Great Basin hydrology, paleoclimate, and connections with the North Atlantic: A speleothem stable isotope and trace element record from Lehman Caves, NV

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**Article Info**

Article history:
Received 7 March 2015
Received in revised form 9 June 2015
Accepted 16 June 2015
Available online xxx

Keywords:
Speleothem
Great Basin
Oxygen isotopes
Carbon isotopes
Trace elements
Termination II
Heinrich Stadial
Effective precipitation

**Abstract**

We present a record of speleothem stable isotope ($\delta^{18}O$, $\delta^{13}C$) and trace element (Mg/Ca, Sr/Ca) variations from Lehman Caves, Nevada for an interval of time (139–128 ka) that encompasses the penultimate glacial termination, Termination II (T-II). Additional growth phases provide data from about 123 ka, a time that correlates with Marine Isotope Stage (MIS) 5e, and about 84 ka and between 82 and 81 ka (MIS 5a). Chronologies from two new stalagmites are anchored by thirty-six uranium–thorium dates. We also present new trace element data from stalagmite LC-2, which has a previously published uranium–thorium chronology and stable isotope record of T-II (Shakun et al., 2011). Our T-II $\delta^{18}O$ record broadly replicates that of the Shakun et al. (2011) and Lachniet et al. (2014) records of this time, recording low values from 139 to 135 ka followed by an approximately 3.5‰ increase over an extended interval between 134 and 129 ka. This rise in $\delta^{18}O$ values occurs during Heinrich Stadial 11 and the associated Weak Monsoon Interval observed in Chinese caves; our record broadly follows the marine termination, rising boreal summer insolation, and the rise in atmospheric CO2. We infer that this shift results from temperature increase due to increasing atmospheric CO2 and potentially a change in moisture source or precipitation seasonality from greater influence of the North American Monsoon accompanying summer insolation increase. It is also plausible that the melting of the ice sheet itself may have contributed to both temperature and precipitation seasonality changes. Trace element ratios and $\delta^{13}C$ values are largely decoupled from $\delta^{18}O$ values, showing minimal variation between 139 and 130 ka, for the duration of the Chinese Weak Monsoon Interval. However, these values increase sharply between 130 and 128 ka, which we interpret as increased prior calcite precipitation driven by a transition from wet to dry conditions. This abrupt drying event coincides within dating uncertainties with the abrupt strengthening of the East Asian summer monsoon, which marks the end of both the Weak Monsoon Interval and Heinrich Stadial 11. Our records demonstrate a link between North Atlantic climate and Great Basin moisture during this time that is consistent with the interpretation of data from the last glacial period and may result from abrupt shifts in Atlantic Meridional Overturning Circulation affecting the strength of the Aleutian Low.

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1. Introduction

The Great Basin has been recognized for over a century as the site of vast periodic hydrologic changes, revealed by the rise and fall of paleolakes Bonneville, Lahontan, and other interbasin lakes (Broecker and Orr, 1958; Eardley et al., 1973; Gilbert, 1890; Oviatt, 1997). Due to dating limitations of lake cores and shoreline
outcrops, however, most studies have focused on the Last Glacial Maximum (LGM) and most recent glacial–interglacial transition Termination I; less is known about the nature and timing of Termination II (T-II) and other times beyond the reach of radiocarbon dating in the Great Basin. This information is critical to define and model the mechanisms controlling orbital and millennium-scale climate and hydrologic change, as well as to understand water balance in a region projected to become even drier in the coming decades (Seager et al., 2007).

Glaciations have occurred following orbital cycles through the Quaternary. Glacial periods are followed by rapid deglaciations, or terminations, that may be triggered when rising summer boreal insolation intensity causes the initiation of melting of large, isostatically-balanced continental ice sheets. After the initiation of melting, the disintegrating ice sheets released massive quantities of meltwater and icebergs into the North Atlantic (Heinrich Events), reducing meridional overturning circulation and increasing ocean stratification, leading to expansive winter sea ice formation (Cheng et al., 2009; Denton et al., 2010). This sea ice cover has been implicated both in creating extreme seasonality in and around the North Atlantic and in quickly propagating the resulting winter anomaly globally (Denton et al., 2010). In the Great Basin, lake level rise is associated with Heinrich Stadials (HS) (Lachniet et al., 2013a; Oviatt, 1997) and the Younger Dryas stadial (Oviatt, 1997) and increased freshwater inputs to Lake Bonneville occur during H2 (McChee et al., 2012). These hydrological changes are possibly due to the alteration of Walker circulation or strengthening of the Aleutian Low, which would enhance moisture delivery to the region (Okumura et al., 2009), or via an ice-sheet modulated jet stream position (Anderson et al., 1988). Speleothems from the southwestern United States also show clear responses to Heinrich stadials (Asmerom et al., 2010; Musgrove et al., 2001; Wagner et al., 2010).

However, the connection between Great Basin and North Atlantic climate is not as clear prior to the last glacial cycle. The first absolute dated, high-resolution record of T-II in the Great Basin was the Devils Hole vein calcite δ18O record (Ludwig et al., 1992; Winograd et al., 1992). Recording a 2% increase in δ18O that seemed to show the penultimate deglaciation greater than 10 kyrs before a rise in Northern Hemisphere summer insolation and H11, this record sparked years of debate on the controls of the Devils Hole isotopic record and challenged the conventional understanding of the mechanisms by which orbital forcing modulates global climate and glacial cycles (Herbert et al., 2001; Imbrie et al., 1993; Winograd, 2002). To test the possibility of an early T-II elsewhere in the Great Basin, Shakun et al. (2011) developed a speleothem record from Lehman Caves (Fig. 1) that showed a 2% rise in δ18O values between 133.5 ± 0.7 and 128.9 ± 0.4 ka, consistent with global marine and terrestrial records of T-II and orbital forcing as shown in Fig. 2 (Cheng et al., 2009; Drysdale et al., 2005; Imbrie et al., 1984; Kawamura et al., 2007). Subsequent speleothem work supported and extended the Shakun et al. (2011) T-II record (Lachniet et al., 2014).

In this study, we test for replication of this key oxygen isotope record immediately before, during, and subsequent to T-II. Moreover, we obtained trace metal (Mg/Ca and Sr/Ca) ratios as a means of characterizing local to regional-scale effective precipitation (Fig. 2). Trace metals have provided valuable hydrological insights in speleothems from diverse locations (Cruz et al., 2007; Orland et al., 2014; Roberts et al., 1998; Tan et al., 2014), and in modern Lehman Caves conditions appear to be a measure of aridity, reflecting the amount of prior calcite precipitation (PCP) occurring in the epikarst above the cave (Steponaitis et al., 2015). By combining oxygen and carbon isotope analysis with trace metal analysis, we can more fully characterize climatic conditions in the region before, during, and after T-II, placing them in the context of changing climatic conditions elsewhere in the world.

In addition to T-II, we report stable isotope data on two short growth phases within MIS 5a (one at ~ 84 ka and another between 82 and 81 ka). At present, there are discrepancies among MIS 5a and 5c cave-based oxygen isotope records at locations near Lehman Caves, notably between the Denniston et al. (2007) and the Lachniet et al. (2014) records. Our data pertain to this discrepancy.

1.1. Site description

Lehman Caves (39.01 N, 114.02 W) in Great Basin National Park (GBNP) is an actively dripping, highly decorated cave in the eastern Great Basin region, near the southwestern edge of paleolake Bonneville (Fig. 1). Located in the eastern flank of the southern Snake Range in a ridge of the Pole Canyon Limestone, Lehman Caves is largely between 30 and 60 m below ground and the blasted-tunnel entrance used by modern visitors is at an elevation of 2080 m. The Pole Canyon Limestone comprises 5 members of alternating light and dark, thin to massive carbonate with minor quartzite and shaly units, and dolomite has been identified in nearby ranges where this limestone crops out (Drewes and Palmer, 1957).

Environmental conditions within the cave are stable year-round, at 100% relative humidity and 11.0 °C, reflecting mean annual temperature. A recent study by Paul and Thodal (2014) indicates that cave pool and spring water δ18O values are similar to the volume-weighted mean annual precipitation δ18O values of ~15.5% (their Fig. 4). Groundwater recharge at Lehman Caves can therefore be attributed to mainly winter precipitation, which is sourced from storms tracking in from the Pacific Ocean (Benson and Klieforth, 1989; Friedman et al., 2002a,b; Paul and Thodal, 2014; Prudic and Clancy, 2009), with potentially minor amounts of summer precipitation (Paul and Thodal, 2014). Hydrologic communication between the cave and surface is rapid, on the order of 1–4 weeks (Steponaitis et al., 2015).

2. Materials and methods

2.1. Sample collection and preparation

WR-41 and IR-3 were collected, already broken, from the West Room and Inscription Room of Lehman Caves, respectively. We prepared them for subsampling by slicing in half along the growth axis and polishing the cut faces. One half of each speleothem was set aside for preservation at the National Park Service office in GBNP. The preparation, dating, and stable isotope subsampling of LC-2 are described in Shakun et al. (2011). All procedures requiring drilling were performed using a handheld drill with a 0.5 mm carbide bit.

2.2. U–Th dating and age modelling

Except for 4 exploratory dates (Fig. 3, large pits), we drilled much smaller samples (5–15 mg of powder) than is typical for speleothem studies because of the relatively low growth rates of our stalagmites. This strategy allowed us to drill within single stratigraphic layers as thin as 1 mm. Forty sample powders were weighed, dissolved in 14N HNO3 and spiked with a 233U/236U/229Th tracer, the concentration of which was periodically calibrated against aliquots of HU-1 from the University of Minnesota’s stock (Cheng et al., 2013). U and Th were separated and purified following the iron co-precipitation and cation exchange procedures of Edwards et al. (1987), and analyzed at the University of Minnesota on the multi-collector Thermo-Finnigan Neptune inductively coupled plasma mass spectrometer using peak-jumping routines.
on the SEM. Despite our small sample sizes of 0.81–7.90 ng of $^{238}$U (3.20 ng of $^{238}$U on average), we maintained relatively high analytical precision due to high (>1%) ionization and transmission efficiencies. Raw data were corrected with instrumental background, tailing, yield, procedural blank data, and an initial $^{230}$Th/$^{232}$Th atomic ratio of $4.4 \times 10^{-6} \pm 2.2 \times 10^{-6}$. This ratio represents a material at secular equilibrium with a bulk earth $^{232}$Th/$^{238}$U value of 3.8. After an early period of somewhat higher procedural blanks, we thoroughly cleaned all teflonware and attained average blank values of 70 $\pm$ 6 a g $^{234}$U, 0.9 $\pm$ 0.2 pg $^{238}$U, 26 $\pm$ 9 a g $^{230}$Th, and 0.3 $\pm$ 0.1 pg $^{232}$Th. Blank corrections are well within analytical (2σ) dating errors, which are on the order of 0.7–2.2%, even with our small sample sizes.

Age models were constructed using OxCal version 4.2 (Ramsey, 1995; Ramsey and Lee, 2013). Among our forty samples we had four pairs of replicates. Each pair is represented in the age models as a single age control point having the pair’s mean age and quadratically combined error. Of the remaining thirty two dates, one from WR-41 and two from IR-3 were discarded to resolve age reversals with non-overlapping error bars, as was an additional date from WR-41 that was significantly younger than surrounding ages.

Separate age models were calculated for each growth interval using the P_Sequence method and a variable $k$ parameter of 0.01 mm$^{-1}$ $\pm$ 2 orders of magnitude to allow OxCal to determine the flexibility and ultimately the slopes of the age models.

2.3. Stable isotopes

A total of 352 samples from WR-41 and IR-3 weighing approximately 100 μg each were extracted at a 0.5 mm interval. Drilling was performed along the growth axis, except in those areas of WR-41 which contain a complicated stratigraphy, where we drilled off to the side. Our transitions to the growth axis are inspected via 1.5–5.5 mm of replication and a Hendy test to attempt to assess disequilibrium crystallization (Hendy, 1971). Sample powders were analyzed at the University of Arizona, Tucson, on a Finnigan MAT 252 gas-source isotope ratio mass spectrometer after dissolution in an attached KIEL-II automated carbonate preparation device. Analytical (2σ) uncertainties for $\delta^{18}$O and $\delta^{13}$C values are approximately $\pm 0.22%o$ and $\pm 0.16%o$, respectively.

2.4. Trace elements

Along our stable isotope drill transects in WR-41 and IR-3 and the growth axis of LC-2, we recovered 203 1–2 mg samples.
Subsampling resolution is 1–2 mm in WR-41 and IR-3, and 5–10 mm in LC-2. At MIT, 1 mg of each sample was dissolved in 8N HNO₃, diluted to ~0.05% Ca in 0.5 M HNO₃, and analyzed for ²⁵Mg, ²⁶Mg, ⁴³Ca, ⁸⁶Sr, and ⁸⁷Sr on a Thermo VG PQ2⁺ quadropole mass spectrometer. Gravimetric Mg–Ca–Sr standards with similar Ca concentrations were run every 10 samples to monitor relative sensitivity for Mg, Ca, and Sr and calculate the Mg/Ca and Sr/Ca sample ratios. Elemental ratios in the standards varied by <10% throughout each run. Raw data were corrected for procedural blanks and instrument background. Analytical uncertainties for elemental ratios are <2%, and mean reproducibility of ratios in replicate samples measured on separate days is better than 10%.

2.5. Mineralogy

To constrain the mineralogy of our speleothems, approximately 1–2 mg of powder was drilled from a total of 5 points in WR-41 and IR-3. The samples were analyzed at the University of Minnesota Characterization Facility on a Bruker-AXS Microdiffractometer for a minimum of 300 s/sample. Phase matching to diffraction peaks was performed using JADE 2010 software.

3. Results

3.1. U–Th dating and mineralogy

We present chronologies from four periods of time: 139.0 ± 1.0–129.2 ± 1.1 ka, 123.1 ± 0.2–122.8 ± 0.2 ka, 84.3 ± 0.2–84.0 ± 0.3 ka, and 82.4 ± 0.3–81.3 ± 0.3 ka (Fig. 3 and Fig. 4). These chronologies were assembled using a total of thirty-two age control points, including four points representing replicate analyses (Table 1). Four dates were not used due to their anomalous ages. Anomalously old samples may potentially be explained by: open system behavior causing the preferential leaching of uranium, thorium being insoluble in most natural surface waters (Ivanovich and Harmon, 1982); greater initial ²³⁰Th (i.e., significant detrital material) than accounted for in our correction (Dorale et al., 2004); or sample contamination with older or younger material forming hiatuses. For the anomalous date at the

Fig. 3. Age models and high-resolution scan of WR-41. The depth scale of the plot is calibrated to match the image. Growth phase age models calculated from dates shown in green are represented by black lines with purple error envelopes. The slopes of our age models are determined by the model rigidity set by OxCal using a small k parameter of 0.01 mm⁻¹ ± two order of magnitude (increasing k increases rigidity). Dates indicated in orange were not used in age models. Data from this stalagmite, WR-41, is distinguished throughout this paper by green coloration and/or circular markers. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Fig. 4. High resolution scan and age model of IR-3. The depth scale of the plot is calibrated to match the image. The age model for the large main growth phase is constructed similarly to those of WR-41 (Fig. 3). Dates not used in the model are significantly older than those shown in blue and are thus not included on the plot. Data from IR-3 is distinguished throughout this paper by blue coloration and/or triangular markers. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
top of oldest growth phase in WR-41, we favor the last explanation. Precise hand-drilling is difficult and we attempted to date the material immediately below a hiatus. Increased initial $^{230}$Th is unlikely to be a problem due to the age of our speleothems (Dorale et al., 2004), as it would have partially decayed, and the $^{230}$Th/$^{232}$Th ratios in our anomalously old samples are high, indicating low detrital content, and are consistent with other samples. The anomalous dates in IR-3 may be a result of localized open system behavior. Unusually, we obtained an anomalously young date in WR-41 (seen in Fig. 3 as the large dating pit near the top of the stalagmite). Geochemical behaviors typically result in the appearance of greater age; this particular sample may intersect younger layers deeper in the stalagmite.

The oldest portion of our chronology is partially replicated in our newly analyzed speleothems, IR-3 (131.5 ± 0.8–129.2 ± 1.1 ka) and WR-41 (139.0 ± 1.0–129.8 ± 0.6 ka), and in LC-2 (133–128 ka). Growth rates are low during this time interval, varying between 0.9 and 33.3 μm/yr with an average of 19.3 μm/yr in WR-41, and in IR-3 ranging from 0.2 to 14.0 μm/yr, with an average of 10.3 μm/yr. Uranium concentrations are also low at this time, ranging between 140.4 and 380.0 ppb $^{238}$U in IR-3 and WR-41. No pattern or trend in $^{238}$U concentration is apparent, but in WR-41 and LC-2, a marked increase in $^{234}$U initial begins around 131–130 ka, rising from respective baselines around 250± and 200± to peak at 433.5± and 387.2±. In IR3, $^{234}$U initial increases, from 391.3± to a peak at 551.6± immediately before the end of the growth phase, which has a $^{234}$U initial of 445.4±. Micro-XRD analyses indicate IR-3 and the oldest growth phase of WR-41 are calcite.

The younger periods (123, 84, and 82–81 ka) presented are each represented by one growth phase in WR-41. The growth period around 123 ka is characterized by relatively high growth rates and $^{238}$U concentrations, on average 185 μm/yr and 768 ppb. The $^{234}$U initial ranges between 276.3± and 317.5±. This phase consists of Mg-calcite; we were unable to find any traces of aragonite. The
intervals at 84 ka and 82–81 ka have slow growth (5–21 μm/yr) and high $\delta^{234}$Uinitial values (408–541‰) and uranium concentrations (426–807 ppb).

3.2. Stable isotopes

Oxygen isotope values in WR-41 and IR-3 range from $-14.3$‰ to $-10.0$‰ (VPDB) as shown in Fig. 5. In WR-41, $\delta^{18}$O values average $-13.5$‰ until 136–135 ka, when they begin a steady rise to a peak of $-10.6$‰ around 129.8 ka. This rise is largely replicated in IR-3 and the previously published LC-2 record (Shakun et al., 2011). Our total range in $\delta^{13}$C values is approximately 8‰. In WR-41, $\delta^{13}$C values are low throughout much of the record, reaching a minimum value in our replicated segment of $-8.4$‰ (on-axis) at 131 ka then rapidly rising to a maximum of $-0.1$‰. The rise in IR-3 $\delta^{13}$C values from $-6.0$‰ to $-0.3$‰ occurs within error at the same time as observed in WR-41.

At 123 ka, $\delta^{18}$O and $\delta^{13}$C values are highly variable, ranging from $-14.5$‰ to $-11.3$‰ and from $-7.4$‰ to $-3.3$‰, respectively. At 84 ka $\delta^{18}$O values average $-12.1$‰, distinctly heavier than $\delta^{18}$O values in the younger phase (82–81 ka), which have a mean of $-13.5$‰. $\delta^{13}$C values are more similar, and range between $-3.4$ and $-6.5$‰. The stable isotope trends in LC-2 are described in Shakun et al. (2011), $\delta^{18}$O and $\delta^{13}$C values are moderately to poorly correlated ($r^2 < 0.4$ in IR-3, $r^2 < 0.1$ in WR-41) in both new stalagmites and growth phases except in WR-41 at 82–81 ka where $r^2 = 0.45$. Values in our Hendy test (Hendy, 1971) transect varied by less than 0.6‰ in both $\delta^{18}$O and $\delta^{13}$C and were not significantly correlated ($r^2 = 0.31$) with each other or with distance from the growth axis ($\delta^{13}$C-distance $r^2 = 0.22$, $\delta^{18}$O-distance $r^2 = 0.01$).

The efficacy of Hendy tests at revealing isotope disequilibrium in speleothems, however, has come under criticism. Sampling along a single stratigraphic horizon is problematic, different equilibrium/disequilibrium conditions may dominate at different points along the growth horizon, and if carbon and oxygen isotopes are responding to the same controls, then a high correlation is likely. As it turns out, Hendy also proposed a replication test, a strategy followed in early studies by Dorale et al. (1998) and Wang et al. (2001). Dorale and Liu (2009) discuss the Hendy and Replication Tests and argue that the Replication Test is the more robust of the two. As the

Fig. 5. Results from stable isotope ($\delta^{18}$O, $\delta^{13}$C) and trace element (Mg/Ca, Sr/Ca) analysis and U–Th dating (ages and $\delta^{234}$Uinitial) of IR-3 (blue triangles), WR-41 (green circles), and LC-2 (red squares). Stable isotope data and U–Th dating results from LC-2 were originally published in Shakun et al. (2011). LC-2 data is distinguished throughout this paper by red coloration and/or square markers. Age control points from WR-41 and IR-3 are shown at the bottom of the panel. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
d\textsuperscript{18}O, d\textsuperscript{13}C, and trace elements (discussed below) of our new speleothem and LC-2 largely replicate each other in both magnitude of variations and absolute values, and the timing of these variations are within error, we find it unlikely that our speleothem records have been drastically altered by disequilibrium effects or vadose-zone processes.

3.3. Trace elements

In WR-41, our Mg/Ca record (Fig. 5) is characterized by an abrupt increase around 130 ka from ratios ranging between 2.1 and 4.5 mmol/mol to a peak of 8.5 mmol/mol. A similar increase is seen in LC-2 around 129 ka, reaching a maximum of 9.6 mmol/mol. Mg/Ca ratios in IR-3 appear to replicate those in LC-2 and WR-41 before their abrupt increases. High Sr/Ca values in LC-2 and WR-41, up to 39 µmol/mol, are recorded around 130–129 ka. Both speleothems also have a period of similarly high Sr/Ca after initiation of growth. In WR-41, this period lasts from about 139 to 135 ka; in LC-2, this occurs between 134 and 132 ka, at the same time that WR-41 records its lowest Sr/Ca ratios (a minimum of 18.3 µmol/mol). Sr/Ca in IR-3 positively trend over a range from 24.7 to 36.1 µmol/mol.

4. Discussion

4.1. Interpretation of d\textsuperscript{18}O

Our Lehman Caves d\textsuperscript{18}O record replicates the 2‰ rise in speleothem d\textsuperscript{18}O values that Shakun et al. (2011) observed between ~133 and 129 ka, as well as (within error) the lead-up to and the initiation of this rise, captured by Lachniet et al. (2014) in other Lehman Caves speleothems. This increase in d\textsuperscript{18}O values has been interpreted in terms of both regional air temperature and moisture source change (Denniston et al., 2007; Lachniet et al., 2014). Previous speleothem studies from the Great Basin and the southwestern United States (Asmerom et al., 2010; Denniston et al., 2007) have derived net temperature dependencies of ~0.25‰/°C, based on a mid-latitude air temperature–precipitation relationship of 0.5–0.69‰/°C (Dansgaard, 1964) and a calcite–water oxygen isotope temperature dependent fractionation of ~−0.2‰/°C. If all of our approximately 3.5‰ increase in d\textsuperscript{18}O values is due to temperature change, this corresponds to a temperature rise of 10–14 °C. A wide range of possible relative Great Basin paleotemperatures during glacial conditions...
has been suggested, from 3 to 12 ºC lower than modern MAT (Hostetler and Benson, 1990; Kaufman, 2003; Quade et al., 2003); we note that MIS 5e was globally ~2 ºC warmer than today (Rohling et al., 2008). Our δ18O signal may track glacial to interglacial temperature change. However, although temperature alone may explain our signal, we cannot rule out a moisture source effect.

Typically, moisture source in the Great Basin is invoked as a change in the latitude of moisture origin of winter Pacific storms (Friedman et al., 2002a; Lachniet et al., 2014), though some studies indicate this relationship may be obscured by intrabasin recycling effects (Ingraham and Taylor, 1991). Most of the northern Great Basin, including Lehman Caves, receives around 20% of its annual precipitation during the summer (Douglas et al., 1993; Xie and Arkin, 1997), as shown in Fig. 1; this precipitation is dominantly small-scale convection and is not connected to the North American summer monsoon (Higgins et al., 1997). Approaching the end of T-II, however, summer insolation was higher than today, and it is possible that an increased land-sea temperature gradient in boreal spring and early summer led to northward movement of isotopically heavy summer precipitation associated with the North American Monsoon originating from the tropical Pacific and Gulf of California. Alternatively, a disintegrating ice sheet may have allowed the polar jet stream to shift northwards, decreasing the winter precipitation component. In each case, increasing δ18O values would reflect changes in both temperature and precipitation seasonality.

Detailed back trajectory and moisture uptake analysis coupled with experimental and modelled precipitation isotopes will assist in understanding the role of moisture source and seasonality in Great Basin precipitation δ18O. Relevant to this discussion, a recent simulation of the impact of precessional changes on precipitation and precipitation δ18O using an isotope-enabled atmospheric general circulation model found a decrease of >1‰ in the annual-mean δ18O value of precipitation in the southern and central Great Basin associated with greater Northern Hemisphere summer insolation at 218 ka relative to 207 ka (Battisti et al., 2014, Fig. 7). As this insolation difference is only ~25% greater than that occurring across T-II, the model results suggest that a similar drop in precipitation-weighted δ18O may have occurred during T-II due to increasing boreal summer insolation. Such a drop in precipitation δ18O appears hard to reconcile with speleothem δ18O changes during T-II, requiring an unrealistically large rise in cave temperature to generate the observed rise in speleothem δ18O if precipitation δ18O was simultaneously falling. As the model does not include ice volume changes, the results do not rule out deglacial changes in Great Basin δ18O related to ice sheet effects on precipitation seasonality.

As for the youngest parts of our record, ~ 84 ka and ~ 82–81 ka, the older interval appears to capture the later part of the increase in δ18O values at the end of the Goshute Cave (Fig. 1) δ18O record associated with Dansgaard-Oeschger (D-O) event 21 (Demistont et al., 2007), as shown in Fig. 7. Neither our record nor the Goshute Cave record at this time replicates the contemporaneous portion of the Lachniet et al. (2014) record from Leviathan Cave (Fig. 1). Applying the Leviathan Cave dripwater correction (Lachniet et al., 2014) to both Lehman and Goshute Caves' δ18O values (Fig. 6, panel 3), neither curve is brought completely into line with the Leviathan Cave record. Lehman and Goshute Caves' δ18O values are 1–5 ‰ lighter, except for the Lehman Caves δ18O values at 84 ka. The Leviathan Cave speleothem dripwater correction also cannot reconcile the difference in curve shape between 84 and 81 ka, when our δ18O values decrease and Leviathan Cave values rise. We observe that Devils Hole δ18O values (Winograd et al., 1992) are similar to Goshute Cave δ18O values around this time, around ~16‰, though Devils Hole is at a lower latitude than

Leviathan Cave (Fig. 1). This is probably due to the relatively higher temperatures (~34 ºC) of calcite precipitation at Devils Hole (Copley et al., 2007).

4.2. Interpretation of trace elements and δ13C

Mg/Ca and Sr/Ca ratios have often provided a perspective on hydrological and environmental conditions that is complementary to δ18O records (Cruz et al., 2007; Orland et al., 2014; Roberts et al., 1998; Tan et al., 2014). Speleothem trace element chemistry reflects drip water geochemistry, which is controlled by a number of factors that can be briefly stated as 1) water-soil-rock interactions 2) amount of prior calcite precipitation (PCP) in the epikarst or the cave above the drip site and 3) mixing of groundwaters with different residence times and/or flow paths (Fairchild et al., 2000; Tremaine and Froelich, 2013). Understanding which factors influence drip water trace elements requires analysis on a cave-by-cave and stalagmite-by-stalagmite basis. Steponaitis et al. (2015) examined soil water and drip water geochemistry at Lehman Caves, and found that while changes in soil geochemistry and groundwater mixing are unlikely to be the dominant controls on our trace element ratios, variations in Mg/Ca and Sr/Ca fit a PCP paradigm. Progressive enrichment in dripwater trace element concentration with greater PCP occurs along a constant Mg/Sr slope, as Mg and Sr have distribution coefficients less than one and therefore preferentially remain in solution during calcite precipitation (Day and Henderson, 2013). Recent studies indicate that this slope depends on initial Mg and Sr dripwater compositions set by water-soil-rock interactions (Tremaine and Froelich, 2013). As such, Mg/Sr is likely to vary within a single cave as well as geographically, since different speleothems receive dripwater sourced from different flow paths through the epikarst, but similar signs and timing of changes are expected.
Observations of Mg/Ca and Sr/Ca in conjunction with δ13C can lead to a more conclusive diagnosis of the likely drivers of both proxies. Controls on δ13C values include hydrologic and biologic factors. One potential mechanism is PCP. During calcite precipitation, degassing of CO2 preferentially releases 12C, leading to δ13C enrichment (Johnson et al., 2006). Longer water residence times or changes in the mode (open vs. closed) of host rock (δ13C value of ~0‰) dissolution may also lead to enriched δ13C values (Genty et al., 2001; Hendy, 1971). δ13C values may also be elevated via changes in vegetation density and soil respiration rates (Baldini et al., 2005; Hellstrom et al., 1998) or soil-water residence times (Baker et al., 1997), as low density and short residence times can increase the proportion of isotopically heavier atmospheric CO2 in seepage water (Genty et al., 2001; McDermott, 2004). However, wetter conditions with higher infiltration rates would be associated with lower trace elemental ratios, particularly Mg/Ca, rather than the higher values that we observe. Although Hellstrom et al. (1998) observed a magnitude of carbon isotopic change similar to ours (~ 8‰ compared to ~ 8‰), the dramatic shift in vegetation density required is highly unlikely to have occurred here. Changes in vegetation type (C3 vs C4) (Dorale et al., 1992) are also unlikely. Pollen and midden evidence indicates that C3 plants have dominated this region for at least the last 40 kys (Madsen et al., 2001). Although we cannot completely rule out contributions to our δ13C signal from soil respiration changes and greater water-rock interaction, the magnitude of our signal and its similarity with our trace element ratios, particularly Mg/Ca, suggest that the major δ13C variations in our record are driven by changes in the amount of CO2 degassing and calcite precipitation.

Enhanced prior calcite precipitation may be driven by CO2 variations and ventilation within the cave environment, but PCP is most often associated with wet-dry variability (Johnson et al., 2006). During dry periods, water residence time is increased, drip rates are decreased, and dry pockets within the epikarst may exist, yielding time and opportunity for degassing in the epikarst and a higher proportion of calcite to precipitate before reaching the speleothem (Fairchild et al., 2000). At Lehman Caves, modern dripwater Mg/Ca and Sr/Ca ratios show high covariability, indicating that PCP is likely the dominant control on modern dripwater trace metals (Steponaitis et al., 2015).

In Figs. 7 and 8, we examine the whole-rock covariations in WR41, IR-3, and LC-2 among δ13C, Mg/Ca, and Sr/Ca. These are generally poor (r² < 0.4), except in IR-3. It is possible that growth rate effects may exert an influence on Sr/Ca in our samples, particularly in WR-41 (r² = 0.45), which experienced large changes in growth rate. Some studies have found enhanced Sr/Ca with increased growth rate (Fairchild and Treble, 2009; Tan et al., 2014), suggesting that above an instantaneous rate of 0.5 mm/yr Sr/Ca will primarily reflect growth rate (Fairchild and Treble, 2009). However, other studies have reported no effect of growth rate on Sr/Ca (Cruz et al., 2007; Day and Henderson, 2013; Treble et al., 2003).

In Fig. 9 we inspect two distinct periods of growth in WR-41, 131–129 ka and 139–135 ka, in order to further examine the temporal variations in our trace element and δ13C relationships. In the older period, Mg/Ca covaries with neither δ13C nor Sr/Ca. In stalagmites spanning the end of the last deglaciation and the early Holocene from Lehman Caves, Steponaitis et al. (2015) observe a similar lack of correlation between Mg/Ca and Sr/Ca before a significant drying event beginning at 8.2 ka. These authors suggest that variations in Sr inputs from Sr-rich dust, a reduced sensitivity to PCP due to the order of magnitude greater distribution coefficient of Sr, or a combination of both factors may play a role. In contrast, Mg/Ca is significantly correlated with both Sr/Ca (r² = 0.50, p = 0.015) and δ13C (r² = 0.81, p = 0.0002) between 131 and 129 ka, similar to what we see in IR-3 (~131.5–128.8 ka). This is diagnostic of PCP control on groundwater trace element and carbon isotope compositions. At this time the LC-2 δ13C (Shakun et al., 2011) does not change significantly and does not correlate well with Mg/Ca. However the large increase in LC-2 Mg/Ca actually postdates the portion of the speleothem that was sampled for stable isotopes, so we do not, at present, have a good measure of the relationship between Mg/Ca and δ13C in LC-2 at times when we expect large changes in both.

Our interpretation is thus that wet conditions in this area persisted at least until ~131 ka. Abrupt drying occurred between 131 and 129 ka, as suggested by dramatic increases in δ13C values, Mg/

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![Fig. 8](image-url). Mg/Ca (left) and Sr/Ca (right) against δ13C. Color and symbology are identical to that of Fig. 7. In the left panel, for IR-3 and WR-41 p < 0.002; for LC-2 p = 0.333. In the right panel, p = 0.0001, 0.025, and 0.864 for IR-3, WR-41, and LC-2, respectively.

Please cite this article in press as: Cross, M., et al., Great Basin hydrology, paleoclimate, and connections with the North Atlantic: A speleothem stable isotope and trace element record from Lehman Caves, NV, Quaternary Science Reviews (2015), http://dx.doi.org/10.1016/j.quascirev.2015.06.016
Ca, and potentially by the cessation in growth of all three of our speleothems immediately afterwards. Taken as a whole, we suggest that these shifts in Mg/Ca and δ13C values are due to increasing amounts of prior calcite precipitation (PCP) at Lehman Caves.

Significant variations in moisture and cave paleohydrology are further supported by δ234Uinitial values. As shown in Fig. 5, δ234Uinitial rises sharply, coincident with our increases in trace metal ratios and δ13C values between 131 and 129 ka. Speleothem δ234Uinitial reflects the δ234U/238U disequilibrium of the water from which it formed (Hellstrom and McCulloch, 2000). This disequilibrium arises from the ejection of 234U into solution via alpha decay (Fleischer and Raabe, 1978; Ivanovich and Harmon, 1982; Osmond, 1980). High 234U may be due to low infiltration rates, or dry conditions with infrequent wetting events, allowing 234U to accumulate to high abundances between events (Ayliffe and Veer, 1988; Hellstrom and McCulloch, 2000).

4.3. Great Basin paleoclimate and paleohydrology

Our record of rising δ18O values between 134 and 129 ka clearly correlates with the global atmospheric CO2 increase from ~180 to 280 ppmv recorded in the Vostok ice core (Petit et al., 1999), with increasing boreal summer insolation (Fig. 6), and with the marine termination in the timescale proposed by Cheng et al. (2009). Thus, we propose that our inferred increase in cave temperature is related to atmospheric CO2 increase. Insolation increase could plausibly have enhanced the North American monsoon, resulting in a higher proportion of summer rainfall, thereby contributing to the rising δ18O values. It is also possible that the melting of the Laurentide ice sheet itself could have affected both temperature and precipitation seasonality, contributing to the observed rise in δ18O values.

However, the abrupt decrease in effective precipitation inferred from trace element data, δ13C and δ234U is essentially decoupled from the more gradual increase in temperature (and possible change in precipitation seasonality) signaled by δ18O. The effective precipitation at Lehman Caves remained high until ~130–128 ka, as demonstrated by the small variability of δ13C and Mg/Ca values prior to this time interval, followed by striking increases in both during this window. Thus, the region experienced abrupt drying synchronous within dating uncertainties with the abrupt rise in East Asian Monsoon intensity recorded in Chinese stalagmites and taken to represent the end of H11, as shown in Fig. 6 (Cheng et al., 2009; Wang et al., 2008). The anti-correlation between effective precipitation in the Bonneville Basin and monsoon intensity in East Asia found during H11 in our record parallels the anti-correlation between δ18O records from Hulu Cave stalagmites and δ18O in precisely dated Lake Bonneville carbonates during H2 and H1 (McGee et al., 2012).

Increased precipitation in western North America during glacial intervals has been attributed to a split or southerly shifted mean polar jet stream position due to the influence of the Laurentide ice sheet (Bromwich et al., 2004; Anderson et al., 1988). It has been suggested that over shorter timescales, e.g. during Heinrich events, moisture delivery to the region was further enhanced via suppression of Atlantic Meridional Overturning Circulation (AMOC) and strengthening of the Aleutian Low (Chiang et al., 2014; Okumura et al., 2014).
et al., 2009). Studies of modern instrumental data demonstrate this connection between Atlantic conditions and North American climate (Enfield et al., 2001; McCabe et al., 2004; Sutton and Hodson, 2005, 2007), indicating that colder Atlantic sea surface temperatures, which would occur during AMOC suppression, lead to reduced winter and summer precipitation in the western United States. Enhanced effective precipitation during Heinrich Stadials is consistent with widespread reports of Great Basin lake highstands clustering around H1 (Munroe and Laabs, 2013a). The decoupling of δ18O from enhanced effective precipitation suggests that winter moisture source may not have a significant role in the δ18O signal; instead we favor temperature and perhaps a changing seasonality of precipitation as the termination progressed.

Other Great Basin paleoclimate records suggest a similar decoupling between warming and hydrological changes during the last deglaciation. For instance, glaciers in the American West exhibited substantial retreat between the LGM and Belling-Allerød (BA) (Shakun et al., accepted, Nature Comm), implying considerable summer warming across the H1 interval. The Leviathan Cave δ18O record also exhibits a general 5‰ enrichment from H1 to the early Holocene (Lachniet et al., 2014), reminiscent of the gradual Lehman Cave δ18O shift across T-I. In contrast, Great Basin pluvial lakes recovered to large highstands during H1 (Munroe and Laabs, 2013b) followed by rapid regressions after the event (Munroe and Laabs, 2013b). These T-I patterns thus mirror the sequence we have identified over T-II, with an abrupt drying following a Heinrich event and commencing several millennia after the onset of a general warming trend.

Speleothem records from the Great Basin and southwestern US also further illustrate connections between western North America and North Atlantic climate and raise new questions. Wetter winter conditions are recorded in a Cave of the Bells, Arizona speleothem during stadials (Wagner et al., 2010). Another speleothem record from Fort Stanton Cave, New Mexico, clearly captures the Younger Dryas stadial and the last five Heinrich Stadials, recording increased contribution of winter precipitation during stadials. In modern conditions, higher winter precipitation sourced from the Pacific is associated with higher mean annual rainfall (Asmerom et al., 2010). In contrast, Feng et al. (2014) argue that the higher growth rates seen in the Fort Stanton record and their Texas speleothems during the warm BA indicate that precipitation during this period was actually higher than during the Younger Dryas (YD) due to increased influx of isotopically-heavy moisture from the Gulf of Mexico, and that even in Arizona the BA was wetter than is typically assumed. However, we note that the variability of growth rates is extremely low, ranging between 0 and 10 mm/yr in the Fort Stanton and Cave of the Bells speleothems, much less than what is seen in the Texas speleothems (up to ~ 90 mm/yr). A study based on the growth histories of six stalagmites from several southern New Mexico caves does not support a wet BA, indicating higher growth rates and thus wetter conditions during the YD and slow/no growth during the BA (Poyak et al., 2004). In northern Nevada, the Goshute Cave (Fig. 1) δ18O record appears to respond strongly to North Atlantic events, specifically D-O events 23–21, recording increases in δ18O values during North Atlantic warming that are interpreted as a change in temperature and perhaps moisture source (Denniston et al., 2007). Our record fits in this framework, appearing to capture the peak of D-O event 21 and cooling afterwards. Why Leviathan Cave δ18O (Lachniet et al., 2014) appears to be insensitive to D-O events and captures the Younger Dryas stadial is unclear, as is why our Lehman Caves δ18O record and the Goshute Cave δ18O record respond to D-O events but not Heinrich Stadials. It is unlikely that recharge or evaporative effects may play a role in the elevated Leviathan δ18O values relative to Lehman and Goshute Caves signals (Fig. 6) during MIS 5c to 5a, as modern Leviathan Cave

drip waters fall along the meteoric water line. Meridional temperature gradients may have been greater than estimated by the Leviathan Cave drip water correction due to the presence of a small Laurentide ice sheet, but this only explains an offset, not the lack of response to D-O events. This lack of regional reproducibility is problematic, and additional Great Basin speleothem studies are needed to address these discrepancies. Further development of complementary proxies like trace elements and δ13C in Great Basin speleothems will promote understanding of the local hydrologic and environmental contexts of speleothem growth, perhaps shedding light on these discrepancies. Future studies will continue to build a history of effective precipitation in the Great Basin, particularly in times beyond the reach of radiocarbon dating, increasing understanding of the controls on and global context of moisture balance in this sensitive region.

5. Conclusions

Past Great Basin speleothem records have used δ18O to investigate large-scale atmospheric circulation changes associated with abrupt climate events such as glacial terminations and Dansgaard-Oeschger events (Denniston et al., 2007; Lachniet et al., 2014; Shakun et al., 2011). We replicated previous Great Basin δ18O orbitally consistent chronologies of T-II using speleothems from Lehman Caves and high-resolution dating techniques, and provided a novel and complementary record of trace element variations supported by δ13C analyses. Our trace element and δ13C records beginning at 139 ka show that wet conditions at Lehman Caves persisted throughout H11 until ~130–128 ka when abrupt drying occurred, appearing to coincide with end of H11 and driving Mg/Ca and δ13C to significantly higher values (from ~ 4–10 mmol/mol and roughly ~8 to 0‰) via greater prior calcite precipitation. Strengthening of the Aleutian Low during Heinrich Stadials enhanced moisture delivery to the region via the winter storm track, which was south-shifted to a mean position over the Great Basin and intensified during glacial conditions (Chiang et al., 2014; Okumura et al., 2009; Oster et al., 2015). Teleconnections to North Atlantic climate are further demonstrated by δ18O variations between ~84 and 81 ka, which correlate with the peak of D-O event 21 and subsequent cooling, consistent with most other regional speleothem responses to these or similar events (Asmerom et al., 2010; Denniston et al., 2007; Wagner et al., 2010).

Acknowledgments

We thank B. Roberts and G.M. Baker of the National Park Service at GBNP. Without their assistance, this project would not have been possible. We thank A. Neary, I. Tal, and E. Steponaitis, who performed large numbers of trace element measurements at MIT, and J. Retrum, who maintained the U-Th dating. We also thank the three anonymous reviewers whose comments greatly improved the manuscript. This study was supported by funding from the National Science Foundation via Grants 1103379 to D. McGee, 1103320 to R.L. Edwards, and EAR-1103066 to J. Quade. Parts of this work were carried out in the Characterization Facility, University of Minnesota, which receives partial support from the NSF through the MRSEC program.

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Please cite this article in press as: Cross, M., et al., Great Basin hydrology, paleoclimate, and connections with the North Atlantic: A speleothem stable isotope and trace element record from Lehman Caves, NV. Quaternary Science Reviews (2015), http://dx.doi.org/10.1016/ j.quascirev.2015.06.016