widely viewed as a dominant control on
286 trace element ratios in cave carbonates (Baker et al., 1997; Day and Henderson, 2013;
287 Fairchild et al., 2006; Johnson et al., 2006; Oster et al., 2009; Sinclair et al., 2012;
288 Tremaine and Froelich, 2012).

289 Mg/Ca and Sr/Ca ratios of shallow soil water above Lehman Cave are similar to
290 the lowest values measured in cave dripwaters, supporting the idea that infiltrating waters
291 start with low ratios and gradually increase as a result of PCP. Consistent with this
292 hypothesis, Rayleigh fractionation curves calculated using distribution coefficients of
293 D_{Mg} (0.0125) and D_{Sr} (0.125) calculated from Day and Henderson (2013) and measured
294 modern cave temperature of 11°C closely match the variation of Mg/Ca, Sr/Ca and Ca
295 concentrations observed in modern cave waters (Figure 5).

The wide range of Mg/Ca and Sr/Ca ratios observed in modern drip waters suggests that they undergo differing degrees of PCP in the epikarst. Other possible controls on Mg/Ca, such as changing soil geochemistry and mixing between different epikarst reservoirs (Fairchild et al., 2006), probably do not play a large role in Lehman Caves due to the thin soil above the cave and uniform bedrock. Because Mg/Ca and Sr/Ca of dripwaters are different under the same conditions, it is reasonable to expect different elemental ratios between coeval stalagmites; however, we would expect changes to be of the same sign and timing in different stalagmites if epikarst flow is responding to changes in infiltration rates due to regional climate changes. Because of the shape of the fractionation curve, the amplitude of trace element variations will increase at higher mean values of PCP.

Tremaine and Froelich (2013) find that Sr/Ca and Mg/Ca ratios covary in speleothem calcite precipitated on plates from cave dripwaters and attribute this covariation to PCP; Sinclair et al. (2012) present model and empirical data suggesting that PCP results in covariation of Sr/Ca and Mg/Ca. The slope of this covariation depends on the initial Sr and Mg contents of the dripwater. In this study, Sr/Ca and Mg/Ca are well correlated after ~10 ka, suggesting a strong control by PCP after this time, but they are not correlated prior to 10 ka. Other studies of farmed cave calcite and speleothem calcite find that Sr/Ca does not always covary with Mg/Ca, even when Mg/Ca is likely to reflect PCP (e.g., Huang and Fairchild, 2001; Orland et al., 2014). In several cave systems, excess Sr supplied from windblown dust is a likely cause of decoupling between Sr and Mg (Frumkin and Stein, 2004; Li et al., 2005; Orland et al., 2014). We speculate that Sr/Ca ratios between 10-16 ka could have been impacted by delivery of aragonite-

rich dust to soils above the cave from newly exposed playa surfaces after the fall of Lake Bonneville and other Great Basin lakes. The greater susceptibility of Sr to control by processes other than PCP is also due to its order-of-magnitude higher partition coefficient (e.g., Day and Henderson, 2013), which causes its enrichment in cave waters during PCP to be substantially less than that for Mg. We therefore focus on the Mg/Ca record as an indicator of PCP and thus infiltration rates, but we note that strong covariation with Sr/Ca exists during the time period from 10-4 ka, which constitutes the main focus of this study.

5.2 Interpretation of $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values in Lehman Caves stalagmites

A variety of factors influence $\delta^{13}\text{C}$ values of speleothem calcite, including temperature (Mühlinghaus et al. 2007, 2009), ratio of C_3 to C_4 plants on the surface above the cave (Fairchild et al., 2006), and degassing in the epikarst (Bar-Matthews et al. 1996; Fairchild et al. 2006) and cave itself (Fairchild and Baker, 2012). Temperature has only a weak effect on $\delta^{13}\text{C}$, impacting the fractionation factor between $\text{HCO}_3^-_{(\text{aq})}$ and CaCO_3 by $<0.01\text{‰}$ per degree C. Pollen and midden records from proximal areas suggest that the ratio of C_3 to C_4 plants is unlikely to have been a major control on this record, as C_3 plants have been dominant throughout the late Pleistocene and Holocene at sites with elevations similar to the cave (Madsen et al., 2001; Rhode and Madsen, 1995). Vegetation density and respiration rates can also affect $\delta^{13}\text{C}$ of speleothem calcite in the underlying cave (Baldini et al. 2005). Baldini et al. (2005) observed changes in $\delta^{13}\text{C}$ of speleothem calcite up to of up to 2‰ resulting from dramatic changes in vegetation cover above a cave. The amplitude of $\delta^{13}\text{C}$ variability in WR11 is around 6‰, and the

vegetation changes above Lehman Caves are unlikely to have been as dramatic as those described by Baldini et al. Thus, vegetation density is probably not a dominant control on $\delta^{13}\text{C}$ of Lehman Caves speleothems. The correlation between Mg/Ca and $\delta^{13}\text{C}$ records in CDR3 and WR11 (Figure 6) points to degassing of CO_2 in the epikarst as the primary control on speleothem $\delta^{13}\text{C}$ values (Fairchild and Baker, 2012; Johnson et al., 2006; Oster et al., 2009). Preferential degassing of $^{12}\text{CO}_2$ enriches residual dissolved inorganic carbon (DIC) in ^{13}C , increasing $\delta^{13}\text{C}$ values in speleothem calcite. This same degassing also drives PCP, enriching Mg/Ca and Sr/Ca ratios in waters and leading to $\delta^{13}\text{C}$ and Mg/Ca covariation.

Controls on $\delta^{18}\text{O}$ values of speleothem calcite are complex; they include condensation temperature, precipitation seasonality, precipitation source, precipitation amount, evaporation in the soil and epikarst, cave temperature, and kinetic fractionation during calcite precipitation (Gascoyne, 1992; Goede et al., 1982; Fairchild et al., 2006; Harmon et al., 1978; Hendy and Wilson, 1968; Lachniet 2009; Mickler et al., 2006; Schwarcz et al., 1976; Yonge et al., 1985). $\delta^{18}\text{O}$ and δD measurements of modern drip and pool waters from the interior of the cave (this study) fall on the same local meteoric water line as $\delta^{18}\text{O}$ and δD measurements of precipitation and stream waters near to the cave (Bryan Hamilton, personal communication); this observation, along with the high relative humidity measured in the modern cave, provides at least coarse support that cave waters are not highly evaporated. During the period of overlap between WR11 and CDR3, differences in $\delta^{18}\text{O}$ values between speleothems (Figure 4) are of similar magnitude to the amplitude of variation in the full length of the $\delta^{18}\text{O}$ records, suggesting that $\delta^{18}\text{O}$ in at least one of these stalagmites may not reflect precipitation $\delta^{18}\text{O}$. The strong covariation

between $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ in WR11 (Figure 6) suggests that $\delta^{18}\text{O}$ in this sample may be impacted by kinetic effects. One mechanism of kinetic fractionation that drives positive $\delta^{18}\text{O}$ - $\delta^{13}\text{C}$ covariation is enrichment of the HCO_3^- (aq) pool in ^{18}O by CO_2 degassing, followed by calcite precipitation proceeding faster than isotopic exchange between HCO_3^- (aq) and H_2O (Hendy, 1971; Mickler et al., 2006, 2004). This mechanism, which is argued to be common in semi-arid caves (Mickler et al., 2006), ties $\delta^{18}\text{O}$ variations to the same degassing and PCP that drive $\delta^{13}\text{C}$ and Mg/Ca variations, offering an explanation for the $\delta^{18}\text{O}$ - $\delta^{13}\text{C}$ -Mg/Ca correlations we observe.

In addition to showing poor replication with sample CDR3, the $\delta^{18}\text{O}$ record from WR11 also does not replicate a previously published stalagmite $\delta^{18}\text{O}$ record from Leviathan Cave in Nevada (Figure 7) (Lachniet et al., 2014). The shorter record from CDR3, which does not show significant covariability between $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ (Supplementary Figure 2), shows reasonable agreement with the overlapping portion of the Leviathan Cave record, but during the mid-Holocene Leviathan Cave $\delta^{18}\text{O}$ values are substantially more positive than values in Lehman Cave sample WR11. This poor reproducibility, along with other studies of $\delta^{18}\text{O}$ in semi-arid regions (Kanner et al., 2014; Mickler et al., 2006), urges caution in linking $\delta^{18}\text{O}$ records from single stalagmites directly to $\delta^{18}\text{O}$ values of precipitation in the Great Basin.

Based on the poor reproducibility of $\delta^{18}\text{O}$ records both within and between caves we do not interpret our $\delta^{18}\text{O}$ records as reflecting precipitation $\delta^{18}\text{O}$. Instead, we take the covariation of $\delta^{18}\text{O}$, $\delta^{13}\text{C}$ and Mg/Ca in much of the Lehman Cave record, and in

particular after 10 ka, as an indication that $\delta^{18}\text{O}$ in these samples primarily reflects kinetic fractionation related to degassing and PCP (Mickler et al., 2004; 2006).

5.3 Regional hydroclimate changes from 16.4-3.8 ka inferred from these results

As detailed in sections 5.1 and 5.2 above, we interpret our trace element and stable isotope records as primarily reflecting infiltration rates above Lehman Caves, which we assume to be related to winter precipitation amount. Reduced infiltration rates allow increased degassing and prior calcite precipitation in the epikarst and in stalactites, resulting in covarying and elevated Mg/Ca, Sr/Ca, $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ in speleothem calcite. We suggest that Mg/Ca and $\delta^{13}\text{C}$ are most simply related to infiltration rates due to their high sensitivity to degassing and PCP. Sr/Ca (due to its higher distribution coefficient in calcite) and $\delta^{18}\text{O}$ (due to partial buffering by isotopic exchange with oxygen in water) are less strongly impacted and are more likely to reflect other environmental factors, such as dust deposition for Sr and cave temperature and precipitation $\delta^{18}\text{O}$ for $\delta^{18}\text{O}$.

Based on this reasoning, low Mg/Ca and $\delta^{13}\text{C}$ values in CDR3 indicate wet conditions at the end of Heinrich Stadial 1 between 16.4 and 15.0 ka, consistent with widespread evidence of wet conditions during this period in the Great Basin (McGee et al., 2012; Munroe and Laabs 2013; Oviatt, 1997). Drying over the interval is suggested by rising $\delta^{13}\text{C}$ values (Figure 8). Mg/Ca values do not rise, perhaps due to the start of PCP being marked by little enrichment in dripwater Mg/Ca (Figure 5) or other previously discussed controls on $\delta^{13}\text{C}$. This drying would be consistent with steadily increasing $\delta^{18}\text{O}$ values of Lake Bonneville carbonates over this time period (Benson et al. 2011; McGee

et al. 2012) as well as with the timing of drying in Jakes Lake and Lake Franklin, located approximately 100 km west of the caves (García and Stokes, 2006; Munroe and Laabs, 2013a). The hiatus from 15.0 to 12.7 ka in CDR3 may be suggestive of locally dry conditions during the Bølling-Allerød warm period (14.7-12.9 ka), consistent with Oster et al. (2009) and records from Lake Bonneville (Oviatt et al. 1992; Godsey et al. 2011).

The record resumes at 12.7 ka in CDR3, showing similarly wet conditions to the first interval of deposition in both Mg/Ca and $\delta^{13}\text{C}$. The speleothem record from WR11 begins at 11.5 ka and the CDR3 record ends at 10.2 ka. The period of overlap is characterized by similar patterns in $\delta^{13}\text{C}$, Mg/Ca and Sr/Ca between the two speleothems. After 10.3 ka, the Mg/Ca and $\delta^{13}\text{C}$ values in WR11 increase slightly, suggesting slightly drier conditions than during the YD and earliest Holocene. After 8.2 ka, Mg/Ca and $\delta^{13}\text{C}$ values increase markedly from 8.2 ka until 6.3 ka, suggesting a sharp reduction in infiltration rates above the cave. The data suggest that locally dry conditions continue until the end of the record at 3.8 ka.

5.4 Mid-Holocene drying: comparison with regional records

The record of hydroclimate changes from Lehman Cave speleothems points to wet conditions beginning at the onset of the Younger Dryas but then persisting well after its end. Inferred highest infiltration rates persist until 10.3 ka, followed by moderately drier conditions from 10.3-8.2 ka and pronounced drying beginning at 8.2 ka. The record from WR11 is the only existing speleothem record from the Great Basin that captures and directly dates the onset of drying after this prolonged wet period. However, many other non-speleothem records from the Bonneville Basin and the Great Basin show the same

pattern of an early Holocene wet period followed by a mid-Holocene dry period. Indeed, as early as 1952, Antevs proposed an early-Holocene “anathermal” characterized by slow drying followed by an extremely dry mid-Holocene “altithermal” during which the Great Basin was substantially drier than present day (Antevs, 1952).

In reviewing these records we note that different types of records may have varying sensitivities to temperature and precipitation due to differences in elevation and proxy type. For instance, low-elevation paleoecological records may be more sensitive to precipitation than to temperature than high-elevation paleoecological records (Power et al., 2011). We review these diverse records in order to provide a broader view of the timing of regional changes in the Great Basin.

Holocene climate changes in the Bonneville Basin have been broadly constrained by the work of many authors (Broughton et al., 2000, 2008; Currey, 1990; Hart et al., 2004; Murchison, 1989; Oviatt, 2005). Radiocarbon dates and sedimentological work on Holocene Lake Bonneville presented by these authors suggest that the lake rose to the Gilbert shoreline sometime after 13 ka. In Homestead Cave, located on the west side of the Great Salt Lake, Madsen et al. (2001) observe a marked disappearance of small mammal species in middens after around 9.1 ka, potentially associated with drying. About 180 km north of Lehman Caves, also on the west side of the Bonneville Basin, a pollen record from Blue Lake marsh shows pronounced desiccation after 8.3 ka (Louderback and Rhode, 2009).

Many recent paleoclimate studies conducted in and around the Great Basin suggest the same general structure of Holocene hydroclimate: wet conditions persisting

during and for a few thousand years after the YD followed by pronounced drying. On the western side of the Great Basin, elemental composition and stable isotope records from a speleothem from Moaning Cave, located in the Sierra Nevada, suggests that wet conditions in the region began near the onset of the YD and persisted until least 10.6 ka and possibly as late as 9.6 ka (Oster et al., 2009). Pollen records from Tulare Lake, located to the southwest of Moaning Cave, show a pronounced dry period between 7 and 4 ka, with wetter conditions before 7 ka (Davis, 1999). Owens Lake, also south of Moaning Cave, rose during the early Holocene from a lowstand during the YD and began to fall again after about 7 ka (Bacon et al., 2006).

In southern Nevada, Quade et al. (1998) report the presence of numerous spring-fed “black mats” beginning around 13.8 ka, peaking at 11.5 ka, and then dropping off completely after around 7.4 ka due to drying. Lachniet et al. (2014) present Holocene speleothem $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ records from Leviathan Cave, located in south-central Nevada about 170 km southwest of Lehman Caves. While the $\delta^{18}\text{O}$ record from Leviathan Cave may indicate relevant changes in precipitation source, the $\delta^{13}\text{C}$ record does not show the pattern of mid-Holocene drying shown by our record and the other records summarized here. This difference may be due to control by some of the confounding factors described in section 5.2 (e.g., changes in C_3 vs C_4 plants), or it may suggest divergent climate histories at the two cave sites.

Evidence for a similar hydrological pattern exists in many locations across the western and southwestern United States. In the Guadalupe Mountains of southern New Mexico, Polyak et al. (2004) use speleothem growth as a proxy for wet conditions. Their data suggest a wet period spanning the duration of the YD and lasting until around 10.6

ka. Polyak and Asmerom (2005) demonstrate a lowering of Lake Estancia in central New Mexico around 8.5 ka. In northern New Mexico, a pollen and charcoal record from a bog in the Jemez Mountains suggests a wet early Holocene followed by desiccation after about 8.5 ka (Anderson et al., 2008). Pollen and plant macrofossil records show a pronounced wet period in the Kaibab Plateau of northern Arizona between 11 and 8 ka, possibly due to an enhanced summer monsoon (Weng and Jackson, 1999). In central Arizona, pollen, macrofossil, and diatom records from Stoneman Lake shows pronounced drying after approximately 9.4 ka (Hasbargen, 1994).

Finally, east of the Great Basin in the Wyoming, Colorado, and the central Great Plains region, record compilations in Shuman et al. (2010) and Pribyl and Shuman (2014) find a consistent pattern of rapid drying after ca. 8-9 ka. Though there is substantial variability in the timing and rate of these changes, many of these records show drying beginning after 9.0 ka, and records that indicate abrupt drying cluster around 8 ka (Williams et al., 2010). Lakes in the Rocky Mountains, in particular, show rapid water-level declines at ca. 9-8 ka (Pribyl and Shuman, 2014; Shuman et al., 2010). The transition at this time is also associated with a large-scale shift in North American moisture gradients captured by both lake-level and pollen records east of the Great Plains, which has been attributed to the rapid reduction of the Laurentide Ice Sheet and its effects on atmospheric circulation (Shuman et al. 2002; Shuman et al. 2006).

Great Basin lake-level data document a pronounced decline in regional lake levels beginning between 8 and 8.5 ka (Figure 9). Because this compilation includes lake level records of varying temporal resolution, it may not capture smaller-scale drying events. However, the very clear onset of drying around 8 ka in this compilation attests to the

drying at that time being more widespread and greater in magnitude than, for example, drying during the Bølling-Allerød. The striking resemblance to Mg/Ca and $\delta^{13}\text{C}$ records presented here suggests that our reconstruction of relative changes in infiltration rates above Lehman Caves is representative of a broad portion of the Great Basin, though future work will be needed to more precisely determine the spatial imprint and temporal evolution of mid-Holocene drying.

5.5 Mechanisms for wet early Holocene conditions and mid-Holocene drying

Here we consider three potential explanations for the transition from relatively wet early Holocene conditions in the Great Basin to a drier mid-Holocene climate. First, and most briefly, orbital changes are an unlikely explanation of wet early Holocene conditions. If the shift toward slightly wetter conditions over the last ~4 ka documented by Great Basin lake level records (Figure 8D) and by the growth rate record from a Leviathan Cave stalagmite (Lachniet et al., 2014) is taken to represent a response to insolation changes between the mid- and late Holocene (declining local summer insolation and increasing winter insolation), then early Holocene insolation (high summer insolation, low winter insolation) should have led to dry conditions in the Great Basin. Insolation changes may drive changes in atmospheric circulation reflected in stalagmite $\delta^{18}\text{O}$ (Lachniet et al., 2014), but it appears that at least in the early Holocene, other factors than insolation control Great Basin water balance. An alternative possibility is that the higher insolation in the early Holocene resulted in an intensified North American Monsoon, however, evidence for this is confined to records far south of Lehman Caves (e.g. Weng and Jackson, 1999). Evidence for such a change in the region proximal to

Lehman Caves is not apparent in either pollen records (Madsen et al., 2001; Rhode and Madsen, 1995) or in the $\delta^{18}\text{O}$ record from WR11.

Studies of modern climate (e.g., Schubert et al., 2004; Seager et al., 2005) have found strong connections between tropical Pacific SSTs and precipitation in the western U.S, raising the possibility that changes in either the mean state of the tropical Pacific or in El Niño-Southern Oscillation (ENSO) variability led to the observed Holocene hydrological changes in the Great Basin. Sea-surface temperature (SST) records from both the eastern and western tropical Pacific based on Mg/Ca measurements on planktonic foraminifera indicate an early Holocene SST maximum (Lea et al., 2006; Stott et al., 2004). A mean state of the tropical Pacific characterized by warm SSTs in the early Holocene is a plausible cause of higher precipitation in the Great Basin; however, alkenone-based tropical Pacific SST estimates suggest an opposite Holocene history (a cooler early Holocene) (Leduc et al., 2010).

Holocene changes in ENSO variance may also have strong impacts on Great Basin hydrology. Some reconstructions of Holocene ENSO variability suggest changes that are in qualitative agreement with our records showing drying in the Great Basin after 8 ka and the driest conditions between 4-6 ka. For instance, records of ENSO variance from fossil surf clams from the Peruvian Coast suggest that between 7.5 and 6.7 ka, the dominant spatial mode of ENSO may have produced weaker El Niño events than in the early Holocene. Further, clam, foraminiferal, and coral records suggest a minimum in ENSO variance at 4-5 ka (Carre et al., 2014; Cobb et al., 2013; Koutavas et al., 2006). Additionally, in the modern climate, ENSO-related precipitation variance shows opposite signs in the northern and southern parts of the Great Basin (Wise, 2010); if

consistent through the Holocene, this could potentially account for differences in hydroclimate records from different latitudes in the Great Basin. Though many Holocene records show some important consistency with our reconstruction, their short duration combined with the lack of agreement over whether the inferred variability in ENSO variance over the Holocene is statistically significant (Cobb et al., 2013) prevents us from making a firm link between our records and changes in ENSO. .

We instead favor a third explanation: that the presence of a remnant of the Laurentide ice sheet during the early Holocene influenced storm tracks in western North America, increasing winter precipitation in the interior western US. The rise of pluvial lakes in the Great Basin coincident with the LGM and late-glacial has been long thought to be due to the southward deflection of the westerly winter storm track by the Laurentide ice sheet (Bromwich et al., 2004; COHMAP Members, 1988), but the storm track's response to ice sheet retreat during the latest Pleistocene and early Holocene is poorly understood. Our record indicates pronounced drying near Lehman Caves after 8.2 ka, coinciding with the timing of the collapse of the remnant Laurentide ice sheet over Hudson Bay (Barber et al., 1999), a precursor of the globally recorded 8.2 ka event (Hughen et al. 2000; Kobashi et al. 2007; Lachniet et al. 2004; Thomas et al. 2007, and others) that is thought to be the result of the draining of glacial lakes previously dammed by the ice sheet. Shuman et al. (2002) showed that moisture gradients in eastern North America shifted rapidly at this time, and Williams et al. (2010) find that most Great Plains records showing rapid mid-Holocene drying are clustered around 8 ka. Great Basin lake records compiled in Figure 8 show a similar pattern, providing corroborating evidence of widespread precipitation changes in the North American interior at this time.

Together, this strong temporal correspondence strongly suggests the collapse in ice sheet area as a cause of the drying. Model experiments simulating ~8 ka climate with and without the 8.5 ka remnant ice sheet will be required to test this hypothesis.

Intriguingly, recent climate model results suggest that the connection between ice sheet collapse and Great Basin hydrology may have been through changes in ENSO, linking our second and third explanations above. Braconnot et al. (2011) find that the early- to mid-Holocene orbital configuration favors a minimum in ENSO variance, but that freshwater fluxes from melting ice sheets in the early Holocene could have offset this orbital control, leading to near-modern ENSO variance in the early Holocene. Under this scenario, the combination of remnant ice sheets and high freshwater fluxes in the early Holocene leads to relatively high ENSO variance and wet conditions in much of the Great Basin, as most of the Great Basin lies in the southern part of the ENSO precipitation dipole described by Wise (2010). The lack of significant ice cover after 8 ka might allow mid-Holocene orbital parameters to drive a minimum in ENSO variance and dry the Great Basin. Again, confirmation of this link awaits further development of Holocene records of ENSO variations.

6. Conclusions

Elemental and stable isotope data from two Lehman Caves, NV speleothems provide precisely dated records documenting the drying of the Great Basin from the deglaciation through the mid-Holocene. The strong covariation of Mg/Ca ratios and $\delta^{13}\text{C}$, and their agreement in two overlapping stalagmites, suggest that they track prior calcite

precipitation and are robust proxies for hydroclimate change. Mg/Ca and $\delta^{13}\text{C}$ data suggest wet conditions with possible slow drying between 16.4-15.0 ka. Drier conditions between 15.0 and 12.7 ka, coinciding with the Bølling-Allerød warm period, are suggested by a depositional hiatus in speleothem CDR3. Relatively wet conditions during the Younger Dryas and earliest Holocene are indicated by both speleothems. Mg/Ca and $\delta^{13}\text{C}$ values suggest a transition to slightly drier conditions at 10.3 ka, followed by pronounced drying after 8.2 ka. Dry conditions then persist until the end of the record at 3.8 ka.

The record presented here is broadly consistent with many available climate records from across the Great Basin, including a new compilation of regional lake level records, indicating that it accurately records the onset of large-scale mid-Holocene drying. The timing of the drying's onset as well as established theories for hydroclimate drivers in the Great Basin point to the collapse of the Laurentide Ice Sheet around 8.2 ka as a possible mechanism for the abrupt drying in the Basin. However, this interpretation must be tested through targeted model experiments and more detailed Holocene tropical Pacific records. As insolation changes would likely have led to an opposite pattern of early-to-mid-Holocene change, this study of stalagmite proxies reflecting local water balance suggests that even relatively small changes in ice extent or tropical SSTs can have larger impacts on Great Basin hydroclimate than insolation changes.

7. Acknowledgements

We thank Larry Edwards and Xianfeng Wang for substantial contributions during the early phases of this study. Ben Roberts, Gretchen Baker, and the staff of Great Basin

National Park provided crucial facilitation and assistance with sampling in the field; Ken Adams also assisted with initial sample collection. We also grateful to Aaron Donohoe, Carrie Morrill, and Allegra LeGrande for helpful discussions of climate model results, and to Bryan Hamilton for sharing oxygen isotope analyses of precipitation. We thank Soumen Mallick at Brown University, Jurek Blusztajn and Scot Birdwhistell at WHOI, and Rick Kayser at MIT for their help with mass spectrometry, and Wendy Salmon at the Whitehead Institute for her assistance with the confocal microscope. Siyi Zhang, Michaela Fendrock and Lucy Page also provided important help in the lab. This work was funded by NSF EAR-1103379, the MIT EAPS Student Research Fund, and the Comer Science and Education foundation.

Figure Captions

Figure 1: Map of the Great Basin (red outline) showing the largest extent of Lake Bonneville (gray outline), and the location of Lehman Caves (yellow star) on the western edge of the Bonneville Basin. Map modified from GeoMapApp (<http://www.geomapapp.org/>); base map from Ryan et al., 2009.

Figure 2: Photograph of CDR3 and WR11. Large sample holes are U-Th samples and small holes on the growth axis are stable isotope samples.

Figure 3: Age models (solid lines) and calculated growth rates (dotted lines) for WR11 (top) and CDR3 (bottom).

Figure 4: Stable isotope and trace element data from CDR3 (green) and WR11 (blue) from 16.4 to 3.8 ka. Included are (A) U-Th age control points, (B) $\delta^{18}\text{O}$, (C) $\delta^{13}\text{C}$, (D) Mg/Ca, and (E) Sr/Ca.

Figure 5: Mg/Ca and Sr/Ca data for cave waters collected in May 2013 (red triangles) and January 2014 (blue circles). Rayleigh fractionation curves (black lines) predicted for PCP are also shown. Rayleigh curves are calculated using $D_{\text{Sr}} = 0.125$ and $D_{\text{Mg}} = 0.0125$ (based on modern cave temperature and equations from Day and Henderson, 2013) and initial Mg/Ca and Sr/Ca ratios of dripwater determined from the minimum ratios measured in soil waters above the cave (this study, Supplementary Table 5). (A) Mg/Ca and Ca concentrations; (B) Sr/Ca and Ca concentrations; and (C) Mg/Ca and Sr/Ca. In all panels, increasing PCP drives values toward the right along the distillation curves.

Figure 6: Trace element and stable isotope values interpolated from WR11 records to demonstrate covariation or lack thereof between variables. (A) $\delta^{13}\text{C}$ vs. Mg/Ca, showing covariation throughout the record. (B) $\delta^{13}\text{C}$ vs. Sr/Ca, showing no clear relationship. (C) Sr/Ca vs. Mg/Ca, demonstrating covariation during approximately the latter half of the record. (D) Covariation between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$.

Figure 7: $\delta^{18}\text{O}$ records from Lehman Caves (WR11: bright blue, CDR3: gray blue) and Leviathan Cave from Lachniet et al. 2014 (black). There is some agreement between CDR3 and LC1 from 13-10 ka, but LC1 has substantially more positive $\delta^{18}\text{O}$ values than WR11 during much of the period from 10-5 ka.

Figure 8: Comparison of (B) Mg/Ca and (C) $\delta^{13}\text{C}$ records from WR11 (blue) and CDR3 (green) with (D) lake level compilation data showing the percent of Great Basin lakes at lowstands over time and (E) shows the area of the Laurentide Ice Sheet (LIS) over time (black line) and the rate of change in LIS area over time (dashed gray line) from (Dyke, 2004). Age control points for WR11 and CDR3 shown on top of figure (A). Vertical gray bar indicates initiation of abrupt drying.

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Figure 1

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