Response of a modern cave system to large seasonal precipitation variability

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Received 17 October 2011; accepted in revised form 27 May 2012; available online 7 June 2012

Abstract

Speleothems are capable of providing information on the response of middle and low-latitude terrestrial environments to global climate change during the Pleistocene and Holocene. Multiproxy speleothem studies, however, have demonstrated that complex interactions can occur in cave settings between processes that are directly and indirectly related to climate change. Thorough and extended monitoring of modern cave environments is necessary in order to fully understand how each cave responds to these processes on seasonal and interannual timescales, and how environmental signals are preserved in speleothem carbonate.

Regular environmental monitoring began at Black Chasm Cavern in the Sierra Nevada foothills, California, during the winter of 2006–2007. Monthly measurements of cave air temperature, humidity, and \( p_{\text{CO}_2} \) in Black Chasm demonstrate that the cave is ventilated in the winter months, when cold, dense surface air sinks into the cave. Cave drip-water flow nearly ceases during the late summer and autumn, increases substantially during the winter and spring, and responds within hours to storm events during the height of the rainy season. Rainwater and drip water \( \delta^{18} \text{O} \) and \( \delta^2 \text{H} \) are controlled by variations in surface air temperature and moisture source. While rainfall source influences rainwater isotopes through individual storm events, it has less influence on drip water isotopic composition due to mixing of recharge waters delivered by different rainfall events in the epikarst. Variations in drip water chemistry (\( \delta^{13} \text{C} \), Mg/Ca, and Sr/Ca) indicate that the greatest level of calcite precipitation upflow from the drip water collection site (prior calcite precipitation) occurs during the autumn (October–November) when drip rates are slow and cave air \( p_{\text{CO}_2} \) is low. The least prior calcite precipitation occurs during the summer (July–August) when drip rates are slow but cave air \( p_{\text{CO}_2} \) is at a maximum. While \( p_{\text{CO}_2} \) is a primary control on prior calcite precipitation during all seasons, the predominant influence of drip rate variability on prior calcite precipitation is evident when considering only those seasons (winter, spring, and autumn) characterized by low cave air \( p_{\text{CO}_2} \). Thus, drip rate variability, and in turn rainfall amount, should provide the primary control on trace element variations ultimately captured in speleothem calcite. The isotopic and chemical variability observed in Black Chasm drip waters supports previous interpretations of speleothem paleoclimate proxy records from a nearby cave where such monitoring is not feasible. Observations of the modern cave environment at Black Chasm provide a reference point for the interpretation of stalagmite proxy records from similar seasonal (Mediterranean) climates.

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1. INTRODUCTION

Speleothem geochemical proxy records have the potential to provide valuable archives of rainfall variability, atmospheric circulation changes, and vegetation response in low and mid-latitude continental environments and have contributed significantly to our understanding of hemispheric
teleconnections linking distal regions (e.g. Wang et al., 2001, 2004, 2007; Partin et al., 2007; Oster et al., 2009; Asmerom et al., 2010; Wagner et al., 2010). However, the interpretation of speleothem isotopic and geochemical proxies is rarely straightforward (e.g. Fairchild et al., 2000; Genty et al., 2001a; Mickler et al., 2004, 2006; Lambert and Aharon, 2011). Geochemical proxies (e.g. $\delta^{18}$O, $\delta^{13}$C, trace element concentrations, $^{87}$Sr/$^{86}$Sr) used in speleothem studies each respond to a variety of environmental influences, and these responses can vary significantly between caves, between speleothems within single cave, and through time. Although the majority of speleothem studies have focused on developing records of oxygen isotope variability, recent multiproxy studies have highlighted the complex interactions that occur in cave settings between processes that are directly related to climate change (rainfall amount, temperature) and those that are indirectly related (degree of water–rock interaction, seepage-water degassing and prior calcite precipitation, changes in cave air $p$CO$_2$) (Banner et al., 1996; Baker et al., 1997; Ayalon et al., 1999; Hellstrom and McCulloch, 2000; Johnson et al., 2006; Oster et al., 2010). Isolating those processes that are directly and indirectly related to climate is challenging, and this limitation can compromise the interpretation of speleothem proxy records. Rapid and extended degassing of CO$_2$ driven by decreased seepage water flow rates, increased evaporation, and/or changing cave air–water $p$CO$_2$ can contribute to non-equilibrium isotopic fractionation of stable isotopes during speleothem growth (Hendy, 1971; Mickler et al., 2004). Such kinetic effects lead to speleothem calcite that is enriched in $^{13}$C and $^{18}$O and possesses elevated concentrations of trace elements such as Mg, Sr, and Ba (Lorens, 1981) that do not accurately reflect the chemical signatures carried by the drip water into the cave.

In light of these complexities, it is necessary to perform thorough and extended monitoring of cave environments in order to fully understand how each cave responds to modern seasonal and interannual changes. Cave monitoring involves regular measurements of cave air temperature, humidity, and $p$CO$_2$ (e.g. Spötl et al., 2005; Banner et al., 2007; Baldini et al., 2008), drip rate (e.g. Genty and Deflandre, 1998; Mattey et al., 2008), cave water and groundwater chemistry (e.g. Bar-Matthews et al., 1996; Fairchild et al., 2000; Musgrove and Banner, 2004; Spötl et al., 2005; Pape et al., 2010; Wong et al., 2011), and soil chemistry (e.g. Musgrove and Banner, 2004; Spötl et al., 2005).

Monitoring of modern cave environments provides a more in-depth understanding of the processes that affect speleothem paleoclimate proxies and allows for more accurate interpretations of speleothem time series. Relatively few monitoring studies, however, have been conducted in highly seasonal and semi-arid to arid environments (Bar-Matthews et al., 1996; Banner et al., 2007; Fernández-Cortes et al., 2008; Mattey et al., 2010; Pape et al., 2010) despite an increase in the number of speleothem-based paleoclimate records emerging from caves in these environments (e.g. Neff et al., 2001; Bar-Matthews et al., 2003; Auler et al., 2004; Genty et al., 2006; Vaks et al., 2006), especially from the western United States (Asmerom et al., 2007, 2010; Denniston et al., 2007; Oster et al., 2009; Wagner et al., 2010). Speleothems provide a crucial window into the history of these climatically sensitive regions where other high-resolution, precisely dated paleoclimate archives are generally scarce. Given the seasonal extremes of precipitation and temperature that can occur in these regions and the potential differences in behavior between these caves and those in humid and sub-humid environments, it is essential to understand how conditions within these caves respond to changing environmental conditions, and how this response is captured in speleothem calcite over time.

Here we discuss the results of a five-year (2006–2011) environmental monitoring program at Black Chasm Cavern, central Sierra Nevada foothills, California that provide insight into the behavior of a cave dominated by fracture flow and precipitation that is highly seasonally variable. Black Chasm provides a suitable setting for environmental monitoring given that it was opened for tourism only recently (2002) and neither ventilation nor water flow within the cave has been artificially modified. Observations drawn from measurements of cave air and drip water at Black Chasm confirm interpretations of speleothem proxy records from other caves developed within proximal marble units in the central Sierra Nevada foothills (e.g. Moaning Cave, Oster et al., 2009) where such monitoring programs cannot be implemented due to anthropogenic modification of the cave.

2. SITE DESCRIPTION

Black Chasm Cavern (38.43°N, 120.63°W, 676 m elevation) (Fig. 1a) is developed along vertical fractures within the Volcano carbonate unit, one of several discrete carbonate masses in the Sierra Nevada foothills. These lenticular masses of metamorphosed Permo-Carboniferous limestone and dolomite are intercalated with other metasedimentary and metavolcanic rocks of the Calaveras Complex (Bowen, 1973). Black Chasm Cavern was opened for tourism in 2002.

Black Chasm Cavern (BLC) consists mainly of two adjacent, vertically oriented chambers, the Colossal Room, and the Landmark Room (Fig. 1b). The Colossal Room is 30 m long by 5 m wide with a ceiling height that ranges from 1.8 to 31 m, and the Landmark Room is ~35 m long by 25 m wide with a ceiling height that ranges from 1.2 to 12.5 m. A series of smaller, short passages surround both rooms, and much of the lowest level of the cave is occupied by standing water (Barton, 2006). A moderately drained, immature soil (Inceptisol) (10–20 cm thick) overlies 10–20 m of marble overburden. During the winter (December through February) and spring (March through May), water drips within the cave near the entrance and at several locations within the Colossal Room. Most drip locations are dry during the late summer and autumn (August–November). Although the Landmark Room is currently dry year-round, it is well decorated with stalactites, helicmites, and stalagmites, and the passages that surround this room do experience seasonal drip water and possess many small, ephemeral lakes. Speleothem growth is ongoing in the Colossal Room and in smaller chambers.

The climate above BLC is characterized by cool, wet winters and warm, dry summers (Fig. 2). The University of California Integrated Pest Management Program has
operated a weather monitoring station 12.4 km from BLC at the Tiger Creek Pumping House (718 m elevation) since January 1951 (TC, Fig. 1a). From 1951 to 2010, this site experienced an average yearly rainfall of ~1100 mm. On average, 92% of this precipitation fell during the autumn, winter, and spring months (October–April), and 70% of this precipitation occurred between December and March. Although Black Chasm is not located within the area of the western United States that is significantly affected by the North American monsoon (Higgins et al., 1999), on average 2% of annual precipitation during this period occurred between July and September, suggesting either that some monsoon precipitation reaches as far west as the central Sierra Nevada foothill region or that this area is occasionally affected by localized summer convective storms. The University of California Integrated Pest Management Program also monitored soil temperature between May 1990 and May 2000 at Plymouth, an irrigated grassland site 18 km northwest of BLC at 472 m elevation (PL, Fig. 1a). Soil temperatures at Plymouth are likely slightly higher than those at BLC due to the location’s lower elevation and more open vegetation; the pattern of soil temperature variation through the year, however, should be similar (Fig. 2). Vegetation above BLC consists of C3 trees and shrubs including California Black Oak (Quercus kelloggii), Interior Live Oak (Quercus wislizenii), Ponderosa Pine (Pinus ponderosa), Gray Pine (Pinus sabiniana), Incense Cedar (Calocedrus decurrens), Pacific Madrone (Arbutus menziesii), Scrub Oak (Quercus berberidifolia), Poison Oak (Toxicodendron diversilobum), and Sword Fern (Polystichum munitum).

3. METHODS

Periodic measurements of cave air temperature, humidity, $pCO_2$, and rainwater and drip water chemistry in BLC were made between December 2006 and August 2011. Monthly (winter and spring) water measurements began in March 2007 and bi-weekly air and water measurements occurred between January 2009 and December 2010. Drip water flow within BLC slows substantially during the late summer and early autumn, so it was not possible to collect regular summer and early autumn water samples for all geochemical measurements, however a series of summer drip water samples was collected during July and August, 2011.

3.1. Air measurements

A HOBO® data logger was placed in a recess in the back of the cave and away from the tourism walkway (Location E, Fig. 1b) in October 2007. The logger performed hourly measurements of air temperature and relative humidity between this date and September 2009. Bi-weekly to monthly measurements of air temperature, humidity, and $pCO_2$ in BLC were taken between January 2009 and March 2011. Air measurements were made between 10 am and 2 pm. Air temperature, humidity, and $pCO_2$ measurements were conducted on the surface and at four platforms within the cave (Locations A–D, Fig. 1). Temperature and humidity measurements were made using a handheld Traceable hygrometer and thermometer. $PCO_2$ measurements were made using a Telaire portable $CO_2$ meter.

3.2. Rain and drip water measurements

Rainwater was collected above the cave in an open area approximately 8 m from the cave entrance. Following the International Atomic Energy Agency (IAEA) protocol, rainwater was collected in a 5 l LDPE carboy, and the bottom of the carboy was coated with approximately 0.5 cm of mineral oil to prevent evaporation. The collector was emptied bi-weekly to monthly and subsamples of the rainwater were removed for $\delta^{18}O$ and $\delta^2H$ analysis.

Drip water samples were collected ~30 m from the cave entrance at Location C (Fig. 1b). Location C is the most consistently wet area within the cave. Several drips fall from draperies approximately 4 m above the collection site. Drip water samples were collected during each cave visit from a group of drips that cover an area of approximately 0.1 m².
Water samples for stable isotope analysis ($\delta^{18}O$, $\delta^2H$) and $\delta^{13}C_{DIC}$ were collected in pre-cleaned 30 ml LDPE bottles, filled until there was no headspace and capped. Due to slow drip rates, it was not always possible to collect samples for $\delta^{13}C_{DIC}$ analysis, especially during the late summer and autumn. Sub-samples (10 ml) for $\delta^{13}C_{DIC}$ analysis were taken, treated with HgCl$_2$, and refrigerated for CO$_2$ extraction. Drip water samples for cation (Ca, Mg, Sr) analysis were collected biweekly to monthly between November 2008 and December 2010 and during July and August 2011. Samples were collected in filter flasks and acidified on site. Measurements of pH were taken prior to acidification using an Orion portable pH meter. Drip rate (in drips/minute) was measured at Location C using a stopwatch for one minute per visit from December 2006 to August 2011. In August 2009, a Stalagmite acoustic drip rate counter was installed in the cave. The counter downloaded drip count measurements every hour from August 2009 until April 2011, when it was removed from the cave.

Oxygen isotope measurements of drip waters were performed on a Finnigan MAT 251 IRMS, and hydrogen isotope measurements were performed on an IsoPrime IRMS. $\delta^{18}O$ and $\delta^2H$ values are reported relative to Standard Mean Ocean Water (V-SMOW) and $\delta^{13}C$ values are reported relative to Pee Dee Belemnite (V-PDB) using standard delta notation. Analytical precision for $\delta^{18}O$ is ±0.05‰ for $\delta^2H$ is ±1.0‰. The $\delta^{13}C_{DIC}$ for water samples collected after January 1, 2010 was analyzed at the Stanford School of Earth Sciences Stable Isotope Biogeochemistry Laboratory. During the analysis, 0.5 ml samples were reacted with 30 μL of phosphoric acid for 12 h at room temperature in 12 ml Labco vials. The headspace gas was analyzed in continuous-flow mode using a Finnigan GasBench-II headspace sampler interfaced with a Finnigan Delta+XL mass spectrometer. The precision, based on 6 repeat analyses of an in-house standard (absolute value 5.53‰, $\delta^{13}C$, calibrated against NBS-18, NBS-19, NBS-20, L-SVEC), was 0.22‰. The $\delta^{13}C_{DIC}$ for water samples collected prior to January 1, 2010 was analyzed in the Stable Isotope Laboratory, Department of Geology, UC Davis. For interlaboratory calibration of the $\delta^{13}C_{DIC}$ measurements, the Stanford in-house standard was dissolved in degassed, weakly acidified deionized water, and the $\delta^{13}C_{DIC}$ was determined by acid-stripping CO$_2$ from 5 ml of the solution under vacuum (using 105% orthophosphoric acid) and purifying the CO$_2$ cryogenically prior to mass spectrometric analysis. The average of three analyses of the Stanford in-house standard is $-5.53^{\circ}\text{C}_{\text{C}}$ overlapping with the mean value ($-5.53^{\circ}\text{C}_{\text{C}}$, n = 3) obtained on the powdered Stanford standard in the same laboratory. The long-term precision of the $\delta^{13}C$ was ±0.03‰.

Analysis to quantify concentrations of Ca, Mg, and Sr, in water samples collected between November 2008 and June 2010 were performed on an Agilent 7500 series quadrupole ICPMS at the Interdisciplinary Center for Plasma Mass Spectrometry, UC Davis. Analyses of waters collected between June 2010 and August 2011 were performed on a Thermo-Scientific X-series 2 quadrupole ICP-MS in the Stanford School of Earth Sciences Environmental Measurement 1 laboratory. External calibration standards were prepared using the Spex2a multi-element standard from 0.01 to 100 ppb for minor elements (Sr) and the Spex Instrument Check Standard 3 multi-element standard from 1000 to 50,000 ppb for major elements (Ca, Mg). Analytical error on cation analyses ranges from <1–3‰.

### 3.3. Statistical tests

In order to examine the sign and strength of relationships among drip water geochemical parameters and environmental factors, Pearson’s product moment correlation coefficients (“r”) and one-tailed tests of significance (“p”) between pairs of measured parameters were calculated using the statistics program R 2.14.1 (http://www.r-project.org/). Data with non-normal distributions were log10-transformed prior to statistical calculations in order to meet the assumptions of the Pearson’s correlation tests. Data with zero or negative values were rescaled with the addition of a constant prior to log-transformation. “p” values of <0.05 were accepted as statistically significant.
4. RESULTS

4.1. Cave air

Biweekly cave air temperature at the site of drip water collection (Location C) varied between 9.4 and 18.3 °C (mean 12.3 °C) between January 2009 and August 2011 (Fig. 3a), with peak temperatures occurring from the late spring to early autumn (May–October) and minimum temperatures occurring during the late autumn to early spring (November–April). Air temperature in BLC is most variable at Locations A and B (Fig. 1b), which are closest to the cave entrance. At Location A, temperature varied between 9.4 and 18.9 °C (mean 14.3 °C), and Location B temperature varied between 9.1 and 19.4 °C (mean 13.5 °C). Near the back of the cave (Location D, Fig. 1), temperature was less variable, ranging from 10.7 to 16.1 °C (mean 12.8 °C). At Location E, in the back of the cave removed from the tourist path, HOBO® logger data indicates temperature varied the least, ranging from 11.8 to 13.3 °C between October 2007 and September 2009.

Air $\text{pCO}_2$ at Location C in BLC is positively correlated with surface air temperature ($r = 0.74$, $p < 0.001$) and negatively correlated with surface air humidity ($r = -0.68$, $p < 0.001$), and is at its maximum in the summer (July–September), begins a steady decline in late October and early November, and is at its minimum in the late autumn (late November) through early spring (Fig. 3b). $\text{pCO}_2$ at Location C varied between 685 and 8020 ppm between January 2009 and March 2011. Cave air $\text{pCO}_2$ did not vary significantly (±150 ppm) between measurement locations within the cave with the exception of slightly lower values at Location A, which is closest to the cave entrance.

Relative humidity at the drip collection site (Location C on Fig. 1) varied between 67% and 100% (mean 94%) between January 2009 and August 2011 (Fig. 3c), with minimum values during the late spring to early autumn (May to October), and maximum values during the winter and early spring (December through March). Elsewhere in the cave, relative humidity varied between 64% and 100%, with the lowest values recorded at Locations A and B, and the highest values near the back of the cave (Location D). Surface air humidity is also higher during the wet winter months.

4.2. Water measurements

BLC rainwater $\delta^{18}$O varies between $-16.5\%$ and $-4.6\%$, $\delta^2$H varies between $-121.7\%$ and $-28.2\%$ (Fig. 4) (Table S1), and values for rainwater $\delta^{18}$O and $\delta^2$H fall on the global meteoric water line (Fig. 4a). Overall, Black Chasm rainwater stable isotopes are more negative during the winter and spring and less negative during the summer and autumn (Fig. 4), with the exception of individual Pineapple Express precipitation events (see Section 5.2). Rainwater $\delta^{18}$O values display a positive correlation with
average air temperature of rain days at the Tiger Creek Pumping House during the collection period \( \left( r = 0.42, \ p = 0.013 \right) \), and show no significant relationship with rainfall amount during the collection period \( \left( r = -0.09, \ p = 0.61 \right) \).

Drip rate at BLC is highly variable, and ranges from 1 drip per 10 min during the dry summer and autumn months to up to 60 drips/minute following winter and spring storm events. As drip rate is so variable and sensitive to storm events during the rainy season, we have normalized drip rates to each winter maximum for the purpose of reducing noise and allowing a more clear comparison of the timing of drip rate variability between years (Fig. 5). The pattern of this normalized drip rate closely resembles the pattern of monthly rainfall at the Tiger Creek Pumping House. Comparison of the Stalagmate logger data from BLC for October 2009 through 2010 with rainfall data demonstrates that drip response to the first storm of the season (e.g. October 24th, 2010, Fig. 6) was somewhat muted, leading to drip rates that were increased by \( \sim 1500 \) drip/day over pre-storm values. Storms later in the season (e.g. January) caused drip rate to increase by \( \sim 3000 \) drips/day, with rainfall events

Fig. 4. (a) Black Chasm rainwater data plotted with global meteoric water line, open circles indicate late spring–autumn rain events (May–November). Closed boxes are winter–early spring rain events (December–April). Stars indicate strong North Pacific-sourced storms. Open boxes indicate Pineapple Express events. (b) \( \delta^{18}\text{O} \) for rainwater from December 2006 to August 2011. Stars indicate strong North Pacific-sourced storms. Open boxes indicate Pineapple Express events. (c) Monthly rainfall totals (black boxes, dashed line), and (d) average monthly temperature (gray boxes and line) at Tiger Creek Pumping House for December 2006 through August 2011. Gray bars highlight winter months, December through February.
that lasted multiple days causing drip rates to increase by up to 5000 drips/day within 2–3 days of the onset of the storm event.

Drip water $\delta^{18}$O values range from $-11.0_{\text{iso}}$ to $-8.0_{\text{iso}}$ (Fig 7b) (Table S2), and $\delta^2$H values range from $-76.0_{\text{iso}}$ to $-54.7_{\text{iso}}$. Cave drip water $\delta^{18}$O and $\delta^2$H follow rainwater $\delta^{18}$O and $\delta^2$H and are generally more negative during the winter and spring months, and less negative during the summer and autumn months. The least negative drip water $\delta^{13}$C and $\delta^2$H values occur during the late autumn (October and November) following 6–8 months of minimal precipitation in the region. Drip water $\delta^{13}$C varies between $-12.5_{\text{iso}}$ and $-9.1_{\text{iso}}$ (Fig. 7c) (Table S2), and displays a strong negative correlation with cave air $\rho$CO$_2$ ($r = -0.81$, $p < 0.001$), and a positive correlation with drip rate ($r = 0.59$, $p = 0.01$) (Table 1).

Drip water Mg/Ca ranges from 0.026 to 0.054 and Sr/Ca ranges from 0.00089 to 0.0020 (Table S2). Drip waters collected during the summer (June–August) have the lowest Mg/Ca and Sr/Ca and highest Ca concentrations, while samples collected in the autumn (October–November) have the highest Mg/Ca and Sr/Ca and the lowest Ca concentrations (Fig. 8). For the entire drip water data set, Mg/Ca and Sr/Ca display significant negative correlations with $\rho$CO$_2$ ($r = -0.56$, $p = 0.0064$; and $r = -0.55$, $p = 0.0081$, respectively) (Table 1), and show no significant relationship with drip rate. However, when the drip water samples collected during the summer months are removed from the data set, the correlation between $\rho$CO$_2$ and Mg/Ca and Sr/Ca is greatly reduced ($r = -0.38$, $p = 0.108$; $r = -0.30$, $p = 0.214$, respectively) and negative correlations between
drip rate and Mg/Ca and Sr/Ca emerge ($r = \frac{1}{C_0^{0.60}}$, $p = 0.024$; $r = \frac{1}{C_0^{0.59}}$, $p = 0.0257$, respectively).

5. DISCUSSION

Prior calcite precipitation (PCP), which can occur within the epikarst between the soil zone and the cave passage or on the cave ceiling any time prior to drip water reaching a collection site or stalagmite, has been demonstrated to significantly influence speleothem proxy records such as carbon isotopic compositions and trace element contents (Baker et al., 1997; Fairchild et al., 2000; Johnson et al., 2006; Cruz et al., 2007; Fairchild and Treble, 2009; Oster et al., 2009; 2010). Variations in PCP have been linked to changes in cave water supply and water residence time in the epikarst, leading to the use of speleothem Mg/Ca and Sr/Ca as proxies of paleo-rainfall (Fairchild et al., 2000; Johnson et al., 2006; Karmann et al., 2007; McDonald et al., 2007). However, recent cave monitoring studies have documented the importance of cave ventilation in controlling calcite precipitation in both the epikarst and the cave, with enhanced calcite precipitation occurring during times when the cave is ventilated and cave air $pCO_2$ is reduced (Spötl et al., 2005; Banner et al., 2007; Kowalczyk and Froelich, 2010; Mattey et al., 2010; Frisia et al., 2011; Lambert and Aharon, 2011; Tremaine et al., 2011; Wong et al., 2011). Thus, it has been suggested that cave ventilation, rather than water supply to the cave, may provide the primary source of variability for speleothem growth rates as well as speleothem $\delta^{13}C$, Mg/Ca, and Sr/Ca.

While some cave monitoring studies have been carried out in monsoonal regions (Johnson et al., 2006) and subhumid regions that experience significant interannual precipitation variability (e.g. McDonald et al., 2007; Wong et al., 2011), to date, few studies have been conducted in semi-arid or arid regions that experience extreme seasonal drying (Bar-Matthews et al., 1996; Banner et al., 2007; Fernandez-Cortes et al., 2008; Mattey et al., 2010). Furthermore, the majority of cave monitoring studies are carried out in caves hosted in large limestone or dolomite units that have high

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Table 1

Pearson’s product moment correlation coefficients ($r$-values) for drip water $\delta^{13}C$, Mg/Ca, and Sr/Ca at BLC.

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<th>Parameter</th>
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<tr>
<td>Mg/Ca (all)</td>
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<tr>
<td>Sr/Ca (all)</td>
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</tr>
<tr>
<td>Mg/Ca (winter, spring, autumn)</td>
<td>$-0.38$</td>
<td>$-0.60^*$</td>
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<td>$-0.59^*$</td>
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$**$  $p < 0.001.$

$*$  $p < 0.05.$

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Fig. 7. From December 2006 through August 2011, (a) BLC rainwater $\delta^{18}O$. As in Fig. 4b, stars indicate strong North Pacific-sourced storms, and open boxes indicate Pineapple Express events. (b) drip water $\delta^{18}O$, (c) drip water $\delta^{13}C$, (d) drip water Mg/Ca, and (e) normalized drip rate at time of water collection. Gray bars highlight winter months, December through February.

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drip rate and Mg/Ca and Sr/Ca emerge ($r = -0.60$, $p = 0.024$; $r = -0.59$, $p = 0.0257$, respectively).
capacity to store and mix water before it enters the cave and in which fracture flow may not be the primary method of transport through the aquifer (e.g. Baker et al., 2000; Banner et al., 2007; Kowalczk and Froelich, 2010; Mattey et al., 2010; Boch et al., 2011). In these cave environments, variations in water supply may have little effect on drip rate within the cave, or may affect only drip sites that are fed by fracture flow and not others that are fed by more diffuse flow through the epikarst (Mattey et al., 2010; Wong et al., 2011). In contrast to these caves, Black Chasm (BLC) is located in a small (~2.5 km² of surface exposure) limestone unit that is bounded by surface drainages, and the cave is mainly fed by fracture flow through a relatively thin epikarst (~15–20 m). Additionally, the Mediterranean climate of the western Sierra Nevada foothills offers a stark seasonal contrast in precipitation, with virtually all of the recharge to the aquifer occurring during the winter and spring. Thus, BLC offers an environment in which to investigate processes of ventilation and drip rate variability and their relative effects on geochemical proxies of stalagmites that form in cave systems in which local recharge is episodic and transmitted rapidly into the cave.

5.1. Cave ventilation

Variations in air temperature, humidity, and pCO₂ in BLC suggest that the cave is most dynamically ventilated from the late autumn through the spring (November–May). During these months, variations in cave air temperature most closely follow changes in average surface air temperature and cave air pCO₂ fluctuates in response (Fig. 3). The monitoring site closest to the cave entrance experiences the largest fluctuations in temperature and humidity, and the overall lowest pCO₂ values (i.e. closest to surface air pCO₂) during this time period. As the spring months progress, temperature in the cave stabilizes, humidity declines, and pCO₂ begins to rise. CO₂ accumulates in the cave as it seeps into the cave from the soil zone and is degassed from drip waters, due to a lack of exchange between cave air and low pCO₂ surface air (Banner et al., 2007). Cave air pCO₂ reaches a maximum in the summer and early autumn (July–September). While pCO₂ within BLC may be modified by tourism, this modification should occur to some degree during all seasons, as the cave is open to tourism year-round. However, increased visitation during the summer could lead to increased summer pCO₂. As a comparison, a brief (4 month) pCO₂ time series collected in a nearby (10 km W of Black Chasm), non-commercial cave using a CO₂meter data logger displays a similar summer increase in pCO₂ (Fig. 3b).

These trends support increased exchange between cave and surface air during the cold winter months into the spring due to sinking of cold, low pCO₂ surface air into the cave. This ventilation continues through the spring, as the rainy season is ending but surface temperatures are still cool. Ventilation slows in early summer (June), and cave air pCO₂ builds within the cave to a summer maximum (July–September). The winter ventilation observed at BLC is consistent with observed seasonal changes in cave ventilation and pCO₂ in many monitored caves including those in the southern United States (Banner et al., 2007; Kowalczk and Froelich, 2010; Lambert and Aharon, 2011), Austria (Spötl et al., 2005; Boch et al., 2011), and Italy (Frisia et al., 2011) that are driven by changes in the relative density of surface and cave air that occur due to seasonal temperature variability.

5.2. Rainwater variability

The observed trend of more negative δ¹⁸O and δ²H values of BLC rainwater during the colder and wetter winter and early spring, and less negative values during the warmer and drier late spring through late autumn/early winter suggests that rainwater isotopes are influenced by seasonal
variation in surface air temperature or rainfall amount (Fig 4). However, the direct relationships between rain $\delta^{18}$O, air temperature on rainy days, and rainfall amount during the collection period suggest that temperature exerts a stronger control on rain isotope values than does rainfall amount. These observations agree with a similar study that suggests temperature rather than rainfall amount controls rainfall $\delta^{18}$O in the Pacific Northwest (Ersek et al., 2010). However, in contrast to observations from that region, and in agreement with a study of rainfall $\delta^{18}$O in southern California (Berkelhammer et al., 2012), the isotopic composition of rainwater from BLC also appears to vary with water vapor source, with North Pacific-sourced vapor associated with more depleted rainwater isotope values, and tropical-sourced water vapor associated with less depleted values. For example, several strong storms sourced in the far North Pacific passed over Northern California during the study interval. The rainwater taken from the collector for these periods had the most depleted $\delta^{18}$O and $\delta^2$H values measured in this study (Fig. 4). In addition, two Pineapple Express storms passed over the cave during the spring of 2009 and 2010. Pineapple Express events occur when the jet stream taps into warm, wet air from the tropical Pacific and funnels it up to the west coast of North America. These storms are uncharacteristically warm and wet for California winter precipitation (Dettinger, 2004). Rainwater collected at BLC after these Pineapple Express events had $\delta^{18}$O and $\delta^2$H values ($-8.2^\circ_C$, $-51.4^\circ_C$, and $-4.6^\circ_C$, $-28.2^\circ_C$, respectively) that were less depleted than other early spring precipitation events, as would be expected of precipitation derived from a more tropical source.

Back-calculated particle trajectories for rainfall events over the central Sierra Nevada foothills using the NOAA Hybrid Single Particle Lagrangian Integrated Trajectory Model (HYSPLIT) and global reanalysis data indicate that the stable isotopic composition of rainfall in this region is influenced by water vapor source. These models, each run at 1500 m above ground level, the approximate height of the 850 hPa level from where rain is expected to originate (http://www.ready.arl.noaa.gov/HYSPLIT.php), simulate the trajectory that an air parcel followed over 72 h before reaching BLC on the day of a particular storm event. Particle trajectories for December 28th, 2010 (Fig. 9a), when rain $\delta^{18}$O was very depleted, suggests a source in the Bering Sea, while the trajectory for March 3rd, 2009 (Fig 9b), when rain $\delta^{18}$O was the most enriched of the four events shown, shows the trajectories dipping southward before reaching California. Trajectories for the rainfall events of December 17th, 2010 (Fig. 9c) and March 20th, 2011 (Fig. 9d) are similar to one another and remain between 30 and 40°N. The rainfall event of December 17th (Fig. 9c) occurred just eleven days prior to the event pictured in Fig. 9a, and average air temperatures for these two days were similar (within 1°C). Thus the observed difference in $\delta^{18}$O between the December 17th and 28th events most likely reflects different vapor sources. However, different $\delta^{18}$O values between the December 17th, 2010 (Fig. 9c) and March 20th, 2011 (Fig. 9d) events could best be explained by variables other than moisture source, as the particle trajectories for these events are similar. Indeed the average air temperature at the weather station was 3.5°C colder on March 20th 2011 than on December 17th, 2010. Thus, on event to
monthly timescales, it appears that both temperature and storm source influence rainwater isotopes at BLC.

5.3. Drip water variability

Drip rate in Black Chasm is strongly tied to rainfall above the cave, with drip rates near zero (~1 drip/10 min) in the dry summer months due to the lack of water recharge from the surface. The first rainfall event of the wet season, usually in October or November, has little effect on cave drip rate, suggesting that the majority of the water from these events is absorbed by the low moisture soil zone, and does not infiltrate into the epikarst. Rainfall events that occur later in the wet season, after the soil has been wetted, trigger up to fourfold increases in drip rate that last for several days, depending on the duration of the rain event (Fig. 6).

Shifts in drip water $\delta^{18}O$ and $\delta^2H$ tend to lag behind analogous changes in rainwater $\delta^{18}O$ and $\delta^2H$ by up to three weeks during the rainiest months, indicating some storage of recharge water in the epikarst. Furthermore, lower variability of drip water $\delta^{18}O$ and $\delta^2H$ relative to that of the rainwater suggests there is some mixing of rainwater recharge with the older stored waters within the epikarst. Such rainwater storage, however, likely consists of waters recharged during each rainy season given that drip water ceases to enter the cave during the late summer months. The storage and mixing of rainwater in the epikarst within each season, however, appears to dampen the effect of singular variations in water vapor source (i.e. North Pacific versus Pineapple Express storms, compare Figs. 4b and 7a and b). Sustained differences in the stable isotopic composition of autumn and winter-spring rainfall due to temperature have a greater influence on drip-water stable isotopes. This would suggest that, in order for changes in vapor source to have a significant long-term effect on drip-water and in turn be recorded by speleothem stable isotope compositions, the shifts would need to be sustained at least throughout a rainy season, and possibly for several years depending on stalagmite growth rate.

BLC drip water $\delta^{13}C$ ($\sim -8.6_{\text{SMOW}}$ to $-12.5_{\text{SMOW}}$) is less negative than the $\delta^{13}C$ composition of DIC anticipated for drip waters in fully open caves systems overlain by C$_3$ vegetation and saturated with respect to CaCO$_3$ ($-14_{\text{SMOW}}$ to $-18_{\text{SMOW}}$) (cf. Hendy, 1971; Dulinski and Rozanski, 1990). In a fully open cave system, the cave seepage water is always in contact with an excess of soil CO$_2$. Less negative $\delta^{13}C$ values of DIC are expected for closed cave systems ($-11_{\text{SMOW}}$) where seepage waters are isolated from soil CO$_2$ after entering the epikarst (Hendy, 1971). Observation of modern cave systems has shown, however, that most cave systems are at least partially open (McDermott, 2004). It is possible, therefore, that the overall $\delta^{13}C$-enriched BLC drip waters can be explained by a higher percentage of carbon contribution from the host limestone due to dissolution that occurs in partially rather than fully open system conditions. This scenario, however, cannot explain the highest drip water $\delta^{13}C$ values that are observed at BLC ($-8.6_{\text{SMOW}}$ to $-10.9_{\text{SMOW}}$), and geochemical modeling of C isotope systematics for a proximal cave (Moaning Cave), which is developed in marble of similar age and depositional and diagenetic history within the Calaveras Formation, suggests that the contribution of host rock dissolution to DIC $\delta^{13}C$ variability is minimal (Oster et al., 2010). Thus, it is unlikely that drip-water $\delta^{13}C$ at BLC is controlled by dissolution of the host limestone. We thus conclude that the $\delta^{13}C$-enriched drip waters at BLC most likely reflect a combination of variations in soil respiration and precipitation of calcium carbonate in the epikarst, upflow from the drip collection site.

The strong positive correlation of Mg/Ca with Sr/Ca ($r = 0.96, p < 0.001$) further argues for a PCP influence on the trace element contents of drip waters and in turn stalagmite calcite. If differential dissolution of calcite and dolomite provided a primary control on elemental concentrations, Mg/Ca should be negatively correlated with Sr/Ca given that dolomite has lower concentrations of Sr than calcite (Fairchild et al., 2000; Hellstrom and McCulloch, 2000). Rather, the positive correlation between drip water Mg/Ca and Sr/Ca and the relationship between Mg/Ca and Sr/Ca and calcium concentration (Fig. 8), as well as a negative correlation between drip water $\delta^{13}C$ and calcium concentration ($r = -0.83, p = 0$) suggests that drip water $\delta^{13}C$, and Mg and Sr concentrations are best explained by degassing and precipitation of calcite upflow from the drip site. During degassing and calcite precipitation, the $\delta^{13}C$ value of the HCO$_3$ reservoir follows Rayleigh-type enrichment in $\delta^{13}C$ with progressive degassing, as the loss of isotopically light $^{12}C$ from degassing cannot be offset by the loss of isotopically heavier $^{13}C$ through calcite precipitation (Mickler et al., 2004), assuming a one for one loss of CO$_2$ and CaCO$_3$. Likewise, as calcite is precipitated from drip water, the remaining water possesses higher concentrations of Mg and Sr, as the distribution coefficients for these elements in calcite are much less than 1 (Lorens, 1981). Prior calcite precipitation (PCP) in a cave will increase due to increased degassing that occurs when slower seepage rates within the epikarst allow saturation to be reached earlier in the water travel path (Fairchild et al., 2000) and when seepage waters encounter air pockets of lower $p$CO$_2$.

In order to further evaluate this hypothesis that PCP is the predominant influence on the observed trace element compositions of drip waters at BLC, we modeled PCP in a CaCO$_3$ saturated drip water, with Mg and Sr contents equal to the maximum measured values of BLC drip waters using Geochemist’s Work Bench (Fig. 8). The model curves were created by moving water from an environment characterized by a CO$_2$ fugacity representing the soil environment ($10^{-2}$) to an environment characterized by a CO$_2$ fugacity representing the cave environment ($10^{-3.5}$). All of the measured drip waters at BLC plot on or proximal to these model curves, further supporting the hypothesis that PCP is the primary control on trace element variation in BLC drip waters.

The significant negative correlation between drip water Mg/Ca and Sr/Ca and cave air $p$CO$_2$ year-round ($r = -0.56, p = 0.0064$; and $r = -0.55, p = 0.0081$, respectively) reflects the overall influence of $p$CO$_2$ driven by seasonal cave ventilation on PCP in this cave system. However, the negative correlation between drip rate, Mg/Ca and
Sr/Ca during the winter, spring, and autumn ($r = -0.62$, $p = 0.0187$; $r = -0.60$, $p = 0.0235$, respectively) supports the hypothesis that water supply is the predominant influence on PCP when overall cave $pCO_2$ is low. Thus, elemental concentrations in Black Chasm drip waters indicate that both $pCO_2$ and water supply influence prior calcite precipitation in this cave system, with drip rates becoming more important during periods of lower cave air $pCO_2$. Given that speleothem precipitation in this cave system most likely occurs during the low $pCO_2$ months we conclude that variations in Mg/Ca and Sr/Ca of BLC stalagmites will primarily record changes in water supply to the cave.

Due to extremely slow drip rates, it was not possible to sample waters for $\delta^{13}C_{DIC}$ during the late summer and fall (between late August and December). Thus, it is not possible to fully assess the relationship between $\delta^{13}C$, drip rates, and $pCO_2$ variability at BLC. The strong negative correlation between $\delta^{13}C$ and $pCO_2$ in the winter through early summer ($r = -0.81$, $p < 0.001$) argues that cave air $pCO_2$ provides the dominant control on drip water $\delta^{13}C$ during the wetter part of the year. Drip water $\delta^{13}C$ should also be influenced by soil respiration, with periods of increased respiration leading to more negative drip water $\delta^{13}C$. Soil respiration in arid and semi-arid environments is controlled by both temperature and soil moisture, and soil CO$_2$ concentrations are generally at a maximum during the spring and early summer when soil temperatures are rising but soils are still moist (Amundson et al., 1988; Quade et al., 1989; Terhune and Harden, 1991). Although soil CO$_2$ was not measured during this study, it is likely that soil respiration rates at BLC reach their maximum during the spring and early summer as soil temperatures increase (Fig. 2) but before soil moisture levels substantially decrease. Indeed, there is a significant negative correlation between drip water $\delta^{13}C$ and surface air temperature ($r = -0.62$, $p = 0.0046$). Thus, it is likely that the minimum $\delta^{13}C$ values for spring and summer drip waters reflect enhanced soil respiration, in addition to reduced PCP, during this interval. The positive correlation observed between $\delta^{13}C$ values and drip rate ($r = 0.59$, $p = 0.05$) is likely a result of the strongly seasonal nature of drip rate variability in this cave, given that both $\delta^{13}C$ values and drip rate are negatively correlated with surface air temperatures. If it were feasible to measure $\delta^{13}C$ during the driest season above the cave, it is possible that a relationship between drip rate and $\delta^{13}C$ would emerge that is parallel to what is observed with Mg/Ca and Sr/Ca, where low fall drip rates are associated with drip waters with less negative $\delta^{13}C$ values reflecting the combined effects of reduced soil respiration and increased PCP relative to summer values.

Observations of cave air and drip water chemistry at BLC illustrate how ventilation and drip rate variability should influence geochemical proxies of stalagmites that form in cave systems in winter-precipitation dominated climates where local recharge is episodic and transmitted rapidly into the cave. Variations in the $\delta^{18}O$ recorded by a stalagmite from BLC should reflect changes in surface air temperature and/or sustained shifts in moisture source, while $\delta^{13}C$ changes should reflect a combination of PCP variability and soil respiration. As stalagmite precipitation in the cave is most likely to occur during periods of lower cave air $pCO_2$, speleothem paleoclimate archives will most likely record autumn through spring climates, rather than annually averaged climate variability, and variations in Mg/Ca and Sr/Ca recorded by a stalagmite are likely to reflect changes in drip rate, and therefore changes in water availability. These conclusions may apply to other central Sierra Nevada foothills caves given their similar geologic setting, paragenetic history, and environmental and climatic conditions.

5.4. Implications for paleoclimate studies

Cave air ventilation and $pCO_2$ have been shown to exert a dominant control on calcite precipitation rates in caves in humid to sub-humid environments, and in those where significant water storage occurs in the epikarst (Banner et al., 2007; Baldini et al., 2008; Kowalczyk and Froelich, 2010; Mattey et al., 2010; Frisia et al., 2011; Lambert and Aharon, 2011; Tremaine et al., 2011; Wong et al., 2011). In these systems, enhanced calcite deposition in the cave and in the epikarst occurs when the difference between the cave air $pCO_2$ and drip water $pCO_2$ is greatest such as when cave air $pCO_2$ is low due to increased exchange between the cave and surface atmosphere (Banner et al., 2007), and/or when drip water $pCO_2$ is high as a result of increased soil $pCO_2$ during the growing season (Genty et al., 2001b; Kowalczyk and Froelich, 2010). Seasonal changes in water supply and drip rate have been shown to exert a secondary control on calcite deposition rates within caves, mainly when comparing calcite precipitation under different drips within the same cave (Banner et al., 2007; Mattey et al., 2010).

Chemical variations of drip water at BLC indicate that, while both $pCO_2$ and drip rate variability have a significant influence on PCP, drip rate, and therefore water supply, provides a dominant influence during periods of low $pCO_2$ when stalagmites are likely to grow, and thus we predict that speleothem proxy records that are sensitive to PCP should respond primarily to variations in water supply to the cave. The observed relationships between stable isotopes and trace elements in BLC rainwater and drip water on a seasonal-scale provide a high-resolution analogue for the decadal to multi-centennial variability inferred from speleothem proxy records from another central Sierran foothill cave, Moaning Cave (Oster et al., 2009). Although BLC is open to tourism, ventilation and water flow within the cave have not been modified. This is not the case for Moaning Cave, 42 km southeast of BLC (MC: Fig. 1a), where spring water from below the cave is artificially pumped through the cave during the dry summer months precluding monitoring of the cave and surface conditions. Like BLC, Moaning Cave (MC) is a vertically oriented cave developed along fractures in a metamorphosed limestone unit within the Calaveras Formation. MC is located at a slightly lower elevation than BLC (520 m a.s.l.) and receives slightly less yearly rainfall (~850 mm/year from 1951 to the present compared to ~1100 mm/year at BLC). Similar to BLC, 75–100% of the yearly rainfall above MC occurs during the winter and early spring.
Comparison of the seasonal-scale variability observed in BLC drip water with the decadal to multi-centennial variability noted in the deglacial to early Holocene proxy records from MC suggests that similar environmental
Processes drive changes on both timescales. Crystal fabrics indicative of continuous precipitation in the Moaning Cave speleothem are associated with lower [Mg] and [Sr] and more negative $\delta^{13}$C and $\delta^{18}$O values. U-series ages indicate these intervals of stalagmite calcite precipitated during Northern Hemisphere cold periods (Fig. 10). Conversely, crystal fabrics in the MC stalagmite that indicate intermittent growth are associated with higher [Mg] and [Sr], and less negative $\delta^{13}$C and $\delta^{18}$O values. These layers of calcite precipitated during Northern Hemisphere warm periods between 16.5 and 8.7 ka (Oster et al., 2009). Enhanced cave ventilation leading to increased calcite precipitation in the epikarst (PCP) on long timescales should be driven by large seasonal temperature differences above the cave (Banner et al., 2007). The largest difference between winter and summer insolation at the Moaning Cave latitude during the period of 16.5–8.7 ka occurs from ~12 to 10 ka (Fig. 10a), from which we infer that the largest seasonal temperature differences above the cave occurred during this interval. The smallest difference between winter and summer insolation occurs from 16 to 14 ka. Therefore, if variations in cave $p$CO$_2$ due to changes in cave ventilation were the only control on PCP in the Moaning Cave system, we would expect to see the highest speleothem Mg and Sr concentrations during the period of enhanced seasonality (~12–10 ka). While this interval does correspond to crystal fabrics that are indicative of continuous stalagmite precipitation, it also corresponds to some of the lowest observed Mg and Sr concentrations. Crystal fabrics indicative of intermittent calcite precipitation as well as higher stalagmite Mg and Sr concentrations and $\delta^{13}$C values are observed when seasonality is reduced (15–14 ka).

The documented relationship between drip rates and trace element ratios of BLC drip waters during periods of low $p$CO$_2$ suggests that, for intervals of stalagmite growth, higher Mg and Sr concentrations in the MC stalagmite should indicate periods of relatively slower drip rate and lower water availability. Taken together, these results suggest that Northern Hemisphere warm periods recorded in the MC stalagmite were characterized by little or no stalagmite growth, punctuated by intervals where the cave was ventilated enough to allow some calcite precipitation. Yet even during these growth intervals, drip rates remained low enough to promote high amounts of PCP in the epikarst, leading to higher concentrations of Mg and Sr in the stalagmite. Conversely, Northern Hemisphere cold periods were characterized by ventilation of the cave that in turn promoted continuous stalagmite growth. Given that crystal fabrics indicate a relatively consistent water supply to the drip site and growing MC stalagmite during these periods, the lower Mg and Sr concentrations are interpreted to record decreased PCP in the epikarst, even during periods of enhanced ventilation, due to high water availability and drip rates.

The $\delta^{13}$C values of BLC drip waters reflect changes in PCP, and possibly also variations in soil respiration rate, as the lowest $\delta^{13}$C values occur during the spring and early summer when soil respiration should be at a maximum. Less negative $\delta^{13}$C values in the MC stalagmite during Northern Hemisphere warm periods could reflect more PCP. If variations in cave air $p$CO$_2$ were the dominant control on stalagmite $\delta^{13}$C, then, as with the trace element records, we would expect to see the highest $\delta^{13}$C values during the periods of increased seasonality. However, we see the lowest stalagmite $\delta^{13}$C values between 12 and 10 ka, when summer insolation is at its peak, and seasonality is enhanced, indicating that cave ventilation is not the primary control on $\delta^{13}$C in the Moaning Cave record. Rather, this relationship suggests that increased water supply to the cave during overall cold periods as is suggested by the stalagmite trace element records, could have lead to decreased $\delta^{13}$C values in stalagmite calcite. This decrease in the $\delta^{13}$C values of stalagmite calcite may have been enhanced by increased soil respiration due to increased water supply.

In comparison with observations from BLC rainwater and drip water, less negative $\delta^{18}$O values in the MC speleothem during Northern Hemisphere warm periods are consistent with warmer temperatures and potentially an enhanced contribution of water vapor from a more subtropical source, possibly from more frequent Pineapple Express events. Recent climate models investigating the effects of future global warming on the occurrence of Pineapple Express events in the western United States suggest that the frequency of years with a larger number of these storms increases with global warming (Dettinger, 2011). It is also possible that moisture transport from the tropics to the mid-latitudes would be enhanced during warmer periods due to the increased water-holding capacity of warmer tropical air, leading to increased $\delta^{18}$O values of precipitation in this region (Schneider et al., 2010; Berkelhammer et al., 2012). Thus a change in the relative contribution of subtropical moisture reaching the central Sierra Nevada during past warmings of the last deglaciation could have amplified the increase in precipitation $\delta^{18}$O driven by increased temperatures in the region.

Observations of drip water chemistry and cave ventilation at Black Chasm provide a useful modern analogue for the interpretation of paleoclimate proxy records from nearby Moaning Cave, and will be essential for the interpretation of speleothem proxy records from other caves in the central Sierra Nevada and similar semi-arid and arid regions that experience large seasonal precipitation variability. The BLC drip water and cave air $p$CO$_2$ variability, coupled with the MC paleoclimate records, suggest that in this region, evidence of variation in PCP can provide a proxy for changes in drip rate, and thus in rainfall amount, above the cave. This is especially true in caves such as those in the Sierra Nevada foothills in which ground water flow is dominated by fracture-flow, and long-term groundwater storage is minimal.

6. CONCLUSIONS

Regular environmental monitoring at Black Chasm Cavern in the Sierra Nevada demonstrates that the cave is ventilated in the winter months, when cold, dense surface air sinks into the cave. Large seasonal rainfall variability causes cave drip-water to slow substantially during the late summer and autumn months, increase significantly during
the wet winter and spring, and respond within hours to storm events during the height of the rainy season. Rainwater $\delta^{18}O$ and $\delta^2H$ are controlled by a combination of surface air temperature and water vapor source. Due to mixing of recharge from different events in the epikarst, seasonal changes in rainwater $\delta^{18}O$ have a larger influence on drip water $\delta^{18}O$ than single storm events with distinct isotopic signatures. However, moisture source could provide a more significant influence on drip water values if vapor sources shift more permanently. Cave drip-water Mg/Ca and Sr/Ca are highest in the autumn when drip rates and cave air $pCO_2$ are low, and lowest during the summer when drip rates are low, but cave air $pCO_2$ is at its highest. Drip water Mg/Ca and Sr/Ca display a significant correlation with $pCO_2$ values throughout the year, but this relationship becomes less significant than the correlation with drip rate if summer data is excluded. This suggests that drip rate may be the primary control on drip water elemental ratios and prior calcite precipitation during the seasons when speleothem growth is most likely to occur. Drip water $\delta^{13}C$ values are most negative during the late spring and early summer and likely reflect both prior calcite precipitation and soil respiration rates.

The isotopic and chemical variability observed in Black Chasm drip waters is consistent with interpretations of speleothem paleoclimate proxy records from a nearby cave where such monitoring is not feasible and suggests that speleothems from these caves can be used to investigate past precipitation variability. While a complete understanding of speleothem proxy records is limited by the inability to observe cave systems under different boundary conditions, such as those that existed during the last glacial period, careful and long-term monitoring of modern cave environments offers the best approach to accurately interpreting these records. Observations of the modern cave environment at Black Chasm provide an analogue for the interpretation of speleothem proxy records from similar fractured-flow dominated cave environments characterized by large seasonal differences in rainfall and minimal long-term ground-water storage and should be useful as more speleothem records from arid regions are developed. Continued cave monitoring work in this and other caves in this region will better constrain the effects of longer-term environmental changes on cave air circulation and water chemistry as well as the degree of variability in the response among caves in this region to these changes. Paleoclimate records from caves in these environments will shed valuable light on how rainfall varies with climate change in regions with highly sensitive water resources.

ACKNOWLEDGEMENTS

We thank Steven Fairchild, President of the Sierra Nevada Recreation Corporation, and Shaundy Farley, manager of Black Chasm Cavern for providing access to the cave. We also thank E. Kleber, M. Kelley, and N. Marks for field assistance. P. Green and Guangchao Li assisted with cation measurements, and H. Spero, D. Winter and. P. Blisnuiu assisted with stable isotope measurements. This work benefitted from comments by S. Zimmerman and three anonymous reviewers. This work was supported by a Cave Research Foundation grant to Jessica Oster and NSF grant NSF-ATM0823656 to Isabel P. Montañez.

APPENDIX A. SUPPLEMENTARY DATA

Supplementary data associated with this article can be found, in the online version, at http://dx.doi.org/10.1016/j.gca.2012.05.027.

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Associate editor: Sidney Hemming