Records of Great Basin precipitation during MIS 11 from two Lehman Cave stalagmites

by

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Submitted to the Department of Earth, Atmospheric, and Planetary Science in partial fulfillment of the requirements for the degree of Bachelor of Science in Earth, Atmospheric, and Planetary Science at the

MASSACHUSETTS INSTITUTE OF TECHNOLOGY

June 2016

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Abstract

Trace elements (Mg/Ca and Sr/Ca) and stable isotopes ($\delta^{13}$C and $\delta^{18}$O) were measured in two coeval stalagmites, LC3 and BT1, from the Lehman Caves, Nevada. BT1 spans 388 to 384 ka. The LC3 record is split into two parts due to a hiatus observed through fluorescent imaging. The section prior to the hiatus spans 411 ka to 402 ka. The post-hiatus section of the stalagmite has a single age of 383 ka; an age model cannot be constructed for this part of the record. These stalagmites span Marine Isotope Stage 11 (MIS 11), a long interglacial that occurred around 424-374 ka. Comparison with more recent stalagmite records has shown prior calcite precipitation to be the dominant control on Mg/Ca and $\delta^{13}$C in the cave. The trace element and stable isotope records obtained point to an arid climate in the Great Basin during MIS 11.

1 Introduction

This study investigates how precipitation in the Great Basin responded to high latitude warming during MIS 11. Trace element and stable isotope proxies are used to reconstruct precipitation trends. Comparisons with trace element and stable isotope data from more recent interglacials (Steponaitis et al., 2015; Cross et al., 2015) will be made to investigate
whether the precipitation response in the Great Basin during MIS 11 was anomalous. Fluorescent imaging was used to examine organic matter rich laminae and to look for a suspected growth hiatus.

Climate models predict that future global warming will be accompanied by a poleward spreading and intensification of the Pacific storm tracks. Subtropical regions, such as the Great Basin, are expected to become drier while the tropics and high latitudes become wetter. This dryness, a result of reduced precipitation minus evaporation, will primarily stem from reduced winter precipitation (Seager & Vecchi, 2010). This is likely similar to what occurred during MIS 11; for this reason it has been used as an analogue for future climate change (Howard, 1997). The MIS 11 record in this study may supplement model predictions to contribute towards a better understanding of Great Basin hydroclimate under warming conditions.

1.1 Lehman Caves speleothem records

The Lehman Caves are located in Nevada at 39°00′20″N, 114°13′13″W at 2130 m elevation. Studies on more recent stalagmites from the Lehman Caves have been done by Steponaitis et al. (2015), Cross et al. (2015), Shakun et al. (2011), and Lachniet et al. (2014).

The study by Steponaitis et al. (2015) spans 16.4 to 3.8 ka. Covariation between Mg/Ca and δ¹³C was observed; this is interpreted as an indication of degassing and prior calcite precipitation being the main controls on them. Covariation of Sr/Ca and Mg/Ca is only observed for part of the record. This is suggested to be due to prior calcite precipitation controlling Mg/Ca throughout the record, controlling Sr/Ca where it shows covariation with Mg/Ca, and dust deposition controlling Sr/Ca where the covariation is not observed. δ¹⁸O of the record is interpreted to reflect cave temperature and precipitation (Steponaitis et al., 2015). However, the authors note the poor reproducibility of the δ¹⁸O record of stalagmites
in the region.

The study by Cross et al. (2015) spans 139 to 128 ka. Similar to the previous record, Cross et al. (2015) found Mg/Ca and Sr/Ca covarying for only parts of the record. For the periods that Mg/Ca and Sr/Ca do covary, $\delta^{13}$C also covaries, which is thought to be indicative of prior calcite precipitation as a control. The $\delta^{18}$O records from the stalagmites studied are interpreted to reflect temperature and changes in moisture source (Cross et al., 2015).

1.2 Marine Isotope Stage 11

Marine Isotope Stage 11 (MIS 11) was a long interglacial, often referred to as a super-interglacial, that occurred around 424-374 ka. Due to the similarity of Earth’s orbital geometry during MIS 11 to the present day, MIS 11 is often used as an analogue for future climate (Howard, 1997). Around 410-400 ka, pronounced high latitude warming occurred in the Northern Hemisphere for which orbital parameters and greenhouse gas concentrations are unlikely to be the sole causes (Melles et al., 2012).

Fawcett et al. (2011) used organic geochemical proxies from a sediment core from the Valles Caldera, New Mexico to reconstruct mean annual temperature. It was found that driest conditions occurred during the warmest periods of interglacials. Three cycles of mean annual temperature during MIS 11 were identified and seem to correspond to precessional cycles. MIS 11 is divided into five substages: warmer substages a, c, and e corresponding to precessional peaks, with substages b and d having temperatures $\sim$2°C lower. MIS 11e is the oldest stage and MIS 11a is the most recent. High calcium concentration in the sediment core during MIS 11e corresponds to closed basin conditions brought on by rapid temperature increase. During early MIS 11e, $\delta^{13}$C$_{TOC}$ is seen to rise, suggesting expansion of C4 plants. This rise in C4 plants is thought to indicate robust summer precipitation. Mudcracks from
later in MIS 11e, along with a decrease in $\delta^{13}\text{C}_{\text{TOC}}$, suggest dry conditions and decline of C4 plants caused by a reduction in summer precipitation. The following substage, MIS 11d, shows evidence for open-basin conditions, increased summer rainfall, and increase of C4 plants. Early MIS 11e, MIS 11c, and MIS 11a exhibit high mean annual temperatures and robust summer precipitation. High temperatures are linked to summer precipitation through land surface heating driving the summer monsoon. However, the warmest parts of MIS 11e do not follow this trend, and are in fact arid periods in the region’s history.

Melles et al. (2012) studied a sediment core from northeastern Russia that records MIS 11. Biological and geochemical proxies are used to reconstruct past environmental conditions. Palynological records indicate that MIS 11c had a mean warmest month temperature 4-5°C higher and annual precipitation ~300 mm higher than more recent interglacials, MIS 1 and 5e. Relatively low intensity summer insolation for a prolonged time during MIS 11c is interpreted to be caused by low eccentricity and obliquity reducing the effect of precession.

Sea level during MIS 11 has been estimated to have been up to 6-13 m higher than present day sea level. Such an increase would require collapse of at least one major ice sheet (Reyes et al., 2014; Howard, 1997). Reyes et al. (2014) suggest that the size of the Greenland ice sheet may have been significantly reduced during MIS 11. Two hypotheses are proposed for the deglaciation of the Greenland ice sheet during MIS 11. The high-latitude warmth exhibited during MIS 11 may have crossed an ice-stability threshold for the Greenland ice sheet or the length of this warm interglacial may have allowed the ice sheet to fully respond unlike during shorter interglacials (Reyes et al., 2014).

### 1.3 Trace element and stable isotope proxies

Trace elements in speleothems offer useful paleoclimatological information. Fairchild et al. (2000) suggests four factors controlling trace element concentrations observed in speleothems:
differences in the rates of dissolution of calcite and dolomite, prior calcite precipitation in the cave system, incongruent dolomite dissolution, and selective leaching of certain elements. Geochemistry of speleothems may be influenced by atmospheric input, soil and vegetation characteristics, the karst aquifer, and crystal growth and alteration (Fairchild & Treble, 2009). Although trace elements may come from other sources, such as aeolian transport, most are sourced from the bedrock and overlying regolith (Fairchild & Treble, 2009).

The incorporation of trace elements into calcite is governed by a distribution coefficient,

\[ D_X = \frac{(X/Ca)_{calcite}}{(X/Ca)_{fluid}} \]  

where \( X \) represents the trace element (Mg or Sr in this case) and \( D_{Mg} \) and \( D_{Sr} \) are less than 1 (Steponaitis et al., 2015). The distribution coefficients for trace elements are relatively constant. Although they may be influenced by temperature, growth rate, growth mechanisms, or structural characteristics of the growth surface, distribution coefficients are consistent within cave environments (Fairchild et al., 2000). Rapid growth rates may drive trace element incorporation away from equilibrium element partitioning (Fairchild & Treble, 2009). Temperature, precipitation, crystal morphology, and solution properties may all affect the distribution coefficient (Fairchild & Treble, 2009). The distribution coefficient for Mg may be dependent on temperature (Huang & Fairchild, 2001).

Trace elements such as Mg and Sr may substitute for Ca in the calcite crystal lattice (Fairchild and Treble, 2009). Mg/Ca is commonly used as a proxy for paleoprecipitation. Its relationship with \( \delta^{13}C \) may be used to argue for the effectiveness of Mg/Ca in recording precipitation (Fairchild & Treble, 2009). As Mg and Sr may both record other factors, covariation of Mg and Sr may provide a more compelling argument for paleoprecipitation records. However, caution should be used when interpreting trace element records as records of rainfall, as they may also reflect changes in the water route or source (Fairchild et al., 2000).
Prior calcite precipitation will increase during drier times as there will be a greater likelihood of degassing into air pockets (Fairchild et al., 2000). When waters percolating through the soil and overlying rock reach a gas with pCO$_2$ lower than that which they equilibrated with, degassing will occur so as to re-equilibrate. In this process, calcite precipitates out (Fairchild & Treble, 2009). Prior calcite precipitation is enhanced by seasonal falls in cave CO$_2$ and seasonally low water flows. Low pCO$_2$ will cause calcite to precipitate from the solution. Conversely, high pCO$_2$ will increase dissolution, thus increasing the solute load (Fairchild & Treble, 2009).

During drier periods, water will move slowly through the soil and epikarst. This gives a greater chance for degassing to occur, causing more calcite to be precipitated from the solution before it reaches the stalagmite. Since $D_{Mg}$ and $D_{Sr}$ are less than 1, the solution will become more enriched in Mg and Sr as calcite precipitates, preferentially incorporating Ca into the calcite. Thus, the Mg/Ca and Sr/Ca ratios of the stalagmite will be higher (Fairchild et al., 2000).

As discussed, degassing of CO$_2$ causes calcite to precipitate. When degassing occurs isotopically light carbon will be incorporated into the gas phase, leaving the solution with an enriched $\delta^{13}C$. Thus, similar to trace element ratios, $\delta^{13}C$ may be used as an indicator of paleoprecipitation in cases where there are no other factors exerting a strong control.

McDermott (2004) discusses open and closed system models and their effects on speleothem $\delta^{13}C$. Under an open system model there is constant equilibration between the solution and soil CO$_2$. This leads to a $\delta^{13}C$ which reflects only the soil CO$_2$. In a closed system model the solution is cut off from soil CO$_2$ once carbonate dissolution in the epikarst begins. The extent of the carbonate dissolution that can occur is limited by the available CO$_2$. Under these conditions, the speleothem $\delta^{13}C$ will reflect that of the host-rock. In reality, the processes taking place in caves are likely to lie somewhere between these two models.

$\delta^{13}C$ may also be used as a proxy for vegetation change. C3 plants have more depleted
$\delta^{13}C$ values than C4 plants. Shifts in the ratio of C3 to C4 plants above the cave will influence the $\delta^{13}C$ value (McDermott, 2004). McDermott (2004) also suggests that the relative proportions of atmospheric and biogenic carbon may be reflected in the speleothem $\delta^{13}C$. Biogenically produced carbon is isotopically lighter; periods of high primary productivity in the soil overlying the cave may be recorded by lighter $\delta^{13}C$ in the speleothem calcite.

$\delta^{18}O$ records from speleothems may be used to reconstruct meteoric precipitation or temperature, as $\delta^{18}O$ values of speleothem calcite will be controlled by only cave temperature and $\delta^{18}O$ value of the drip water if equilibrium fractionation is occurring (Lachniet, 2009). However, complexity exists in interpreting $\delta^{18}O$ records due to the many potential controls such as moisture source, kinetic effects when calcite is precipitating, and isotope effects during phase changes along the meteoric water line. In regions where amount effect is strong, $\delta^{18}O$ may directly reflect rainfall rather than temperature (Fairchild & McMillan, 2007). Lachniet (2009) stresses the importance of studying modern day cave processes in aiding interpretation of speleothem $\delta^{18}O$ records, as processes and moisture sources vary between sites. Trace element paleoprecipitation records allow for the source of $\delta^{18}O$ in speleothems to be better understood (Fairchild & Treble, 2009).

1.4 Fluorescence and annual laminae

A speleothem may show annual laminae if there is an annual rhythm in the climate at the surface and some mechanism that transfers this rhythm to the speleothem (Baker et al., 2008). The larger the magnitude of this rhythm, the more likely that annual laminae will form.

Several types of annual laminae may be present. Visible laminae may result from alterations of the crystal arrangements due to seasonal variations in drip rate or drip water supersaturation (Baker et al., 2008). van Beynen et al. (2001) found that darker layers of
calcite had, on average, four times more humic acid than lighter calcite. This study also
found that the particulate organic matter in darker calcite may act so as to quench fluores-
cence through self-absorbance. Laminae of calcite-aragonite couplets resulting from seasonal
alteration of aragonite and calcite growth layers may also be present; however, the controls
on these are poorly understood. Fairchild & McMillan (2007) suggest that seasonal calcite-
aragonite changes may be caused by reduced drip rate and prior calcite precipitation. Trace
element laminae appear to be near universally present. Controls on trace element laminae
may be similar to those on fluorescent or visible laminae. Additionally, transport of ma-
terial in colloidal form may increase trace element concentrations and fluorescent organic
matter (Baker et al., 2008). Organic acids concentration and inorganic detrital content in
speleothems are positively correlated, suggesting material transported in solid phase (Mc-
Garry & Baker, 2000).

Fluorescent laminae are caused by the presence of organic matter. Organic acids are
produced through humification in the soil. Humic acid and fulvic acid are the primary organic
acids observed in speleothems (McGarry & Baker, 2000). Amino-acid groups of proteins
formed from decomposed plant material in the soil may also contribute to fluorescence (Baker
& Genty, 1999). The organic acids become trapped in the speleothem calcite due to the high
zero point charge of the calcite surface and the negative charge of organic matter in the
solution (Baker & Genty, 1999).

Differences in fluorescence within a lamination may reflect seasonal processes. In regions
with Mediterranean climates (dry summers, wet winters), the first flush in autumn/winter
after recharging of groundwater will cause organic material built up from the spring and
summer to enter the groundwater. Increases in discharge increases the amount of organic
matter transported to the speleothem (Baker et al., 2008; McGarry & Baker, 2000). It has
been found that the intensity of fluorescence peaks in autumn/winter after this first flush
(Baker & Genty, 1999). Baker & Genty (1999) found that the second flush showed a decrease
in wavelength of fluorescence. This can be accounted for by the difference in the composition of organic matter from that of the first flush, as the winter flush will contain more poorly humified material (McGarry & Baker, 2000; Baker & Genty, 1999). The additional winter moisture will also increase the availability of dissolved organic matter in the soil (Baker & Genty, 1999). Increase in organic acid concentration will increase the intensity of the fluorescence observed but does not affect the wavelength (McGarry & Baker, 2000). The decreases in wavelength may also be due to metal-ion interactions from the higher calcium ion concentration during the first flush, or due to dissolution of limestone bedrock by the flowing water (Baker & Genty, 1999).

Additionally, porosity of the calcite will affect the intensity of the fluorescence due to total internal reflection. Variable porosity may make paleoprecipitation records over long timescales difficult. However, fluorescence may be a useful high resolution proxy on annual or sub-annual timescales (McGarry & Baker, 2000).

Fluorescence may also be influenced by pH, presence of metal ions, soil composition, or climate. pH may increase the intensity of fluorescence observed by changing characteristics of the molecules functional groups or changing conformations of the organic molecules (Baker & Genty, 1999). However, karst drip waters typically have a pH between 7.0 and 8.5, so the effect of pH on fluorescence is small (Baker & Genty, 1999). Metal ions may act so as to increase or decrease the intensity of fluorescence (Baker & Genty, 1999). In karst waters, calcium dominates metal-ion interactions, allowing for the effect of metal ions to be ignored (Baker & Genty, 1999). Thus, in the case of speleothem records, soil and climate are the dominant controls on fluorescence. Climate controls the rate of humification, affecting the ratio of humic acid to fulvic acid produced (McGarry & Baker, 2000). Rate of humification will increase with temperature and soil moisture (Baker & Genty, 1999). Wetter periods will be accompanied by an increase in dissolved organic matter. It has been observed that the proportion of more hydrophobic acids, such as humic acid, will increase more than more
hydrophilic acids, such as fulvic acids. Colder, wetter conditions may decrease humification, causing a greater proportion of the organic acids to be humic acid. In this case, fluorescence at higher wavelengths will be observed. Conversely, warmer drier conditions will increase humification, and fluorescence at shorter wavelengths will reflect the greater proportion of fulvic acid (McGarry & Baker, 2000).

To determine whether observed laminae are annual in nature they may be compared with ages obtained through U-Th dating. The number of laminae between well dated layers is used for this comparison. The age error on the U-Th analysis is important in whether or not the annual nature can be determined. In the case of the laminae being annual they may be used to provide a precise age-depth model. This is especially useful in the absence of constant growth rates, as growth rate may increase or decrease due to climate trends or changes in stored groundwater characteristics.

2 Methods

2.1 Trace elements

LC3 was sampled every 2 mm along its growth axis while BT1 was sampled every 3 mm. Samples of 2 mg were drilled using a vertical mill. The drill bit was cleaned with dilute HCl, methanol, and compressed air between samples. Compressed air was used to ensure the sample was free of powders before drilling the next sample. Samples were dissolved in 1.5 ml of 0.5 M nitric acid. After dissolution, 20 μl of the solution was diluted in 12 ml of nitric acid. These were analyzed using a quadrupole ICP-MS at the Massachusetts Institute of Technology. A standard with Mg/Ca of 42.22 mmol/mol and Sr/Ca of 0.12 mmol/mol and a blank were prepared. Results for samples were corrected for using the blank and standard.
2.2 Stable isotopes

Both stalagmites were sampled for δ13C and δ18O every 10 mm along the growth axis. The powders were dissolved in dehydrated phosphoric acid at 70°C. They were analyzed with a gas ratio mass spectrometer at the University of Arizona.

2.3 Uranium-Thorium dating

Samples of 200-700 mg were taken from five places on LC3 and eight places on BT1 for dating. Stalagmites were cleaned with ethanol and compressed air prior to drilling. The drill bit was cleaned as described for trace element sampling. Extra caution was taken to ensure that the workspace was clean of any dust that may contaminate the samples. Places for dating were chosen so as to be spread over length of the stalagmite and from places where laminae were visible that could be drilled across. Uranium and thorium concentrations were measured on an ICP-MS at MIT. Age models were constructed using OxCal.

2.4 Fluorescence

Small areas from both stalagmites were imaged using confocal microscopy at MIT. BT1 was imaged between two dated regions. LC3 shows visible variations in calcite color and opacity. This was thought to have been caused by a hiatus in deposition. The area to image was chosen so as to extend across the variations in order to investigate the potential hiatus.

3 Results

3.1 Trace elements

Mg/Ca in BT1 ranges from 0.75 to 22.8 mmol/mol and Sr/Ca ranges from 0.031 to 0.083 mmol/mol. In LC3, Mg/Ca ranges from 6.1 to 191.3 mmol/mol, while Sr/Ca ranges from
Figure 1: Time series for LC3 trace element and stable isotope record before hiatus (top) and after hiatus (bottom). The record for after the hiatus is plotted against depth from the top of the stalagmite as no age model could be constructed for it.
0.035 to 1.3 mmol/mol. The LC3 record may be split into two sections as shown in Figure 1. The first section is more similar to BT1 with Mg/Ca ranging from 6.1 to 32.8 mmol/mol and Sr/Ca ranging from 0.041 to 0.163 mmol/mol, while the second section Mg/Ca ranges from 7.82 to 191.26 mmol/mol and Sr/Ca from 0.035 to 1.286 mmol/mol. Timeseries for the records are shown in Figures 1 and 2.

### 3.2 Stable isotopes

In BT1, $\delta^{18}$O ranged from -13 to -11.5 $\%_\circ$, while $\delta^{13}$C ranged from -7 to -4 $\%_\circ$. In LC3, $\delta^{18}$O ranged from -13.5 to -11 $\%_\circ$, while $\delta^{13}$C ranged from -6 to -2.5 $\%_\circ$. The stable isotope records are shown in Figures 1 and 2.
3.3 Uranium-Thorium dating

The U concentrations from BT1 were between 520 and 700 ng/g. In LC3, U concentrations were between 630 and 1090 ng/g. Ages were calculated from data in Table 1 using the equation

\[
\left[ \frac{^{230}Th}{^{238}U} \right]_{\text{activity}} = 1 - e^{-\lambda_{230}T} + \left( \frac{\delta^{234}U_{\text{measured}}}{1000} \right) \left[ \frac{\lambda_{230}}{\lambda_{230} - \lambda_{234}} \right] \left( 1 - e^{(\lambda_{234} - \lambda_{230})T} \right) \quad (2)
\]

Where \( \lambda \) represents a decay constant and \( T \) is the age. Decay constant was \( 9.1705 \times 10^{-6} \) year\(^{-1} \) for \(^{230}\)Th and \( 2.82206 \times 10^{-6} \) year\(^{-1} \) for \(^{234}\)U (Cheng et al., 2013). A decay constant of \( 1.55125 \times 10^{-10} \) year\(^{-1} \) was used for \(^{238}\)U (Jaffey et al., 1971). An initial \( ^{230}\)Th/\( ^{232}\)Th of \( 4.4 \pm 2.2 \times 10^{-6} \) was assumed to correct for detrital \(^{230}\)Th. A corrected age was calculated using

\[
\left[ \frac{^{230}Th}{^{238}U} \right] - \left[ \frac{^{232}Th}{^{238}U} \right] \left[ \frac{^{230}Th}{^{232}Th} \right] (e^{-\lambda_{230}T}) - 1 = -e^{-\lambda_{230}T} + \left( \frac{\delta^{230}U_{m}}{1000} \right) \left( \frac{\lambda_{230}}{\lambda_{230} - \lambda_{234}} \right) \left( 1 - e^{-(\lambda_{230} - \lambda_{234})T} \right) \quad (3)
\]

where \( T \) is the corrected age (Edwards et al., 2003).

Age models constructed using OxCal are shown in Figure 3. The deepest age taken LC3 has been excluded from the age model due to it being an anomalous outlier. These age models give BT1 an age range of 388±5.8 ka to 384±7.5 ka and average growth rate of 89.69 mm/kyr. LC3 has an age range of 411±4 ka to 402±4 ka and average growth rate of 18.98 mm/kyr (uncertainties are 2\( \sigma \)). A single point dated above the hiatus yields a date of 383±6 ka.
Table 1: Results for U-Th dating. Ages were calculated using Equation 2 and corrected using Equation 3. Uncertainties are 2σ.

<table>
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<th>Depth (cm)</th>
<th>$^{238}$U (ng/g)</th>
<th>$^{232}$Th (pg/g)</th>
<th>$\delta^{234}$U (activity)</th>
<th>$^{230}$Th/$^{238}$U</th>
<th>U-Th Age (ka)</th>
<th>$\delta^{234}$U$_{initial}$</th>
<th>Model Age (ka)</th>
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<td>428.2±9.2</td>
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3.4 Fluorescence

Figure 4 shows the images obtained from the confocal microscope. Images from BT1 (a and b) show fine uninterrupted laminations of varying width. Fig 4(c) shows evidence of a hiatus in LC3, as seen by the curved laminations ending and truncating the later, flatter laminations. Unfortunately, only one date could be obtained from above the hiatus, so it is unclear for how long it lasted.

The images show approximately 11 lamina/mm in BT1 and 27 lamina/mm in LC3. Combining these with the growth rates (89.69 mm/kyr for BT1 and 18.98 mm/kyr for LC3) we get approximately 0.94 lamina/year for BT1 and 0.52 lamina/year for LC3. The fluorescent lamina in BT1 are thus likely to be annual in nature. Other stalagmites from the cave have growth rates ranging from 7 mm/kyr to 30 mm/kyr (Steponaitis et al., 2015). These would correspond to annual laminations of 143 lamina/mm to 33 lamina/mm. Based upon the range of growth rates for this cave it is possible, but unlikely that the lamina in LC3 are recording annual signals.

Figure 4: (a) and (b) show fine lamination present in BT1. (c) shows the hiatus (indicated by arrows) in LC3.
4 Discussion

4.1 U-Th dates and age models

U-Th dating is a method of dating based on disequilibrium between $^{238}\text{U}$ and its daughter product $^{230}\text{Th}$. In time, $^{238}\text{U}$ and $^{230}\text{Th}$ will reach secular equilibrium, a state where their activities are equal (Dorale et al., 2004). Secular equilibrium occurs where the slope is zero as shown on Fig 5. This method allows for stalagmites to be dated to about 600,000-700,000 years BP (Spotl & Boch, 2012). Figure 5 shows change in $\left[\frac{^{230}\text{Th}}{^{238}\text{U}}\right]_{\text{activity}}$ with time for three initial $\delta^{234}\text{U}$ values. High initial $\delta^{234}\text{U}$ allows for more accurate dating of older samples. The initial $\delta^{234}\text{U}$ for BT1 and LC3 were relatively low, which contributes to the difficulty in
Figure 5: From (Dorale et al., 2004). Secular equilibrium corresponds to when solid lines have a slope of zero. $\delta^{234}$U$_{initial}$ ranges from 320 to 457 % in our samples while the $\left[\frac{230}{238}U\right]_{activity}$ ranges from 1.11 to 1.18. As can be seen from the graph, our samples have neared secular equilibrium.

Table 2: Table showing the effects of small change in $\left[\frac{230}{238}U\right]_{activity}$ on ages obtained.

<table>
<thead>
<tr>
<th>Difference from median</th>
<th>$\left[\frac{230}{238}U\right]_{activity}$ (activity)</th>
<th>Age (years BP)</th>
<th>Difference from actual age (years)</th>
<th>Difference from actual age (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0%</td>
<td>1.154</td>
<td>374,873</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>0.1%</td>
<td>1.155</td>
<td>377,685</td>
<td>-2,812</td>
<td>-0.75%</td>
</tr>
<tr>
<td>0.5%</td>
<td>1.160</td>
<td>389,643</td>
<td>-14,770</td>
<td>-3.94%</td>
</tr>
<tr>
<td>1%</td>
<td>1.166</td>
<td>406,489</td>
<td>-31,613</td>
<td>-8.43%</td>
</tr>
</tbody>
</table>

obtaining good ages for them.

Table 2 summarizes the effects that small loss of $^{238}$U would have on the obtained ages. A higher initial $^{230}$Th content than accounted for would also lend to this effect. These calculations demonstrate the large effect on the obtained age that can be caused by $^{238}$U loss, high initial $^{230}$Th, or other error for samples of the age of BT1 and LC3. The age of these samples and the low U content (due in part to age also) contribute to the difficulty in placing precise ages on them.
Table 3: Comparison of trace element and stable isotope ratios from BT1 and LC3 (split into before and after hiatus) with CDR3 and WR11 (Steponaitis et al., 2015) and WR41 and IR3 (Cross et al., 2015).

<table>
<thead>
<tr>
<th></th>
<th>Mg/Ca (mmol/mol)</th>
<th>Sr/Ca (mmol/mol)</th>
<th>δ¹³C (‰)</th>
<th>δ¹⁸O (‰)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>min</td>
<td>max</td>
<td>min</td>
<td>max</td>
</tr>
<tr>
<td>BT1</td>
<td>0.75</td>
<td>22.8</td>
<td>0.031</td>
<td>0.083</td>
</tr>
<tr>
<td>(before hiatus)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>LC3</td>
<td>6.1</td>
<td>32.8</td>
<td>0.041</td>
<td>0.163</td>
</tr>
<tr>
<td>(after hiatus)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CDR3</td>
<td>1.3</td>
<td>2.6</td>
<td>0.083</td>
<td>0.14</td>
</tr>
<tr>
<td>WR11</td>
<td>2.1</td>
<td>6.1</td>
<td>0.087</td>
<td>0.16</td>
</tr>
<tr>
<td>WR41 &amp; IR3</td>
<td>2.1</td>
<td>9.6</td>
<td>0.018</td>
<td>0.039</td>
</tr>
</tbody>
</table>

### 4.2 Trace element and stable isotope records

Table 3 compares the trace element and stable isotope ratios for BT1 and LC3 with those of CDR3 and WR11 from Steponaitis et al. (2015) and WR41 and IR3 from Cross et al. (2015). δ¹³C and δ¹⁸O are very similar in all samples. Mg/Ca has a much larger range in BT1 and LC3 than in the other samples, with the post hiatus Mg/Ca of LC3 being exceptionally high. BT1, WR41 and IR3 have low Sr/Ca compared to all other samples. The Sr/Ca before the hiatus in LC3 is similar to the Sr/Ca of CDR3 and WR11, while that after the hiatus is much higher and has a great range.

While controls on trace element concentrations vary by location even within a particular cave system the comparison of the post-hiatus LC3 record to the others is significant. The much larger maximum values and large range in trace element ratios may indicate an exceptionally drier time but with large variability. The large Mg/Ca ratios of BT1 and pre-hiatus LC3 may also be significant in relation to the other samples; the large magnitudes may indicate a drier climate during MIS 11.

Fig 6 compares Mg/Ca to δ¹³C for LC3 and BT1, and two stalagmites from the same
Figure 6: Mg/Ca vs δ¹³C shows a linear trend for all four stalagmites, indicating that prior calcite precipitation is the main control in the cave.

cave studied by Stepounaitis et al. (2015). The linear trend in all of them indicates that prior calcite precipitation is the dominant control in this cave. Thus, these can be interpreted as proxies for rainfall above the cave.

For almost all of the BT1 record Mg/Ca and Sr/Ca are inversely correlated. Figure 7 shows Mg/Ca vs Sr/Ca for BT1. The linear trend of negative slope is clear. Three outliers have been removed from the data, leaving 86 points in the fit. This trend does not appear to have any dependence on when the calcite was deposited. Rather, it seems to have been
present throughout the record.

Mg/Ca is interpreted as a proxy for paleoprecipitation. Sr may also record paleoprecipitation, however, it may also be an indicator of other processes.

Calcite dissolves much more rapidly than dolomite. Equilibrium with calcite is reached rapidly, while dolomite dissolution slows far from equilibrium (Fairchild & Treble, 2009). Thus, in regions of mixed calcite and dolomite bedrock, increased residence time is expected to increase the Mg content of the solution due to increased dolomite dissolution. Variations in dolomite dissolution may be shown in the inverse relationship between Mg/Ca and Sr/Ca as dolomite is depleted in Sr relative to calcite (Fairchild et al., 2000). Thus, dolomite dissolution occurring in the bedrock may be responsible for the inverse correlation between Mg/Ca and Sr/Ca in BT1. In this case, the high Mg/Ca and low Sr/Ca times would indicate drier climate, as water would need to reside longer in the epikarst for dolomite dissolution to occur.

Sr may be carried to the site by aeolian dust (Fairchild & Treble, 2009). It is possible that this dust is washed down to the cave during wetter periods, which would explain high Sr/Ca coinciding with low Mg/Ca.

Sr is observed to be incorporated more at higher growth rates. This effect seems to be significant if the growth rate exceeds 0.5 mm/year (Fairchild & Treble, 2009). The growth rate for BT1 is 0.09 mm/year, much lower than 0.5 mm/year. It is possible that at times a growth rate equivalent to 0.5 mm/year was exceeded but this is unlikely to be the entire
Figure 8: Mg/Ca vs Sr/Ca for before and after the hiatus in LC3.

cause of the Sr trend.

The time series for LC3 has been split into two parts: before and after the hiatus. After the hiatus in LC3, ages based on a continuation of the projected age model would not be correct. Instead, the record has been plotted against depth from the top of the stalagmite. Greater depths are older and lesser depths are more recent, but no age model could be constructed for this portion. The drastically different trace element profiles suggest that the stalagmite was subject to different growth conditions before and after the hiatus, so growth rates were likely different.

Prior to the hiatus, LC3 spans MIS substage 11c and part of 11d (described by Fawcett et al. (2011)), but does not show any evidence for distinctions between the substages. Rather, it indicates a consistent, generally dry climate.

The relationship between Mg/Ca and Sr/Ca differs before and after the hiatus. Prior to the hiatus, the data as a whole do not show any significant trend. However, the later part of the record before the hiatus shows a positive trend. Influences on the controls of these proxies may have changed during the time period recorded by this stalagmite.
After the hiatus, Mg/Ca and Sr/Ca vary together, as seen in Figure 8, with a positive slope. This suggests that they are both controlled by prior calcite precipitation later in the record.

BT1 and the record after the hiatus in LC3 show opposite Mg/Ca vs Sr/Ca trends. Barring BT1 having grown under an isolated patch of dolomitic epikarst which was not present over LC3 or BT1 having significant growth rate effects, these trends would suggest that the hiatus in LC3 lasted at the very least until the end of the BT1 record, 384 ka.

Lachniet (2009) discusses how processes occurring in the ocean, atmosphere, soil, epikarst, and cave affect the $\delta^{18}O$ values recorded in speleothems. Within the soil and epikarst in arid regions, evaporation effects are likely to play a large role in enriching the $\delta^{18}O$ value of waters. Under equilibrium fractionation conditions, the $\delta^{18}O$ value of the carbonate precipitated will be dominated by the $\delta^{18}O$ value of the water it precipitated from, rather than the carbonate in solution. However, lack of time for isotope exchange reactions to occur or differences in mineralogy will affect the $\delta^{18}O$ value of the speleothem.

Kinetic effects are most prevalent during rapid reactions, such as the rapid CO$_2$ degassing that may occur during drier times. Covariation in $\delta^{18}O$ and $\delta^{13}C$, which may result from evaporation due to cave air being below 100% humidity during drier periods, indicate kinetic fractionation occurring (Fairchild & McMillan, 2007; Lachniet, 2009). Figure 9 compares $\delta^{18}O$ and $\delta^{13}C$ for BT1 and LC3. A general positive trend may be observed. While the correlation is somewhat weak it may indicate that $\delta^{18}O$ is being

![Figure 9: $\delta^{18}O$ vs $\delta^{13}C$ for LC3 and BT1. A general positive correlation can be seen.](image)
controlled by kinetic effects to some extent.
This suggests that the $\delta^{18}O$ records may be recording factors other than the $\delta^{18}O$ of precipitation above the cave, helping to explain the poor reproducibility of $\delta^{18}O$ records from different stalagmites in the region (Steponaitis et al., 2015).

5 Conclusion

The stalagmites in this study span part of MIS 11. While date constraints on the stalagmites have greater uncertainties than those of more recent stalagmites from the cave, the age models present good constraints. The covariation between Mg/Ca and $\delta^{13}C$ indicate that prior calcite precipitation is dominating in the cave, and thus atmospheric precipitation above the cave is being recorded in the stalagmites. The trace element ratios in the stalagmites seem to indicate MIS 11 to have been a time of aridity in the Great Basin.

In BT1 Mg/Ca vs Sr/Ca may be controlled by dolomite dissolution in the epikarst, Sr carried by silicate dust to the site, or growth rate affecting Sr incorporation into the stalagmite. In LC3 no correlation is observed between Mg/Ca and Sr/Ca until after the hiatus, at which time the positive correlation suggests both trace elements are being controlled by prior calcite precipitation and thus recording atmospheric precipitation above the cave. The relationship between $\delta^{13}C$ and $\delta^{18}O$ in both stalagmites indicates that $\delta^{18}O$ is, to some extent, controlled by kinetic effects during calcite precipitation, suggesting dry conditions.

Examining the stalagmites for records of seasonality would be of interest for future study. The annual laminae in BT1 make it an especially strong candidate for such a study. Seager & Vecchi (2010) describe the modelled drying of the Great Basin in the next century to be primarily due to a reduction in winter precipitation. A study of seasonality of the records from the MIS 11 stalagmites could determine whether this was also the mechanism for the
MIS 11 drying.

Another direction for further study is wavelengths of fluorescence. The examination of fluorescent laminae in this study was primarily qualitative. Baker & Genty (1999) identified three excitation-emission wavelength pairs which correspond to different organic acids, amino-acids, and proteins. The relative proportions of organic acids in speleothem calcite has been found to be related to vegetation and climate (McGarry & Baker, 2000). A study of these wavelengths and the corresponding organic acids in BT1 may provide a record of vegetation and climate at an annual resolution.

Having only two stalagmites places limits on this study in terms of reproducibility and extent of the record. It is possible for something happening only at a very localized point in the cave system to effect the trace element and stable isotope composition of a stalagmite. Even between LC3 and BT1 many differences are evident such as the different growth rates, the annual lamina in BT1, and the hiatus in LC3. The records for BT1 and LC3 do not overlap, so reproducibility cannot be tested from them. In an ideal scenario, more stalagmites would be available to reproduce and extend the record. However, despite these limitations, valuable conclusions can be drawn from these records.

The results from BT1 and LC3 seem to indicate an arid climate. The dramatically higher trace element ratios in these MIS 11 stalagmites as compared to Holocene stalagmites from the same cave suggest that during MIS 11 the Great Basin experienced greater aridity than has been experienced in the Holocene. This is largely consistent with the results of other work done on Great Basin climate during MIS 11, such as the drought conditions predicted by Fawcett et al. (2011).
6 Acknowledgments

I am grateful to Professor David McGee for his guidance and support on this project. I would also like to thank the rest of the McGee Lab, especially Irit Tal for her invaluable help. Additionally, I would like to thank Jane Connor for her help with the writing and presentation.

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