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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Abstract

We present a record of speleothem stable isotope ($\delta^{18}\text{O}$, $\delta^{13}\text{C}$) 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 $\delta^{18}\text{O}$ 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 $\delta^{18}\text{O}$ 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 CO_2 . We infer that this shift results from temperature increase due to increasing atmospheric CO_2 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 $\delta^{13}\text{C}$ values are largely decoupled from $\delta^{18}\text{O}$ 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.

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 millennial-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 rises are associated with Heinrich Stadial (HS) 1 (Munroe and Laabs, 2013a; Oviatt, 1997) and the Younger Dryas stadial (Oviatt, 1997) and increased freshwater inputs to Lake Bonneville occur during H2 (McGee 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

69 delivery to the region (Okumura et al., 2009), or via an ice-sheet modulated jet stream position (Anderson
70 et al., 1988). Speleothems from the southwestern United States also show clear responses to Heinrich
71 stadials (Asmerom et al., 2010; Musgrove et al., 2001; Wagner et al., 2010).

72 However, the connection between Great Basin and North Atlantic climate is not as clear prior to the last
73 glacial cycle. The first absolute dated, high-resolution record of T-II in the Great Basin was the Devils
74 Hole vein calcite $\delta^{18}\text{O}$ record (Ludwig et al., 1992; Winograd et al., 1992). Recording a 2‰ increase in
75 $\delta^{18}\text{O}$ that seemed to show the penultimate deglaciation greater than 10 kyrs before a rise in Northern
76 Hemisphere summer insolation and H11, this record sparked years of debate on the controls of the Devils
77 Hole isotopic record and challenged the conventional understanding of the mechanisms by which orbital
78 forcing modulates global climate and glacial cycles (Herbert et al., 2001; Imbrie et al., 1993; Winograd,
79 2002). To test the possibility of an early T-II elsewhere in the Great Basin, Shakun et al. (2011)
80 developed a speleothem record from Lehman Caves (Figure 1) that showed a 2‰ rise in $\delta^{18}\text{O}$ values
81 between 133.5 ± 0.7 and 128.9 ± 0.4 ka, consistent with global marine and terrestrial records of T-II and
82 orbital forcing as shown in Figure 2 (Cheng et al., 2009; Drysdale et al., 2005; Imbrie et al., 1984;
83 Kawamura et al., 2007). Subsequent speleothem work supported and extended the Shakun et al. (2011) T-
84 II record (Lachniet et al., 2014).

85 In this study, we test for replication of this key oxygen isotope record immediately before, during, and
86 subsequent to T-II. Moreover, we obtained trace metal (Mg/Ca and Sr/Ca) ratios as a means of
87 characterizing local to regional-scale effective precipitation (Figure 2). Trace metals have provided
88 valuable hydrological insights in speleothems from diverse locations (Cruz et al., 2007; Orland et al.,
89 2014; Roberts et al., 1998; Tan et al., 2014), and in modern Lehman Caves conditions appear to be a
90 measure of aridity, reflecting the amount of prior calcite precipitation (PCP) occurring in the epikarst
91 above the cave (Steponaitis et al., in press). By combining oxygen and carbon isotope analysis with trace
92 metal analysis, we can more fully characterize climatic conditions in the region before, during, and after
93 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 (Figure 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 $\delta^{18}\text{O}$ values are similar to the volume-weighted mean annual precipitation $\delta^{18}\text{O}$ values of ~ -15.5‰ (their Figure 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 Glancy, 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., in press).

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 (Figure 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 ²³⁸U (3.20 ng ²³⁸U 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 ²³⁰Th/²³²Th 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 ²³²Th/²³⁸U 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 ± 6 ag ²³⁴U, 0.9 ± 0.2 pg ²³⁸U, 26 ± 9 ag ²³⁰Th, and 0.3 ± 0.1 pg ²³²Th. 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 $0.01 \text{ mm}^{-1} \pm 2$ 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 $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values are approximately $\pm 0.22\text{‰}$ and $\pm 0.16\text{‰}$, 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 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 HNO_3 , diluted to $\sim 0.05\%$ Ca in 0.5 M HNO_3 , and analyzed for ^{25}Mg , ^{26}Mg , ^{43}Ca , ^{86}Sr , and ^{87}Sr on a Thermo VG PQ2+ quadrupole 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 seconds/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: $139.0 \pm 1.0 - 129.2 \pm 1.1$ ka, $123.1 \pm 0.2 - 122.8 \pm 0.2$ ka, $84.3 \pm 0.2 - 84.0 \pm 0.3$ ka, and $82.4 \pm 0.3 - 81.3 \pm 0.3$ ka (Figure 3, Figure 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 ^{230}Th (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 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 $^{230}\text{Th}/^{232}\text{Th}$ 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 Figure 3 as the large

dating pit near the bottom 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 ($131.5 \pm 0.8 - 129.2 \pm 1.1$ ka) and WR-41 ($139.0 \pm 1.0 - 129.8 \pm 0.6$ ka), and in LC-2 (133 – 128 ka). Growth rates are low during this time interval, varying between 0.9 and 33.3 $\mu\text{m/yr}$ with an average of 19.3 $\mu\text{m/yr}$ in WR-41, and in IR-3 ranging from 0.2 to 14.0 $\mu\text{m/yr}$, with an average of 10.3 $\mu\text{m/yr}$. 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 $\delta^{234}\text{U}_{\text{initial}}$ begins around 131 – 130 ka, rising from respective baselines around 250‰ and 200‰ to peak at 433.5‰ and 387.2‰. In IR3, $\delta^{234}\text{U}_{\text{initial}}$ increases, from 391.3‰ to a brief peak at 551.6‰ immediately before the end of the growth phase, which has a $\delta^{234}\text{U}_{\text{initial}}$ 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 $\mu\text{m/yr}$ and 768 ppb. The $\delta^{234}\text{U}_{\text{initial}}$ 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 $\mu\text{m/yr}$) and high $\delta^{234}\text{U}_{\text{initial}}$ 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 Figure 5. In WR-41, $\delta^{18}\text{O}$ 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 $\delta^{13}\text{C}$ values is approximately 8‰. In WR-41, $\delta^{13}\text{C}$ 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 $\delta^{13}\text{C}$ values from -6.0‰ to -0.3‰ occurs within error at the same time as observed in WR-41.

At 123 ka, $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values are highly variable, ranging from -14.5‰ to -11.3‰ and from -7.4‰ to -3.3‰, respectively. At 84 ka $\delta^{18}\text{O}$ values average -12.1‰, distinctly heavier than $\delta^{18}\text{O}$ values in the younger phase (82 – 81 ka), which have a mean of -13.5‰. $\delta^{13}\text{C}$ values are more similar, and range between -3.4 to -6.5‰. The stable isotope trends in LC-2 are described in Shakun et al. (2011). $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ 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 $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ and were not significantly correlated ($r^2 = 0.31$) with each other or with distance from the growth axis ($\delta^{13}\text{C}$ -distance $r^2 = 0.22$, $\delta^{18}\text{O}$ -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 $\delta^{18}\text{O}$, $\delta^{13}\text{C}$, and trace elements (discussed below) of our new speleothems 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 (Figure 5) is characterized by an abrupt increase around 130 ka from ratios ranging between 2.1 to 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 $\mu\text{mol/mol}$, 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 $\mu\text{mol/mol}$). Sr/Ca in IR-3 positively trend over a range from 24.7 to 36.1 $\mu\text{mol/mol}$.

4. Discussion

4.1 Interpretation of $\delta^{18}\text{O}$

Our Lehman Caves $\delta^{18}\text{O}$ record replicates the 2‰ rise in speleothem $\delta^{18}\text{O}$ 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 $\delta^{18}\text{O}$ 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 – 0.35‰/°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 $\delta^{18}\text{O}$ 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°C lower than modern MAT (Hostetler and Benson, 1990; Kaufman, 2003; Quade et al., 2003); we note that MIS 5e was globally ~ 2°C warmer than today (Rohling et al., 2008). Our $\delta^{18}\text{O}$ 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 Figure 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 $\delta^{18}\text{O}$ 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 $\delta^{18}\text{O}$. Relevant to this discussion, a recent simulation of the impact of precessional changes on precipitation and precipitation $\delta^{18}\text{O}$ using an isotope-enabled atmospheric general circulation model found a *decrease* of $>1\text{‰}$ in the annual-mean $\delta^{18}\text{O}$ 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, Figure 7). As this insolation difference is only $\sim 25\%$ greater than that occurring across T-II, the model results suggest that a similar drop in precipitation-weighted $\delta^{18}\text{O}$ may have occurred during T-II due to increasing boreal summer insolation. Such a drop in precipitation $\delta^{18}\text{O}$ appears hard to reconcile with speleothem $\delta^{18}\text{O}$ changes during T-II, requiring an unrealistically large rise in cave temperature to generate the observed rise in speleothem $\delta^{18}\text{O}$ if precipitation $\delta^{18}\text{O}$ was

simultaneously falling. As the model does not include ice volume changes, the results do not rule out deglacial changes in Great Basin $\delta^{18}\text{O}$ related to ice sheet effects on precipitation seasonality.

As for the youngest parts of our record, ~ 84 ka and $\sim 82 - 81$ ka, the older interval appears to capture the later part of the increase in $\delta^{18}\text{O}$ values at the end of the Goshute Cave (Figure 1) $\delta^{18}\text{O}$ record associated with Dansgaard-Oeschger (D-O) event 21 (Denniston et al., 2007), as shown in Figure 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 (Figure 1). Applying the Leviathan Cave dripwater correction (Lachniet et al., 2014) to both Lehman and Goshute Caves' $\delta^{18}\text{O}$ values (Figure 6, panel 3), neither curve is brought completely into line with the Leviathan Cave record. Lehman and Goshute Caves' $\delta^{18}\text{O}$ values are 1 – 5 ‰ lighter, except for the Lehman Caves $\delta^{18}\text{O}$ 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 $\delta^{18}\text{O}$ values decrease and Leviathan Cave values rise. We observe that Devils Hole $\delta^{18}\text{O}$ values (Winograd et al., 1992) are similar to Goshute Cave $\delta^{18}\text{O}$ values around this time, around -16‰, though Devils Hole is at a lower latitude than Leviathan Cave (Figure 1). This is probably due to the relatively higher temperatures ($\sim 34^\circ\text{C}$) of calcite precipitation at Devils Hole (Coplen et al. 2007).

4.2 Interpretation of Trace Elements and $\delta^{13}\text{C}$

Mg/Ca and Sr/Ca ratios have often provided a perspective on hydrological and environmental conditions that is complementary to $\delta^{18}\text{O}$ 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. (in press) 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 $\delta^{13}\text{C}$ can lead to a more conclusive diagnosis of the likely drivers of both proxies. Controls on $\delta^{13}\text{C}$ values include hydrologic and biologic factors. One potential mechanism is PCP. During calcite precipitation, degassing of CO_2 preferentially releases ^{12}C , leading to $\delta^{13}\text{C}$ enrichment (Johnson et al., 2006). Longer water residence times or changes in the mode (open vs. closed) of host rock ($\delta^{13}\text{C}$ value of $\sim 0\text{‰}$) dissolution may also lead to enriched $\delta^{13}\text{C}$ values (Genty et al., 2001; Hendy, 1971). $\delta^{13}\text{C}$ 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 CO_2 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 ($\sim 10\text{‰}$ compared to $\sim 8\text{‰}$), the dramatic shift in vegetation density required is highly unlikely to have ever occurred here. Changes in vegetation type (C_3 vs C_4) (Dorale et al., 1992) are also unlikely. Pollen and midden evidence indicates that C_3 plants have dominated this region for at least the last 40 kyrs (Madsen et al., 2001). Although we cannot completely rule out contributions to our $\delta^{13}\text{C}$ 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 $\delta^{13}\text{C}$ variations in our record are driven by changes in the amount of CO_2 degassing and calcite precipitation.

Enhanced prior calcite precipitation may be driven by CO_2 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 shows high covariability, indicating that PCP is likely the dominant control on modern dripwater trace metals (Steponaitis et al., in press).

In Figures 7 and 8, we examine the whole-rock covariations in WR41, IR-3, and LC-2 among $\delta^{13}\text{C}$, Mg/Ca, and Sr/Ca. These are generally poor ($r^2 < 0.4$), 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 ($r^2 = 0.45$), 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 Figure 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 $\delta^{13}\text{C}$ relationships. In the older period, Mg/Ca covaries with neither $\delta^{13}\text{C}$ nor Sr/Ca. In stalagmites spanning the end of the last deglaciation and the early Holocene from Lehman Caves, Steponaitis et al. (in press) 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 ($r^2 = 0.50$, $p = 0.015$) and $\delta^{13}\text{C}$ ($r^2 = 0.81$, $p = 0.0002$) between 131 and 129 ka, similar to what we see in IR-3 ($\sim 131.5 - 128.8$ ka). This is diagnostic of PCP control on groundwater trace element and carbon isotope compositions. At this time the LC-2 $\delta^{13}\text{C}$ (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 $\delta^{13}\text{C}$ 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 $\delta^{13}\text{C}$ values, Mg/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 $\delta^{13}\text{C}$ 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 $\delta^{234}\text{U}_{\text{initial}}$ values. As shown in Figure 5, $\delta^{234}\text{U}_{\text{initial}}$ rises sharply, coincident with our increases in trace metal ratios and $\delta^{13}\text{C}$ values between 131 and 129 ka. Speleothem $\delta^{234}\text{U}_{\text{initial}}$ reflects the $^{234}\text{U}/^{238}\text{U}$ disequilibrium of the water from which it formed (Hellstrom and McCulloch, 2000). This disequilibrium arises from the ejection of ^{234}U into solution via alpha recoil, or from the preferential leaching of ^{234}U from sites damaged by α -decay (Fleischer and Raabe, 1978; Ivanovich and Harmon, 1982; Osmond, 1980). High ^{234}U may be due to low infiltration rates, or dry conditions with infrequent wetting events, allowing ^{234}U 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 $\delta^{18}\text{O}$ values between 134 and 129 ka clearly correlates with the global atmospheric CO_2 increase from ~ 180 to 280 ppmv recorded in the Vostok ice core (Petit et al., 1999), with increasing

boreal summer insolation (Figure 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 $\delta^{18}\text{O}$ 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 $\delta^{18}\text{O}$ values.

However, the abrupt decrease in effective precipitation inferred from trace element data, $\delta^{13}\text{C}$ and $\delta^{234}\text{U}$ is essentially decoupled from the more gradual increase in temperature (and possible change in precipitation seasonality) signaled by $\delta^{18}\text{O}$. The effective precipitation at Lehman Caves remained high until $\sim 130 - 128$ ka, as demonstrated by the small variability of $\delta^{13}\text{C}$ 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 Figure 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 $\delta^{18}\text{O}$ records from Hulu Cave stalagmites and $\delta^{18}\text{O}$ 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., 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

405 suppression, lead to reduced winter and summer precipitation in the western United States. Enhanced
406 effective precipitation during Heinrich stadials is consistent with widespread reports of Great Basin lake
407 highstands clustering around H1 (Munroe and Laabs, 2013a). The decoupling of $\delta^{18}\text{O}$ from enhanced
408 effective precipitation suggests that winter moisture source may not have a significant role in the $\delta^{18}\text{O}$
409 signal; instead we favor temperature and perhaps a changing seasonality of precipitation as the
410 termination progressed.

411 Other Great Basin paleoclimate records suggest a similar decoupling between warming and hydrological
412 changes during the last deglaciation. For instance, glaciers in the American West exhibited substantial
413 retreat between the LGM and Bølling-Allerød (BA) (Shakun et al., accepted, Nature Comm), implying
414 considerable summer warming across the H1 interval. The Leviathan Cave $\delta^{18}\text{O}$ record also exhibits a
415 general 5‰ enrichment from H1 to the early Holocene (Lachniet et al., 2014), reminiscent of the gradual
416 Lehman Cave $\delta^{18}\text{O}$ shift across T-II. In contrast, Great Basin pluvial lakes tended to reach highstands
417 during H1 (Munroe and Laabs, 2013a) followed by rapid regressions after the event (Munroe and Laabs,
418 2013b). These T-I patterns thus mirror the sequence we have identified over T-II, with an abrupt drying
419 following a Heinrich event and commencing several millennia after the onset of a general warming trend.

420 Speleothem records from the Great Basin and southwestern US also further illustrate connections between
421 western North America and North Atlantic climate and raise new questions. Wetter winter conditions are
422 recorded in a Cave of the Bells, Arizona speleothem during stadials (Wagner et al., 2010). Another
423 speleothem record from Fort Stanton Cave, New Mexico, clearly captures the Younger Dryas stadial and
424 the last five Heinrich stadials, recording increased contribution of winter precipitation during stadials. In
425 modern conditions, higher winter precipitation sourced from the Pacific is associated with higher mean
426 annual rainfall (Asmerom et al., 2010). In contrast, Feng et al. (2014) argue that the higher growth rates
427 seen in the Fort Stanton record and their Texas speleothems during the warm BA indicate that
428 precipitation during this period was actually higher than during the Younger Dryas (YD) due to increased
429 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 $\mu\text{m}/\text{yr}$ in the Fort Stanton and Cave of the Bells speleothems, much less than what is seen in the Texas speleothems (up to $\sim 90 \mu\text{m}/\text{yr}$). 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. In northern Nevada, the Goshute Cave (Figure 1) $\delta^{18}\text{O}$ record appears to respond strongly to North Atlantic events, specifically D-O events 23 – 21, recording increases in $\delta^{18}\text{O}$ 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 $\delta^{18}\text{O}$ (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 $\delta^{18}\text{O}$ record and the Goshute Cave $\delta^{18}\text{O}$ 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 $\delta^{18}\text{O}$ values relative to Lehman and Goshute Caves signals (Figure 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 $\delta^{13}\text{C}$ 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 $\delta^{18}\text{O}$ 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 $\delta^{18}\text{O}$ 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 $\delta^{13}\text{C}$ analyses. Our trace element and $\delta^{13}\text{C}$ records beginning at 139 ka show that wet conditions at Lehman Caves persisted throughout H11 until $\sim 130 - 128$ ka when abrupt drying occurred, appearing to coincide with end of H11 and driving Mg/Ca and $\delta^{13}\text{C}$ to significantly higher values (from ~ 4 to 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 $\delta^{18}\text{O}$ 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).

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479 Figure captions

480 Figure 1: A map of the Great Basin indicating the locations of Lehman Caves, Goshute Cave, Leviathan
481 Cave, and Devils Hole. The color scale indicates the winter (December, January and February) and
482 summer (June, July, and August) contributions to mean annual precipitation (NOAA Climate Prediction
483 Center). Much of the modern Great Basin receives an approximately equal distribution of annual
484 precipitation between winter and summer, but studies of cave and spring waters and groundwaters
485 indicate that the majority of recharge is from winter precipitation (Paul and Thodal, 2014; Winograd et al.
486 1988).

487 Figure 2: The rising limb of boreal summer insolation (gray) and CO₂ (bottom, black) (Petit et al., 1999)
488 is broadly paralleled by Lehman Caves $\delta^{18}\text{O}$ in red (Shakun et al., 2011), purple (Lachniet et al., 2014),
489 and blue (this study). Great Basin cave records appear to be largely decoupled from HS 11 (green bar),
490 which is associated with Weak Monsoon Interval II (dry conditions) in Chinese speleothems (top, green
491 and black) (Cheng et al., 2009; Wang et al., 2008). Younger Heinrich stadials are associated with lake
492 highstands in the Great Basin (McGee et al., 2012; Munroe and Laabs, 2013a) and increased contributions
493 of winter precipitation in the American southwest (Asmerom et al., 2010; Wagner et al., 2010).

494 Figure 3: Age models and high-resolution scan of WR-41. The depth scale of the plot is calibrated to
495 match the image. Growth phase age models calculated from dates shown in green are represented by
496 black lines with purple error envelopes. The slopes of our age models are determined by the model
497 rigidity set by OxCal using a small k parameter of $0.01 \text{ mm}^{-1} \pm$ two order of magnitude (increasing k
498 increases rigidity). Dates indicated in orange were not used in age models. Data from this stalagmite,
499 WR-41, is distinguished throughout this paper by green coloration and/or circular markers.

500 Figure 4: High resolution scan and age model of IR-3. The depth scale of the plot is calibrated to match
501 the image. The age model for the large main growth phase is constructed similarly to those of WR-41
502 (Figure 3). Dates not used in the model are significantly older than those shown in blue and are thus not
503 included on the plot. Data from IR-3 is distinguished throughout this paper by blue coloration and/or
504 triangular markers.

505 Figure 5: Results from stable isotope ($\delta^{18}\text{O}$, $\delta^{13}\text{C}$) and trace element (Mg/Ca, Sr/Ca) analysis and U-Th
506 dating (ages and $\delta^{234}\text{U}_{\text{initial}}$) of IR-3 (blue triangles), WR-41 (green circles), and LC-2 (red squares). Stable
507 isotope data and U-Th dating results from LC-2 were originally published in Shakun et al. (2011). LC-2
508 data is distinguished throughout this paper by red coloration and/or square markers. Age control points
509 from WR-41 and IR-3 are shown at the bottom of the panel.

510 Figure 6: Panel 1: Comparison of Lehman Caves records with regional and global records (Andersen et
511 al., 2004; Cheng et al., 2009; Coplen et al., 1994; Denniston et al., 2007; Lachniet et al., 2014; Petit et al.,
512 1999; Winograd et al., 1992; Wang et al., 2008). Panel 2: Zoomed-in view of dashed box in panel 1,
513 showing the remarkable antiphasing between the Asian monsoon record (Cheng et al., 2009; Wang et al.,
514 2008) and the Lehman Caves Mg/Ca record. Panel 3: Lehman Caves (blue), Leviathan Cave (purple), and
515 Goshute Cave (green) speleothem records from dotted box in panel 1, plotted on same axis. In both panel
516 1 and panel 3, gray curves indicate original data. In panel 1, Goshute and Lehman Caves records are sea
517 water corrected only; in panel 3 the colored curves are additionally corrected to Leviathan Cave drip

water (Lachniet et al., 2014). This procedure does not bring Lehman and Goshute Caves $\delta^{18}\text{O}$ values completely into line with Leviathan Cave $\delta^{18}\text{O}$ values; Devils Hole $\delta^{18}\text{O}$ values are also relatively low at this time (panel 1). Our correction factors are +0.32 and +1.03‰ for Lehman and Goshute Caves.

Figure 7: Trace element data from IR-3 (triangles), LC-2 (squares), and WR-41 (circles) plotted in Sr/Ca vs. Mg/Ca space and colored according to age. The darkest red color is included for samples from WR-41 dating to ~ 123 ka. The blue, red, and green lines are whole stalagmite trend lines of Mg/Ca and Sr/Ca in IR-3, LC-2, and WR-41, respectively; also shown are their r^2 values ($p < 0.005$).

Figure 8: Mg/Ca (left) and Sr/Ca (right) against $\delta^{13}\text{C}$. Color and symbology are identical to that of Figure 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.

Figure 9: Correlations between Mg/Ca and Sr/Ca, Mg/Ca and $\delta^{13}\text{C}$ in sample WR-41 for intervals around the beginning (blue) end (orange) of Heinrich Stadial 11. Mg/Ca and Sr/Ca (upper panels) are significantly correlated between 129 and 131 ka ($r^2 = 0.50$, $p = 0.015$) but not between 135 and 139 ka ($r^2 = 0.02$, $p = 0.435$). Mg/Ca and Sr/Ca (upper panels) are significantly correlated between 129 and 131 ka ($r^2 = 0.50$, $p = 0.015$) but not between 135 and 139 ka ($r^2 = 0.02$, $p = 0.435$). Mg/Ca and $\delta^{13}\text{C}$ (lower panels) are similarly related, significantly and strongly correlated in the younger period of interest ($r^2 = 0.81$, $p = 0.0002$) but not in the older interval ($r^2 = 0.07$, $p = 0.143$). We suggest the high correlations between Mg/Ca and $\delta^{13}\text{C}$ and Mg/Ca and Sr/Ca during the younger period indicate increasing prior calcite precipitation, probably due to drying. Data point color in the two right panels is on the same scale used in Figures 7 and 8. Data from WR-41 not inside the orange or blue boxes are shown as hollow gray circles.

539 Bibliography

- 540 Andersen, K. K., Azuma, N., Barnola, J.-M., Bigler, M., Biscaye, P., Caillon, N., Chappellaz, J.,
541 Clausen, H. B., Dahl-Jensen, D., Fischer, H., et al. (2004). High-resolution record of Northern
542 Hemisphere climate extending into the last interglacial period. *Nature*, 431(7005):147–151.
- 543 Anderson, P., Barnosky, C., Bartlein, P., Behling, P., Brubaker, L., Cushing, E., Dodson, J., Dworetsky,
544 B., Guetter, P., Harrison, S., et al. (1988). Climatic changes of the last 18,000 years:
545 Observations and model simulations. *Science*, 241:1043–1052.
- 546 Asmerom, Y., Polyak, V. J., and Burns, S. J. (2010). Variable winter moisture in the southwestern
547 United States linked to rapid glacial climate shifts. *Nature Geoscience*, 3(2):114–117.
- 548 Ayliffe, L. and Veer, H. (1988). Uranium-series dating of speleothems and bones from Victoria Cave,
549 Naracoorte, South Australia. *Chemical Geology: Isotope Geoscience Section*, 72(3):211–234.
- 550 Baker, A., Ito, E., Smart, P. L., and McEwan, R. F. (1997). Elevated and variable values of ^{13}C in
551 speleothems in a British cave system. *Chemical Geology*, 136(3):263–270.
- 552 Baldini, J., McDermott, F., Baker, A., Baldini, L., Matthey, D., and Railsback, L. B. (2005). Biomass
553 effects on stalagmite growth and isotope ratios: A 20th century analogue from Wiltshire,
554 England. *Earth and Planetary Science Letters*, 240(2):486–494.
- 555 Battisti, D., Ding, Q., and Roe, G. (2014). Coherent pan-Asian climatic and isotopic response to orbital
556 forcing of tropical insolation. *Journal of Geophysical Research: Atmospheres*, 119(21):11–997.
- 557 Benson, L. and Klieforth, H. (1989). Stable isotopes in precipitation and ground water in the Yucca
558 Mountain region, southern Nevada: Paleoclimatic implications. *Aspects of Climate Variability
559 in the Pacific and the western Americas*, pages 41–59.
- 560 Broecker, W. S. and Orr, P. C. (1958). Radiocarbon chronology of Lake Lahontan and Lake Bonneville.
561 *Geological Society of America Bulletin*, 69(8):1009–1032.
- 562 Bromwich, D. H., Toracinta, E. R., Wei, H., Oglesby, R. J., Fastook, J. L., and Hughes, T. J. (2004).
563 Polar MM5 Simulations of the Winter Climate of the Laurentide Ice Sheet at the LGM*.
564 *Journal of Climate*, 17(17):3415–3433.
- 565 Cheng, H., Edwards, R. L., Broecker, W. S., Denton, G. H., Kong, X., Wang, Y., Zhang, R., and Wang,
566 X. (2009). Ice age terminations. *Science*, 326(5950):248–252.
- 567 Cheng, H., Edwards, R. L., Shen, C.-C., Polyak, V. J., Asmerom, Y., Woodhead, J., Hellstrom, J.,
568 Wang, Y., Kong, X., Spötl, C., et al. (2013). Improvements in ^{230}Th dating, ^{230}Th and ^{234}U half-
569 life values, and U–Th isotopic measurements by multi-collector inductively coupled plasma
570 mass spectrometry. *Earth and Planetary Science Letters*, 371:82–91.
- 571 Chiang, J. C., Lee, S.-Y., Putnam, A. E., and Wang, X. (2014). South Pacific Split Jet, ITCZ shifts, and
572 atmospheric North–South linkages during abrupt climate changes of the last glacial period.
573 *Earth and Planetary Science Letters*, 406:233–246.

574 Coplen, T. B. (2007). Calibration of the calcite–water oxygen-isotope geothermometer at Devils Hole,
575 Nevada, a natural laboratory. *Geochimica et Cosmochimica Acta*, 71(16):3948–3957.

576 Coplen, T. B., Winograd, I. J., Landwehr, J. M., and Riggs, A. C. (1994). 500,000-year stable carbon
577 isotopic record from Devils Hole, Nevada. *Science*, 263(5145):361–365.

578 Cruz, F. W., Burns, S. J., Jercinovic, M., Karmann, I., Sharp, W. D., and Vuille, M. (2007). Evidence of
579 rainfall variations in Southern Brazil from trace element ratios (Mg/Ca and Sr/Ca) in a Late
580 Pleistocene stalagmite. *Geochimica et Cosmochimica Acta*, 71(9):2250–2263.

581 Dansgaard, W. (1964). Stable isotopes in precipitation. *Tellus*, 16(4):436–468.

582 Day, C. C. and Henderson, G. M. (2013). Controls on trace-element partitioning in cave-analogue
583 calcite. *Geochimica et Cosmochimica Acta*, 120:612–627.

584 Denniston, R. F., Asmerom, Y., Polyak, V., Dorale, J. A., Carpenter, S. J., Trodick, C., Hoye, B., and
585 González, L. A. (2007). Synchronous millennial-scale climatic changes in the Great Basin and
586 the North Atlantic during the last interglacial. *Geology*, 35(7):619–622.

587 Denton, G., Anderson, R., Toggweiler, J., Edwards, R., Schaefer, J., and Putnam, A. (2010). The last
588 glacial termination. *Science*, 328(5986):1652–1656.

589 Dorale, J. A. and Liu, Z. (2009). Limitations of Hendy test criteria in judging the paleoclimatic
590 suitability of speleothems and the need for replication. *Journal of Cave and Karst Studies*,
591 71(1):73–80.

592 Dorale, J. A., González, L. A., Reagan, M. K., Pickett, D. A., Murrell, M. T., and Baker, R. G. (1992).
593 A high-resolution record of Holocene climate change in speleothem calcite from Cold Water
594 Cave, northeast Iowa. *Science*, 258(5088):1626–1630.

595 Dorale, J. A., Edwards, R. L., Ito, E., and González, L. A. (1998). Climate and vegetation history of the
596 midcontinent from 75 to 25 ka: a speleothem record from Crevice Cave, Missouri, USA.
597 *Science*, 282(5395):1871–1874.

598 Dorale, J. A., Edwards, R. L., Alexander Jr, E. C., Shen, C.-C., Richards, D. A., and Cheng, H.
599 Uranium-series dating of speleothems: current techniques, limits, & applications. In *Studies of*
600 *Cave Sediments*, pages 177–197. *Springer*, (2004).

601 Douglas, M. W., Maddox, R. A., Howard, K., and Reyes, S. (1993). The Mexican monsoon. *Journal of*
602 *Climate*, 6(8):1665–1677.

603 Drewes, H. and Palmer, A. R. (1957). Cambrian rocks of southern Snake Range, Nevada. *AAPG*
604 *Bulletin*, 41(1):104–120.

605 Drysdale, R. N., Zanchetta, G., Hellstrom, J. C., Fallick, A. E., and Zhao, J.-x. (2005). Stalagmite
606 evidence for the onset of the Last Interglacial in southern Europe at 129±1 ka. *Geophysical*
607 *Research Letters*, 32(24).

608 Eardley, A., Shuey, R., Gvosdetsky, V., Nash, W., Picard, M. D., Grey, D., and Kukla, G. (1973). Lake
609 cycles in the Bonneville basin, Utah. *Geological Society of America Bulletin*, 84(1):211–216.

610 Edwards, R. L., Chen, J., and Wasserburg, G. (1987). ^{238}U - ^{234}U - ^{230}Th - ^{232}Th systematics and the precise
611 measurement of time over the past 500,000 years. *Earth and Planetary Science Letters*,
612 81(2):175–192.

613 Enfield, D. B., Mestas-Nuñez, A. M., and Trimble, P. J. (2001). The Atlantic multidecadal oscillation
614 and its relation to rainfall and river flows in the continental US. *Geophysical Research Letters*,
615 28(10):2077–2080.

616 Fairchild, I. J. and Treble, P. C. (2009). Trace elements in speleothems as recorders of environmental
617 change. *Quaternary Science Reviews*, 28(5):449–468.

618 Fairchild, I. J., Borsato, A., Tooth, A. F., Frisia, S., Hawkesworth, C. J., Huang, Y., McDermott, F., and
619 Spiro, B. (2000). Controls on trace element (Sr–Mg) compositions of carbonate cave waters:
620 implications for speleothem climatic records. *Chemical Geology*, 166(3):255–269.

621 Feng, W., Hardt, B. F., Banner, J. L., Meyer, K. J., James, E. W., Musgrove, M., Edwards, R. L., Cheng,
622 H., and Min, A. (2014). Changing amounts and sources of moisture in the US southwest since
623 the Last Glacial Maximum in response to global climate change. *Earth and Planetary Science
624 Letters*, 401:47–56.

625 Fleischer, R. L. and Raabe, O. (1978). Recoiling alpha-emitting nuclei. Mechanisms for uranium-series
626 disequilibrium. *Geochimica et Cosmochimica Acta*, 42(7):973–978.

627 Friedman, I., Harris, J. M., Smith, G. I., and Johnson, C. A. (2002)a. Stable isotope composition of
628 waters in the Great Basin, United States 1. Air-mass trajectories. *Journal of Geophysical
629 Research: Atmospheres* (1984–2012), 107(D19):ACL–14.

630 Friedman, I., Smith, G. I., Johnson, C. A., and Moscati, R. J. (2002)b. Stable isotope compositions of
631 waters in the Great Basin, United States 2. Modern precipitation. *Journal of Geophysical
632 Research: Atmospheres* (1984–2012), 107(D19):ACL–15.

633 Genty, D., Baker, A., Massault, M., Proctor, C., Gilmour, M., Pons Branchu, E., and Hamelin, B.
634 (2001). Dead carbon in stalagmites: carbonate bedrock paleodissolution vs. ageing of soil
635 organic matter. Implications for ^{13}C variations in speleothems. *Geochimica et Cosmochimica
636 Acta*, 65(20):3443–3457.

637 Gilbert, G. K. (1890). Lake Bonneville, volume 1. United States Geological Survey.

638 Hellstrom, J. and McCulloch, M. (2000). Multi-proxy constraints on the climatic significance of trace
639 element records from a New Zealand speleothem. *Earth and Planetary Science Letters*,
640 179(2):287–297.

641 Hellstrom, J., McCulloch, M., and Stone, J. (1998). A detailed 31,000 year record of climate and
642 vegetation change, from the isotope geochemistry of two New Zealand speleothems.
643 *Quaternary Research*, 50(2):167–178.

644 Hendy, C. H. (1971). The isotopic geochemistry of speleothems—I. The calculation of the effects of
645 different modes of formation on the isotopic composition of speleothems and their applicability
646 as palaeoclimatic indicators. *Geochimica et cosmochimica Acta*, 35(8):801–824.

647 Herbert, T., Schuffert, J., Andreasen, D., Heusser, L., Lyle, M., Mix, A., Ravelo, A., Stott, L., and
648 Herguera, J. (2001). Collapse of the California Current during glacial maxima linked to climate
649 change on land. *Science*, 293(5527):71–76.

650 Higgins, R., Yao, Y., and Wang, X. (1997). Influence of the North American monsoon system on the
651 US summer precipitation regime. *Journal of Climate*, 10(10):2600–2622.

652 Hostetler, S. and Benson, L. V. (1990). Paleoclimatic implications of the high stand of Lake Lahontan
653 derived from models of evaporation and lake level. *Climate Dynamics*, 4(3):207–217.

654 Imbrie, J., Hays, J. D., Martinson, D. G., McIntyre, A., Mix, A. C., Morley, J. J., Pisias, N. G., Prell, W.
655 L., and Shackleton, N. J. The orbital theory of Pleistocene climate: Support from a revised
656 chronology of the marine $\delta^{18}\text{O}$ record. In *Milankovitch and climate: Understanding the response*
657 to astronomical forcing, volume 1, page 269, (1984).

658 Imbrie, J., Mix, A., and Martinson, D. (1993). Milankovitch theory viewed from Devils Hole. *Nature*,
659 363(6429):531–533.

660 Ingraham, N. L. and Taylor, B. E. (1991). Light stable isotope systematics of large-scale hydrologic
661 regimes in California and Nevada. *Water Resources Research*, 27(1):77–90.

662 Ivanovich, M. and Harmon, R. S. (1982). Uranium series disequilibrium: applications to environmental
663 problems. Oxford University Press, USA.

664 Johnson, K. R., Hu, C., Belshaw, N. S., and Henderson, G. M. (2006). Seasonal trace-element and
665 stable-isotope variations in a Chinese speleothem: the potential for high-resolution
666 paleomonsoon reconstruction. *Earth and Planetary Science Letters*, 244(1):394–407.

667 Kaufman, D. S. (2003). Amino acid paleothermometry of Quaternary ostracodes from the Bonneville
668 Basin, Utah. *Quaternary Science Reviews*, 22(8):899–914.

669 Kawamura, K., Parrenin, F., Lisiecki, L., Uemura, R., Vimeux, F., Severinghaus, J. P., Hutterli, M. A.,
670 Nakazawa, T., Aoki, S., Jouzel, J., et al. (2007). Northern Hemisphere forcing of climatic cycles
671 in Antarctica over the past 360,000 years. *Nature*, 448(7156):912–916.

672 Lachniet, M. S., Denniston, R. F., Asmerom, Y., and Polyak, V. J. (2014). Orbital control of western
673 North America atmospheric circulation and climate over two glacial cycles. *Nature*
674 *communications*, 5.

675 Ludwig, K. R., Simmons, K. R., Szabo, B. J., Winograd, I. J., Landwehr, J. M., Riggs, A. C., and
676 Hoffman, R. J. (1992). Mass-spectrometric ^{230}Th - ^{234}U - ^{238}U dating of the Devils Hole calcite
677 vein. *Science*, 258(5080):284–287.

678 Madsen, D. B., Rhode, D., Grayson, D. K., Broughton, J. M., Livingston, S. D., Hunt, J., Quade, J.,
679 Schmitt, D. N., and Shaver, M. (2001). Late Quaternary environmental change in the Bonneville
680 basin, western USA. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 167(3):243–271.

681 McCabe, G. J., Palecki, M. A., and Betancourt, J. L. (2004). Pacific and Atlantic Ocean influences on
682 multidecadal drought frequency in the United States. *Proceedings of the National Academy of*
683 *Sciences*, 101(12):4136–4141.

684 McDermott, F. (2004). Palaeo-climate reconstruction from stable isotope variations in speleothems: a
685 review. *Quaternary Science Reviews*, 23(7):901–918.

686 McGee, D., Quade, J., Edwards, R. L., Broecker, W. S., Cheng, H., Reiners, P. W., and Evenson, N.
687 (2012). Lacustrine cave carbonates: Novel archives of paleohydrologic change in the Bonneville
688 Basin (Utah, USA). *Earth and Planetary Science Letters*, 351:182–194.

689 Munroe, J. S. and Laabs, B. J. (2013)a. Temporal correspondence between pluvial lake highstands in the
690 southwestern US and Heinrich Event 1. *Journal of Quaternary Science*, 28(1):49–58.

691 Munroe, J. S. and Laabs, B. J. (2013)b. Latest Pleistocene history of pluvial Lake Franklin, northeastern
692 Nevada, USA. *Geological Society of America Bulletin*, 125(3-4):322–342.

693 Musgrove, M., Banner, J. L., Mack, L. E., Combs, D. M., James, E. W., Cheng, H., and Edwards, R. L.
694 (2001). Geochronology of late Pleistocene to Holocene speleothems from central Texas:
695 Implications for regional paleoclimate. *Geological Society of America Bulletin*, 113(12):1532–
696 1543.

697 Okumura, Y. M., Deser, C., Hu, A., Timmermann, A., and Xie, S.-P. (2009). North Pacific climate
698 response to freshwater forcing in the subarctic North Atlantic: Oceanic and atmospheric
699 pathways. *Journal of Climate*, 22(6):1424–1445.

700 Orland, I. J., Burstyn, Y., Bar-Matthews, M., Kozdon, R., Ayalon, A., Matthews, A., and Valley, J. W.
701 (2014). Seasonal climate signals (1990–2008) in a modern Soreq Cave stalagmite as revealed by
702 high-resolution geochemical analysis. *Chemical Geology*, 363:322–333.

703 Osmond, J. Uranium disequilibrium in hydrologic studies. In *Handbook of environmental isotope*
704 *geochemistry*. Vol. 1. (1980).

705 Oster, J. L., Ibarra, D. E., Winnick, M. J., and Maher, K., (2015). Steering of westerly storms over
706 western North America at the Last Glacial Maximum. *Nature Geoscience*.

707 Oviatt, C. G. (1997). Lake Bonneville fluctuations and global climate change. *Geology*, 25(2):155–158.

708 Paul, A. P. and Thodal, C. E. (2014). Preliminary Geochemical Assessment of Water from Selected
709 Streams, Springs, and Caves in the Upper Baker and Snake Creek Drainages in Great Basin
710 National Park, Nevada, 2009. Technical report, United States Geological Survey.

711 Petit, J.-R., Jouzel, J., Raynaud, D., Barkov, N. I., Barnola, J.-M., Basile, I., Bender, M., Chappellaz, J.,
 712 Davis, M., Delaygue, G., et al. (1999). Climate and atmospheric history of the past 420,000
 713 years from the Vostok ice core, Antarctica. *Nature*, 399(6735):429–436.

714 Polyak, V. J., Rasmussen, J. B., and Asmerom, Y. (2004). Prolonged wet period in the southwestern
 715 United States through the Younger Dryas. *Geology*, 32(1):5–8.

716 Prudic, D. E. and Glancy, P. A. (2009). *Geochemical Investigation of Source Water to Cave Springs,*
 717 *Great Basin National Park, White Pine County, Nevada.* US Department of the Interior, US
 718 Geological Survey.

719 Quade, J., Forester, R. M., and Whelan, J. F. (2003). Late Quaternary paleohydrologic and
 720 paleotemperature change in southern Nevada. *SPECIAL PAPERS-GEOLOGICAL SOCIETY OF*
 721 *AMERICA*, pages 165–188.

722 Ramsey, C. B. (1995). Radiocarbon calibration and analysis of stratigraphy; the OxCal program.
 723 *Radiocarbon*, 37(2):425–430.

724 Ramsey, C. B. and Lee, S. (2013). Recent and planned developments of the program OxCal.
 725 *Radiocarbon*, 55(2–3):720–730.

726 Roberts, M. S., Smart, P. L., and Baker, A. (1998). Annual trace element variations in a Holocene
 727 speleothem. *Earth and Planetary Science Letters*, 154(1):237–246.

728 Rohling, E. J., Grant, K., Hemleben, C., Siddall, M., Hoogakker, B., Bol-shaw, M., and Kucera, M.
 729 (2008). High rates of sea-level rise during the last interglacial period. *Nature Geoscience*,
 730 1(1):38–42.

731 Seager, R., Ting, M., Held, I., Kushnir, Y., Lu, J., Vecchi, G., Huang, H.-P., Harnik, N., Leetmaa, A.,
 732 Lau, N.-C., et al. (2007). Model projections of an imminent transition to a more arid climate in
 733 southwestern North America. *Science*, 316(5828):1181–1184.

734 Shakun, J.D., Clark, P.U., He, F., Lifton, N., Liu, Z., and Otto-Bliesner, B.L. Regional and global
 735 forcing of glacier retreat during the last deglaciation. *Nature Communications*, accepted.

736 Shakun, J. D., Burns, S. J., Clark, P. U., Cheng, H., and Edwards, R. L. (2011). Milankovitch-paced
 737 Termination II in a Nevada speleothem? *Geophysical Research Letters*, 38(18).

738 Steponaitis, E., Andrews, A. L., McGee, D., Quade, J., Hsieh, Y.-T., Broecker, W. S., Shuman, B. N.,
 739 Burns, S. J., and Cheng, H. (in press). Mid-Holocene drying of the U.S. Great Basin recorded in
 740 Nevada speleothems. *Quaternary Science Reviews*.

741 Sutton, R. T. and Hodson, D. L. (2005). Atlantic Ocean forcing of North American and European
 742 summer climate. *Science*, 309(5731): 115–118.

743 Sutton, R. T. and Hodson, D. L. (2007). Climate response to basin-scale warming and cooling of the
 744 North Atlantic Ocean. *Journal of Climate*, 20(5):891–907.

745 Tan, L., Shen, C.-C., Cai, Y., Lo, L., Cheng, H., and An, Z. (2014). Trace-element variations in an
746 annually layered stalagmite as recorders of climatic changes and anthropogenic pollution in
747 Central China. *Quaternary Research*, 81(2):181–188.

748 Treble, P., Shelley, J., and Chappell, J. (2003). Comparison of high resolution sub-annual records of
749 trace elements in a modern (1911–1992) speleothem with instrumental climate data from
750 southwest Australia. *Earth and Planetary Science Letters*, 216(1):141–153.

751 Tremaine, D. M. and Froelich, P. N. (2013). Speleothem trace element signatures: A hydrologic
752 geochemical study of modern cave dripwaters and farmed calcite. *Geochimica et Cosmochimica*
753 *Acta*, 121:522–545.

754 Wagner, J., Cole, J., Beck, J., Patchett, P., Henderson, G., and Barnett, H. (2010). Moisture variability in
755 the southwestern United States linked to abrupt glacial climate change. *Nature Geoscience*,
756 3(2):110–113.

757 Wang, Y., Cheng, H., Edwards, R. L., Kong, X., Shao, X., Chen, S., Wu, J., Jiang, X., Wang, X., and
758 An, Z. (2008). Millennial-and orbital-scale changes in the East Asian monsoon over the past
759 224,000 years. *Nature*, 451(7182):1090–1093.

760 Wang, Y.-J., Cheng, H., Edwards, R. L., An, Z., Wu, J., Shen, C.-C., and Dorale, J. A. (2001). A high-
761 resolution absolute-dated late Pleistocene monsoon record from Hulu Cave, China. *Science*,
762 294(5550):2345–2348.

763 Winograd, I. J. (2002). The California current, Devils Hole, and Pleistocene climate. *Science*,
764 296(5565):7a–7b.

765 Winograd, I. J., Riggs, A. C., & Coplen, T. B. (1998). The relative contributions of summer and cool-
766 season precipitation to groundwater recharge, Spring Mountains, Nevada, USA. *Hydrogeology*
767 *Journal*, 6(1), 77-93.

768 Winograd, I. J., Coplen, T. B., Landwehr, J. M., Riggs, A. C., Ludwig, K. R., Szabo, B. J., Kolesar, P.
769 T., and Revesz, K. M. (1992). Continuous 500,000-year climate record from vein calcite in
770 Devils Hole, Nevada. *Science*, 258(5080):255–260.

771 Xie, P. and Arkin, P. A. (1997). Global precipitation: A 17-year monthly analysis based on gauge
772 observations, satellite estimates, and numerical model outputs. *Bulletin of the American*
773 *Meteorological Society*, 78(11):2539–2558.

Table 1. U-Th dating results

Depth	^{238}U	^{238}U	^{232}Th	$\delta^{234}\text{U}^*$	$^{230}\text{Th} / ^{238}\text{U}$	^{230}Th Age (kyr)	^{230}Th Age (kyr BP)	$\delta^{234}\text{U}_{\text{Initial}}^{**}$
(mm)	(ppb)	(ng)	(ppt)	(meas.)	(activity)	(uncorr.)	(corrected)	(corrected)
<i>WR-41</i>								
6.0	807 $\pm$ 1.8	4.20	727 $\pm$ 27	325.3 $\pm$ 2.4	0.7121 $\pm$ 0.0038	80.8 $\pm$ 0.7	80.7$\pm$0.7	408.7 $\pm$ 3.1
6.5	530 $\pm$ 2.4	-	690 $\pm$ 14	336.8 $\pm$ 3.4	0.7235 $\pm$ 0.0034	81.6 $\pm$ 0.6	81.5$\pm$0.6	424.0 $\pm$ 4.3
10.0	661 $\pm$ 0.8	5.22	1771 $\pm$ 39	342.7 $\pm$ 1.7	0.7329 $\pm$ 0.0024	82.5 $\pm$ 0.4	82.4$\pm$0.4	432.6 $\pm$ 2.2
18.0	522 $\pm$ 1.1	2.92	5394 $\pm$ 111	324.6 $\pm$ 2.4	0.7187 $\pm$ 0.0040	81.8 $\pm$ 0.7	81.7$\pm$0.7	408.9 $\pm$ 3.1
20.0	426 $\pm$ 0.5	4.13	1617 $\pm$ 35	383.9 $\pm$ 1.7	0.7688 $\pm$ 0.0028	84.4 $\pm$ 0.5	84.3$\pm$0.5	487.0 $\pm$ 2.3
22.0	438 $\pm$ 1.6	-	1089 $\pm$ 22	402.6 $\pm$ 3.1	0.7789 $\pm$ 0.0031	84.2 $\pm$ 0.6	84.1$\pm$0.6	510.5 $\pm$ 4.0
23.0	688 $\pm$ 1.4	4.33	1787 $\pm$ 41	426.9 $\pm$ 2.3	0.7932 $\pm$ 0.0041	84.2 $\pm$ 0.7	84.1$\pm$0.7	541.3 $\pm$ 3.0
24.0	786 $\pm$ 1.6	4.40	611 $\pm$ 24	205.7 $\pm$ 2.0	0.8345 $\pm$ 0.0040	122.3 $\pm$ 1.1	122.2$\pm$1.1	290.5 $\pm$ 3.0
26.0	836 $\pm$ 1.3	5.27	1967 $\pm$ 44	209.1 $\pm$ 1.8	0.8463 $\pm$ 0.0030	124.7 $\pm$ 0.9	124.5$\pm$0.9	297.2 $\pm$ 2.7
26.0	844 $\pm$ 1.3	5.57	1918 $\pm$ 43	205.8 $\pm$ 2.0	0.8437 $\pm$ 0.0060	124.7 $\pm$ 1.6	124.5$\pm$1.6	292.5 $\pm$ 3.1
30.0	718 $\pm$ 0.8	7.90	5134 $\pm$ 104	223.9 $\pm$ 1.6	0.8550 $\pm$ 0.0022	123.9 $\pm$ 0.7	123.6$\pm$0.7	317.5 $\pm$ 2.3
46.0	892 $\pm$ 1.2	7.05	2365 $\pm$ 50	198.9 $\pm$ 1.5	0.8303 $\pm$ 0.0024	122.5 $\pm$ 0.7	122.4$\pm$0.7	281.1 $\pm$ 2.2
74.0	547 $\pm$ 2.4	-	773 $\pm$ 16	204.2 $\pm$ 3.1	0.8371 $\pm$ 0.0043	123.2 $\pm$ 1.3	123.1$\pm$1.3	289.2 $\pm$ 4.5
74.5	754 $\pm$ 1.6	4.37	1293 $\pm$ 33	195.5 $\pm$ 2.0	0.8283 $\pm$ 0.0038	122.7 $\pm$ 1.1	122.6$\pm$1.1	276.3 $\pm$ 3.0
76.0*	333 $\pm$ 0.6	1.90	277 $\pm$ 21	295.9 $\pm$ 2.5	0.9573 $\pm$ 0.0055	135.3 $\pm$ 1.5	135.2$\pm$1.5	433.5 $\pm$ 4.1
77.0	218 $\pm$ 0.6	-	194 $\pm$ 4	275.1 $\pm$ 2.7	0.9183 $\pm$ 0.0028	129.7 $\pm$ 0.9	129.6$\pm$0.9	396.6 $\pm$ 4.0
79.0	313 $\pm$ 0.4	3.38	266 $\pm$ 12	246.2 $\pm$ 1.8	0.9000 $\pm$ 0.0035	131.0 $\pm$ 1.0	130.9$\pm$1.0	356.4 $\pm$ 2.8
80.0	243 $\pm$ 0.6	1.39	190 $\pm$ 21	234.7 $\pm$ 3.5	0.8874 $\pm$ 0.0074	130.1 $\pm$ 2.1	130.0$\pm$2.1	338.8 $\pm$ 5.4
80.0	243 $\pm$ 0.5	1.39	168 $\pm$ 21	232.3 $\pm$ 3.2	0.8878 $\pm$ 0.0071	130.8 $\pm$ 2.0	130.7$\pm$2.0	335.9 $\pm$ 5.0
83.0	298 $\pm$ 0.5	1.82	663 $\pm$ 23	155.5 $\pm$ 2.2	0.8210 $\pm$ 0.0052	129.5 $\pm$ 1.6	129.4$\pm$1.6	224.1 $\pm$ 3.3
85.0	229 $\pm$ 0.6	1.35	426 $\pm$ 22	148.9 $\pm$ 3.4	0.8231 $\pm$ 0.0073	131.7 $\pm$ 2.3	131.6$\pm$2.3	215.9 $\pm$ 5.1
86.0	167 $\pm$ 0.4	1.00	121 $\pm$ 20	149.2 $\pm$ 3.7	0.8330 $\pm$ 0.0092	134.7 $\pm$ 3.0	134.6$\pm$3.0	218.1 $\pm$ 5.7
87.0	286 $\pm$ 0.7	1.69	840 $\pm$ 26	159.6 $\pm$ 3.3	0.8370 $\pm$ 0.0063	133.3 $\pm$ 2.1	133.1$\pm$2.0	232.4 $\pm$ 4.9
91.0	258 $\pm$ 0.5	1.50	655 $\pm$ 24	165.3 $\pm$ 2.6	0.8573 $\pm$ 0.0059	138.0 $\pm$ 2.0	137.9$\pm$2.0	243.9 $\pm$ 4.0
96.0	260 $\pm$ 0.5	1.59	750 $\pm$ 25	164.0 $\pm$ 2.7	0.8504 $\pm$ 0.0057	136.2 $\pm$ 1.9	136.1$\pm$1.9	240.9 $\pm$ 4.2
113.0	320 $\pm$ 0.6	1.70	525 $\pm$ 24	170.2 $\pm$ 2.3	0.8621 $\pm$ 0.0056	138.2 $\pm$ 1.9	138.1$\pm$1.9	251.4 $\pm$ 3.7
119.0	333 $\pm$ 0.6	2.23	601 $\pm$ 21	169.8 $\pm$ 2.4	0.8555 $\pm$ 0.0054	136.3 $\pm$ 1.7	136.2$\pm$1.7	249.5 $\pm$ 3.7
119.0	318 $\pm$ 0.5	2.58	560 $\pm$ 18	172.5 $\pm$ 2.1	0.8568 $\pm$ 0.0043	136.0 $\pm$ 1.4	135.9$\pm$1.4	253.2 $\pm$ 3.2
140.0	344 $\pm$ 0.4	3.92	1665 $\pm$ 35	169.4 $\pm$ 1.8	0.8549 $\pm$ 0.0032	136.2 $\pm$ 1.1	136.0$\pm$1.1	248.8 $\pm$ 2.8
141.0	380 $\pm$ 0.6	2.93	991 $\pm$ 25	170.5 $\pm$ 2.1	0.8520 $\pm$ 0.0044	135.1 $\pm$ 1.4	134.9$\pm$1.4	249.5 $\pm$ 3.3
142.0*	208 $\pm$ 0.4	-	1014 $\pm$ 20	179.6 $\pm$ 9.6	0.8528 $\pm$ 0.0021	133.1 $\pm$ 2.4	132.9$\pm$2.4	261.4 $\pm$ 14.1
151.0	318 $\pm$ 0.5	2.16	4254 $\pm$ 87	173.5 $\pm$ 2.0	0.8699 $\pm$ 0.0050	139.8 $\pm$ 1.7	139.4$\pm$1.7	257.2 $\pm$ 3.1
<i>IR-3</i>								
1.5	140 $\pm$ 0.4	0.81	537 $\pm$ 23	312.3 $\pm$ 5.7	0.9327 $\pm$ 0.0120	125.8 $\pm$ 3.1	125.7$\pm$3.1	445.4 $\pm$ 9.0
2.0*	189 $\pm$ 0.2	2.66	506 $\pm$ 11	321.1 $\pm$ 2.0	1.1542 $\pm$ 0.0040	191.7 $\pm$ 1.8	191.6$\pm$1.8	551.6 $\pm$ 4.4
3.0	189 $\pm$ 0.4	1.42	136 $\pm$ 16	305.2 $\pm$ 3.2	0.9443 $\pm$ 0.0087	130.0 $\pm$ 2.3	130.0$\pm$2.3	440.4 $\pm$ 5.4
3.0	197 $\pm$ 0.5	1.30	126 $\pm$ 18	306.5 $\pm$ 3.8	0.9471 $\pm$ 0.0080	130.5 $\pm$ 2.1	130.4$\pm$2.1	442.9 $\pm$ 6.1
8.0	195 $\pm$ 0.3	2.59	1042 $\pm$ 23	295.8 $\pm$ 2.1	0.9427 $\pm$ 0.0045	131.5 $\pm$ 1.2	131.4$\pm$1.2	428.7 $\pm$ 3.4
16.0	195 $\pm$ 0.3	2.67	331 $\pm$ 11	279.7 $\pm$ 1.9	0.9254 $\pm$ 0.0044	130.5 $\pm$ 1.2	130.4$\pm$1.2	404.2 $\pm$ 3.1
19.0*	188 $\pm$ 0.2	2.50	237 $\pm$ 5	274.3 $\pm$ 1.9	0.9863 $\pm$ 0.0033	148.7 $\pm$ 1.1	148.6$\pm$1.1	417.3 $\pm$ 3.2
22.0	180 $\pm$ 0.3	1.03	375 $\pm$ 22	270.3 $\pm$ 4.6	0.9205 $\pm$ 0.0091	131.2 $\pm$ 2.6	131.1$\pm$2.6	391.3 $\pm$ 7.3

The uncertainties quoted are $\pm 2\sigma$.

$$* \delta^{234}\text{U} = \left(\left[^{234}\text{U}/^{238}\text{U} \right]_{\text{activity}} - 1 \right) \times 1000.$$

** $\delta^{234}\text{U}_{\text{initial}}$ was calculated based on ^{230}Th age (T), i.e., $\delta^{234}\text{U}_{\text{initial}} = \delta^{234}\text{U}_{\text{measured}} \times e^{\lambda_{234}T}$

Decay constants are $\lambda_{230} = 9.17052 \times 10^{-6} \text{ y}^{-1}$, $\lambda_{234} = 2.82206 \times 10^{-6} \text{ y}^{-1}$, $\lambda_{238} = 1.55125 \times 10^{-10} \text{ y}^{-1}$.

Corrected ^{230}Th ages assume the initial $^{230}\text{Th}/^{232}\text{Th}$ atomic ratio of $4.4 \pm 2.2 \times 10^{-6}$; those are the values for a material at secular equilibrium, with the bulk earth $^{232}\text{Th}/^{238}\text{U}$ value of 3.8. ^{230}Th ages (kyr BP, where present is the year 1950) are indicated in bold.

In the ^{238}U (ng) column, dashes denote exploratory dates, for which much larger (>50 mg) powders were drilled. Dates with the same depth are replicates; depths with asterisks indicate rejected dates.

Figure %

Figure 8

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