AN ABSTRACT OF THE DISSERTATION OF

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Abstract approved:

__________________________________________________________
Peter U. Clark

This dissertation presents the results of three studies that assessed climate variability on short and long timescales in western United States. The growth of carbonate formations in caves (speleothems) is used to infer the timing and amplitude of past climate variability. We first assess the controls on speleothem growth for the past 380 000 years by combining high-density U-Th dating with a regional climate model and an energy balance model. The majority of speleothem growth occurred overwhelmingly during interglacial periods and glacial periods were characterized by little or no growth. We assess the mechanisms responsible for growth cessation during the peak of the last glaciation (LGM), 21 000 years ago, with a regional climate model and determine that a combination of drier LGM conditions with a change in the seasonal cycle of the surface water balance provides a feedback mechanism that limited recharge to the deeper soil and reduced or eliminated drip water in the cave. The energy balance model supports this mechanism by indicating that cave temperatures did not drop below the freezing point of water at any time during the last glaciation. We then evaluate the climatic significance of stable isotopes of oxygen in rainwater collected in southwestern Oregon in order to accurately interpret the isotopic
record of speleothems. We establish that temperature plays an important role in controlling the distribution of oxygen isotopes in rainwater and therefore speleothems in southwestern Oregon can preserve a record of temperature changes through time. Finally, we present a high-resolution paleoclimate record using speleothems from a cave in Oregon that grew during the last 9000 years and during the middle of the last interglaciation. We find the winter temperatures were sensitive to winter solar insolation and varied in a quasi-cyclical pattern on millennial timescales. This pattern was characteristic of both regional and hemispheric variability as indicated by similarities of our record with other regional and hemispheric climate reconstructions. It appears that higher climate variability in the past was associated with warming trends, with the implication that global warming may make the climate more unstable in the future. Climate variability during the last interglacial period was characterized by different scales of millennial and centennial climate variability compared with the Holocene.
Past Climate Variability in Southwestern Oregon and Relationships with Regional and Hemispheric Climate

by

Vasile Ersek

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I understand that my thesis will become part of the permanent collection of Oregon State University libraries. My signature below authorizes release of my thesis to any reader upon request.

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Vasile Ersek, Author
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CONTRIBUTION OF AUTHORS

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I dedicate this work to Carmen, who has been a constant source of strength and inspiration, and to our son, Vlad, who brought so much joy into our lives.
Past climate variability in southwestern Oregon and relationships with regional and hemispheric climate

Chapter 1

Introduction

1.1 Purpose

The purpose of this dissertation is to describe and understand climate variability in western North America on both short and long timescales. While the current interglacial period has often been viewed as a climatically stable period in Earth’s history, it has become increasingly apparent that climate changes during the last ~10 000 years (the Holocene Epoch) were of sufficient magnitude to affect human civilization. Much remains unknown, however, about the spatial and temporal character of these changes.

The Pacific Ocean variability has an important effect on Earth’s climate, yet the past behavior of the North Pacific and its influence on adjacent land masses is not well constrained. This study makes a contribution towards a better understanding of the land-ocean interactions in the eastern North Pacific by applying a combination of high-resolution stable isotope measurements with precise U-Th dating and with insight into the controls of oxygen isotopes in modern rainwater at Oregon Caves National Monument in southwestern Oregon.

1.2 Summary of chapters

Chapter 2 discusses the growth of speleothems at Oregon Caves National Monument (OCNM) which is sensitive to changes in environmental conditions at the surface (temperature, precipitation and vegetation) and can provide useful
paleoclimatic and paleoenvironmental information. The results of this study indicate that most speleothem growth occurred during interglacial periods, whereas little speleothem growth occurred during glacial periods. Through a combination of two modeling approaches the potential environmental controls on speleothem growth are evaluated and the results indicate that the influence of the Laurentide and Cordilleran ice sheets on atmospheric circulation induced substantial changes in water balance in southwestern Oregon and affected speleothem growth.

In an effort to understand the controls of oxygen isotopes in speleothems, chapter 3 is a study of the relationship between climate variables and oxygen isotopes in rainwater collected at OCNM. Storm trajectories for the past 48 hours before arrival were calculated using a Lagrangian model. The storms span over ~30° of latitude and a large gradient in sea surface temperatures and δ¹⁸O, but no significant relationships are found between trajectory origin or path and the isotopic composition of rainfall events sampled on synoptic timescales. Storm and local temperatures explain ~30% of oxygen isotope variability and the slope between temperature and δ¹⁸O is 0.7 ‰/°C. Changes in specific humidity along the storm track suggest that continuous condensation and entrainment of moisture along the storm path may also influence the δ¹⁸O composition of rainfall events in southwestern Oregon.

Chapter 3 presents a high-resolution record of climate variability during winter in western North America using speleothems that grew during the Holocene and the last interglacial period. We use speleothem δ¹⁸O as an indicator of temperature change, while the δ¹³C is interpreted as reflecting mainly changes in precipitation and soil biomass. Holocene winter climate at our site was influenced by long-term changes in solar insolation, but also exhibits shorter, millennial and centennial climate variability. The phase relationship between carbon and oxygen isotopes in speleothems suggests a delayed response of vegetation to climate change at the millennial time band, and synchronized changes in temperature and precipitation on multi-centennial timescales. We identify maximum climate variability corresponding with the peak rate of warmings, which implies that global warming may cause higher climate instability in the future. Compared with the Holocene, the last interglacial
period was characterized by similar average temperatures, but was possibly wetter and/or had higher soil productivity.
Chapter 2

Environmental influences on speleothem growth in southwestern Oregon during the last 380 000 years

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2.1 Abstract

The growth of carbonate formations in caves (speleothems) is sensitive to changes in environmental conditions at the surface (temperature, precipitation and vegetation) and can provide useful paleoclimatic and paleoenvironmental information. We use 73 $^{230}$Th dates from speleothems collected from a cave in southwestern Oregon (USA) to constrain speleothem growth for the past 380,000 years. Most speleothem growth occurred during interglacial periods, whereas little speleothem growth occurred during glacial periods. We use two new modeling approaches to evaluate potential environmental controls on speleothem growth: i) a one-dimensional thermal advection-diffusion model to estimate cave temperatures during the last glacial cycle, and ii) a regional climate model simulation for the Last Glacial Maximum (21,000 years before present) that evaluates a range of potential controls on speleothem growth under peak glacial conditions. The two models are mutually consistent in indicating that permafrost formation did not influence speleothem growth during glacial periods. Instead, the regional climate model simulation combined with proxy data suggest that the influence of the Laurentide and Cordilleran ice sheets on atmospheric circulation induced substantial changes in water balance in southwestern Oregon and affected speleothem growth.

2.2 Introduction

Speleothems are important archives of terrestrial paleoclimate because they can preserve high temporal resolution environmental records derived from chemical and isotopic proxies, and can be precisely dated using uranium-series disequilibrium (McDermott, 2004; Richards and Dorale, 2003). Changes in speleothem growth can also provide an important constraint on paleoenvironmental conditions because speleothem formation requires water availability and elevated CO$_2$ levels in the soil zone above the cave, which in turn are dependent on local water balance, biomass, and temperature (Dreybrodt 1980, 1999; Dorr and Munnich, 1989; Fairchild et al., 2006). Accordingly, the growth of many speleothems at nonglaciated mid- and high-latitudes slowed or ceased altogether at various times during the Quaternary because of some
combination of increase in the areal extent of permafrost or glaciers, a decrease in soil pCO$_2$, or changes in water availability (Baker et al., 1995; Brook et al., 2006; Genty et al., 2003; Gordon et al., 1989; Harmon et al., 1977; Spötl et al., 2002; Thompson et al., 1974; Turgeon and Lundberg, 2001; 2004). Distinguishing among these potential controls from information on growth changes alone, however, is generally unconstrained, requiring additional proxy information or modeling.

Previous investigations of speleothems at Oregon Caves National Monument (OCNM) in southwestern Oregon showed that they are suitable for paleoclimatic studies. Turgeon and Lundberg (2001; 2004) studied two flowstones and one 5-cm long stalagmite from OCNM. A chronology established by 14 thermal ionization mass spectrometry (TIMS) $^{230}$Th ages suggested that growth of these three samples occurred primarily during early to mid-interglacial periods, with one interval of deposition during marine isotope stage (MIS) 10. They proposed that growth ceased during glacial intervals because the glaciation limit was lower, favoring permafrost conditions at the cave site. Vacco et al. (2005) studied the $\delta^{18}$O record from a 26.5 cm long stalagmite from OCNM and identified a cold interval that is synchronous with the Younger Dryas (12.5 - 11.7 ka) period recognized elsewhere in the Northern Hemisphere.

Here we present 46 new $^{230}$Th ages from five OCNM stalagmites to further constrain the growth history of speleothems from OCNM. These ages support previous work (Turgeon and Lundberg, 2001; 2004) in indicating that times of speleothem growth largely correspond to warm (interglacial) conditions. We also find, however, that several growth episodes occurred during cold (glacial) intervals and we investigate possible factors affecting speleothem growth during glacial conditions. Previous studies of speleothem growth have relied on regional or global climate records to infer possible environmental controls on speleothem formation (Baker et al., 1995; Genty et al., 2003; Gordon et al., 1989; Harmon et al., 1977; Spötl et al., 2002; Thompson et al., 1974), but this approach can be complicated by uncertainties in the chronologies of the paleoproxies used (e.g. Baker et al., 1993), particularly for periods beyond the limit of radiocarbon dating. In this study, we use a new approach to evaluate controls on speleothem growth by combining paleoproxy data with a regional
climate model to simulate the climate of SW Oregon at the peak of the last global glaciation, 21,000 years ago (21 ka), and with a thermal advection-diffusion model to calculate the sub-surface temperature evolution at the OCNM site through the last glacial cycle.

2.3 Regional Setting

OCNM is situated in the Klamath Mountains of SW Oregon (42°05′N, 123°25′W), approximately 75 km east of the Pacific coast (Fig. 2.1). The Klamath Mountains were formed by tectonic accretion along an active continental margin and are divided into four arcuate lithic belts bounded by east-dipping thrust faults (Snok and Barnes, 2006). The OCNM system is developed in a small Triassic marble lens and has several subhorizontal levels developed along fractures and faults at an elevation of 1150 to 1270 meters. The average mean annual air surface temperature (MAT) at OCNM for the period 1981-2006 was 8.8 °C ± 0.7°C (OCNM internal document). The active cave level is ~60 m below the surface and the cave temperature is approximately constant over the whole year at ~7°C. The modern vegetation in the area is dominated by *Pseudotsuga menziesii* (Douglas-fir) and *Abies concolor* (white fir).

The seasonal and annual climate of the Pacific Northwest (PNW) is strongly influenced by atmosphere-ocean interactions that originate in the Pacific Ocean (Taylor and Hannan, 1999). Regional patterns and seasonal variations in temperature track variations in ocean temperature, resulting in coherent changes of PNW climate west of the Cascade Range. Regional climate responds to contrasts in ocean-continent heating that lead to strong seasonality, with relatively warm and dry summers associated with high pressure over the North Pacific, and cool and wet winters associated with an intensification of the Aleutian low pressure cell (Fig. 2.1). Along the coast, seasonal temperature extremes are moderated by strong summer ocean upwelling, driven by northerly winds.

At interannual-to-decadal timescales, the climate of the PNW demonstrates significant covariance with large-scale dynamic oscillations such as the tropical El
Niño-Southern Oscillation, and the high-latitude Pacific Decadal Oscillation (Mantua et al., 1997). Over longer timescales, the geologic record clearly demonstrates significant variability in Pacific sea surface temperatures (SST) on millennial and orbital timescales (Mix et al., 1999; Ortiz et al., 1997). Variability at these timescales perhaps corresponds to sustained spatial patterns in the ocean-atmosphere system similar to those that occur at interannual and decadal timescales. However, additional large-scale controls also influenced the climate of the Pacific Northwest, including changes in atmospheric greenhouse gases, the seasonal and latitudinal distribution of solar insolation, and the size of North American ice sheets (Bartlein et al., 1998; Hostetler and Bartlein, 1999).

The cave hydrologic system responds quickly to atmospheric precipitation. All cave passages have clear seasonal cycles in drip rates, with dry passages during summer and high drip rates during winter and spring. Data from an automatic driplogger installed in the cave show that for at least one cave site, there is a one-month lag between time of first significant winter-season surface precipitation and reactivation of drips. As a consequence of the strong seasonality of precipitation, calcite deposition at OCNM is significantly reduced or terminated during summer when cave drip rates are minimal or nonexistent.

2.4 Controls on speleothem growth

Speleothems are secondary cave carbonates formed by the interaction of water and CO₂ with soluble carbonate bedrock (Dreybrodt 1980, 1999; White, 1988). Speleothem growth is largely determined by the climate régime above the cave because the chemical kinetics of the carbonate system are a function of water availability, partial pressure of soil CO₂ (pCO₂), and temperature (Kaufmann, 2003). In general, at mid- and high-latitudes, temperature is the principal controlling factor of speleothem growth on long timescales because it determines the physical state of water (liquid vs. ice) and controls biologic productivity (thus influencing soil pCO₂ and carbonate dissolution rate). Plant respiration and decay of organic matter in organic-rich soils supply CO₂ to percolating waters, leading to elevated pCO₂ in the
infiltrating solutions. When this slightly acidic water comes in contact with the carbonate bedrock, dissolution occurs until the water approaches saturation with respect to CO$_2$, and its dissociation products HCO$_3^-$ and CO$_3^{2-}$ (White, 1988).

Because cave pCO$_2$ is usually lower than that of the infiltrating water, drip waters are generally out of equilibrium with respect to cave pCO$_2$, so the CO$_2$ is released to the cave atmosphere and the water becomes supersaturated with respect to CO$_3^{-}$. Carbonate minerals (typically calcite) precipitate to form stalagmites, stalactites and flowstones, commonly known as speleothems (Moore, 1952). Thus, speleothem growth depends on a steady supply of water that is enriched in pCO$_2$, HCO$_3^-$ and CO$_3^{2-}$. In certain cases, calcite supra-saturation can also occur due to evaporation or due to common ion effects (Atkinson, 1983; Harmon et al., 1983).

2.5 Materials and methods

2.5.1 Chronology

Changes in the growth rate of any individual stalagmite may reflect some combination of variability in localized factors along the water path as well as climate (Baker et al., 1996). Accordingly, we collected five stalagmites from different locations in OCNM, as well as include the three samples from the Turgeon and Lundberg (2001; 2004) study, to account for possible localized factors in modulating the climate signal in growth. We sampled our five stalagmites from OCNM (Figure A1) by sectioning them parallel to the growth axis. Subsamples of 30 mg to 200 mg for $^{230}$Th dating were drilled perpendicular to the growth direction. In addition to sampling within intervals of similar calcite petrography to determine growth variability, we also targeted areas of visible changes in the color and texture of the stalagmites to determine whether these petrographic changes were associated with changes in growth.

Stalagmite OCNM02-1 is 26.5 cm long, up to 8 cm in diameter and XRD analyses of powdered samples indicate a composition of 99.9% calcite (Vacco et al., 2005). Between 15 and 16.5 cm from the top, the stalagmite has a distinctly darker interval, but no dissolutional features were identified either below or above this area
(Figure A1). Stalagmite OCNM02-2 is 29 cm long, up to 14.5 cm in diameter and is formed of several intervals of large prismatic calcite crystals bounded by intervals of darker calcite that include visible discontinuities. Stalagmite OCNM05 is 32 cm long, up to 12.5 cm wide, and comprises two coalesced stalagmites (OCNM05-1A and OCNM05-1B) that became connected after the dripwater location moved slightly sometime after 119 ka and before 57 ka. Sample OCNM07-1 is 5 cm long and 4 cm in diameter, yellow-brown in color and is formed of columnar calcite crystals.

Thirty-eight $^{230}\text{Th}$ ages were measured at the Minnesota Isotope Laboratory, University of Minnesota on a Thermo-Finnigan Element equipped with a double-focusing sector-field magnet in reversed Nier-Johnson geometry and a single MasCom multiplier in peak-jumping mode. Eight additional small-size (100 mg or less) samples (OCMN02-1-15.26 to 16.50 in Table A1) were analyzed on a multi-collector inductively coupled plasma mass spectrometer (MC-ICP-MS, Thermo-Finnigan-Neptune) using a MasCom multiplier behind the retarding potential quadrupole (RPQ) in peak-jumping mode; the Neptune has higher sensitivities for both uranium and thorium (close to one order of magnitude higher than the Element), with 1 to 2% combined ionization and transmission efficiency for both uranium and thorium. The chemical procedures used to separate uranium and thorium for $^{230}\text{Th}$ dating are similar to those described in Edwards et al. (1987). The procedures of characterizing multiplier and instrumental approaches were similar to those described in Cheng et al. (2000) and Shen et al. (2002). The Neptune measurements are similar to those described by Shen et al. (2002) for the Element, with the following two differences: (a) the higher ionization/transmission efficiency, and (b) we do not peak jump through background positions because the retarding potential quadrupole removes any significant tailing.

2.5.2 Climate model simulations

We used our regional climate model to simulate the climate of the region during the Last Glacial Maximum (LGM, 21 ka) and the present. The regional climate model comprises the RegCM2 atmosphere (Giorgi et al., 1993) and the full physics land-surface model (Thompson and Pollard, 1995) with six soil layers. The model
domain covers most of North America at a 50-km horizontal grid space and includes the North American ice sheets. Boundary conditions for the regional model simulations were obtained from global simulations of the 21 and 0 ka climates that were run with the GENESIS v2.3 general circulation model with appropriate large-scale boundary conditions which include continental ice sheets, atmospheric composition (180 ppmV CO$_2$ for LGM and 365 ppmV CO$_2$ for control), and specified monthly sea surface temperatures (SSTs) (LGM from Hostetler et al. (2006) and present from Levitus (1982)). The boundary conditions for the RegCM (vertical profiles of temperature, humidity and winds and SSTs) are updated every 6 hours of simulation. Further details about the models and the modeling methodology are given elsewhere (Hostetler and Bartlein, 1999; Hostetler et al., 2000; Hostetler et al., 2006).

2.5.3 Cave Temperature Modeling

Turgeon and Lundberg (2001; 2004) suggested the possibility that the freezing of dripwater sources, or of the water itself, associated with permafrost conditions above the cave caused growth hiatuses in the speleothems. As an additional method to evaluate the role of temperature over a longer time scale we applied a one-dimensional thermal advection-diffusion model to calculate the sub-surface temperature evolution at the OCNM site through the last glacial cycle. Inputs to the model are the geothermal heat flux from below, and air temperature and perturbations in solar radiation at the surface.

Movement of heat in the subsurface is governed by the supply of energy from the earth’s interior and the supply of energy from the surface, which depends on the surface energy budget as the advection of heat due to water percolating into the cave. The advection-diffusion equation representing heat transport can be written as:

$$\rho_m c_m \frac{\partial T}{\partial t} + \rho_w c_w w \frac{\partial T}{\partial z} = k \frac{\partial^2 T}{\partial z^2} + \Phi$$

where $T$ is temperature, $t$ is time, $\rho_m$ is the density of marble (2700 kg m$^{-3}$), $c_m$ is the specific heat capacity of marble (2639 J K$^{-1}$ kg$^{-1}$), $z$ is the vertical spatial coordinate, $w$ is advection velocity of percolating water assuming a 30 day surface-to-cave transit time ($1.897 \times 10^{-5}$ m s$^{-1}$), $\rho_w$ is density...
of water (1000 kg m\(^{-3}\)), \(c_w\) is the specific heat capacity of water (4181 J K\(^{-1}\) kg\(^{-1}\)), \(k\) is thermal conductivity (3.0 W m\(^{-1}\) K\(^{-1}\)), and \(\Phi\) represents heat sources (orbital insolation anomalies and geothermal heat fluxes). We use finite differences to implement Eq. 1 over a 1000 m domain with a 1-m vertical grid spacing and a 10-year time step. A small geothermal heat source at the bottom is specified as 90 mW m\(^{-2}\). We use the record of variations in SSTs of the California Current from ODP 1020 in the northeast Pacific (Lat: 41° 0.051’N, Long: 126° 26.024’W) (Herbert et al., 2001) as a surrogate for temperature variability at millennial scales and longer. This site lies 200 km to the west of OCNM, and present-day SSTs at the site are within ~2\(^\circ\)C of the MAT at OCNM, reflecting the strong influence of Pacific SSTs on the climate at OCNM. We adjust the SST record uniformly by this 2\(^\circ\)C such that the mean temperature for the last 1000 years equals the current cave temperature. For the upper boundary condition, the topmost cell is set to the time-varying SST record.

We also test the possibility that changes in direct solar inputs due to changes in orbital parameters could have led to warmer or colder cave temperatures. To do this, the time-varying anomaly in June solar energy flux at 40\(^\circ\)N (Berger and Loutre, 1991) is applied as an additional heat source to the topmost grid cell. The radiation anomaly in reality is modulated by cloud-cover, surface albedo, and possible shading by vegetation. We cannot quantify these factors, so we allow for a varying strength of the solar radiation signal by sensitivity tests in which we vary the range of the base radiation anomaly from 10 % to 70 % of the total anomaly and use the radiation range in combination with the present-day cave temperature to select those modelled time histories of cave temperature that simulated modern temperatures of 7 to 8.8\(^\circ\)C.

2.6 Results

2.6.1 Speleothem growth history

Based on 17 \(^{230}\)Th ages on three speleothems, Turgeon and Lundberg (2001, 2004) concluded that speleothem growth in OCNM was restricted to interglaciations. We constrain growth history for each of our five stalagmites from the age-depth model.
provided by our 46 new $^{230}$Th dates as well as the 10 dates previously published in Vacco et al. (2004) (Table A1). These data constrain two general features of stalagmite growth: intervals of no-to-slow growth and intervals when most growth occurred. Combining these data with previously published ages from Turgeon and Lundberg (2001, 2004) clearly demonstrates that most growth occurred during interglaciations (Fig. 2.2).

The 17 new $^{230}$Th dates from stalagmite OCNM02-1 are all in stratigraphic order, ranging from 0.46 ka to 64 ka (Fig. 2.2, Table A1). When combined with the 13 previously published $^{230}$Th dates for this stalagmite (Vacco et al., 2005) (Table A1), these ages suggest that the majority of the speleothem (94%) grew during MIS 5e and the Holocene (Fig. 2.2). The remaining 6% of the formation is a darker-colored interval, approximately 1.5 cm thick, which occurs between these two interglacial intervals (Figure A1). The youngest age from the MIS 5e interval (120.6 ka at 171 mm) occurs immediately adjacent to an age of 64.1 ka at 165 mm marking the start of the dark-colored interval, suggesting an interval of no-to-slow growth between MIS 5e and MIS 4 (Fig. 2.2). The contact is paraconformable, and the absence of dissolutional features at this contact suggests that water availability was likely the limiting factor in reducing growth. Ten ages from the dark-colored interval range from 17.7 ka and 64.1 ka, with half of them concentrated around 60 ka (Table A1). These ages indicate at least some deposition occurring during the Last Glacial Interval (LGI – MIS 2, 3 and 4), but growth was either episodic in response to changing environmental conditions above the cave, or was continuous, but very slow, and the $^{230}$Th dates average large intervals of time.

The eight $^{230}$Th dates from stalagmite OCNM02-2 yielded ages between 10 ka and 241 ka, with 87% of the formation growing during interglaciations (MIS 1, 5a, 5e, and 7) and 13% growing during glaciations (MIS 2, 3, 4, and 6) (Fig. 2.2). The speleothem material that grew during the last interglaciation is characterized by a large diameter and coarsely crystalline structure, suggesting abundant water supply and a thick water film on the speleothem surface. Of particular interest is the fact that growth began at 135 ± 1.2 ka, or when atmospheric CO$_2$ and summer insolation at the cave site were still near glacial levels (Fig. 2.2). The early warming suggested from
this interval of growth, however, is similar to records elsewhere that suggest an early start to the last interglaciation (Stirling et al., 1998; Henderson and Slowey, 2000; Spötl et al., 2002; Gallup et al., 2002). There is a dissolitional discontinuity between late MIS 6 and MIS 7, indicating that water undersaturated with respect to calcite was entering the cave during some of the glacial interval of MIS 6. Growth occurred during MIS 7, but an apparent age inversion (Table A1), likely due to contamination by detrital clays as suggested by the high $^{232}$Th concentrations of sample OCNM02-2-18 (Table A1), prevents an assessment of growth rates.

The seven $^{230}$Th dates for OCNM05-1A range from 119 ka to 123 ka, with all ages in stratigraphic order except for sample OCNM05-1-20.6. Accordingly, all of this formation grew during MIS 5e. The ten $^{230}$Th dates from stalagmite OCNM05-1B indicate that 81% of growth occurred during the present interglaciation and the remaining 19% occurring during the last glaciation (Figure A1, Fig. 2.2). The four $^{230}$Th dates for OCNM07-1 range from 2.3 ka to 6.1 ka, with all ages in stratigraphic order, suggesting that all of this formation grew during the present interglaciation (Table A1, Fig. 2.2).

In summary, the compilation of all of our new data indicates that 92% of speleothem growth occurred during interglaciations and 8% during glaciations. Although growth of any given individual speleothem was not necessarily continuous through the entirety of any given interglaciation, suggesting some internal hydrological variability in controlling growth, a primary climatic control is clearly indicated when including the results of all five samples (Fig. 2.2). Combining our new ages with previously published $^{230}$Th dates from OCNM (n = 73) (Turgeon and Lundberg (2004) and Vacco et al. (2005)) further supports this strong relation between speleothem growth and climate for the past 380 ka, indicating that interglacial environmental conditions (warm temperatures, high soil pCO$_2$) were favorable to speleothem formation (Fig. 2.2). However, speleothem growth also occurred intermittently during cold (glacial) periods (MIS 3,4 and 6), indicating that at these times temperatures were above 0°C, soil development was sufficient to allow carbonate dissolution, and water was dripping in the cave.
2.6.2 Temperature modeling

Figure 2.3 shows the results of temperature modelling for radiation values ranging between 10% and 70% and with the presence and absence of percolating water. Only the time series with values of 10% and 30% perturbations with dripwater present, and the 30% perturbation with no dripwater approximate the modern cave temperature of ~7°C. More importantly, the cave temperature does not drop below the freezing point of water during the entire 100-kyr modeled interval, except for short periods in the extreme 70% perturbation of radiation intensity. Accordingly, these results suggest that cave temperature remained above the freezing point throughout the LGI.

2.6.3 Climate Modeling

With the exception of a few months during the spring, our RegCM simulations produce monthly LGM surface air temperatures at OCNM that were about 4°C colder than present (Fig. 2.4). The colder temperatures are the result of colder LGM SSTs, lower trace-gas concentrations, and regional atmospheric circulations that are forced by the presence of the North American ice sheets. The influence of the ice sheets on atmospheric flow also induces substantial changes in the water balance at OCNM. Topographic and thermal effects from the ice sheet, accompanied by anticyclonic flow immediately south of the Cordilleran ice sheet, produces drier conditions over the Pacific Northwest in all but a few months when the LGM is as wet or slightly wetter than present (Fig. 2.4).

The simulated 2-meter MAT for present at OCNM is ~10°C, which is about a degree warmer than the observed value of ~9°C. At 4.5m soil depth, the mean-annual ground temperature (MAGT) in the control simulation is 8°C; the 2°C difference is similar to the observed difference. The simulated MAT for the LGM is 7°C. Soil temperatures are colder throughout the soil zone than those of the present in response to this surface cooling and changes in the surface energy balance (Fig. 2.5 A,C). Deeper soil temperatures, however, remain well above freezing. At the lowest level of the soil zone (4.5 m), the LGM MAGT is ~6°C, which is about 2°C colder than that of the present and 1°C colder than the LGM MAT.
Liquid water is reduced substantially at the LGM relative to the control throughout the soil zone (Fig. 2.5 D), whereas from November through April there is a substantial increase in frozen soil water during the LGM (Fig. 2.5 E). The reduction in liquid water in part reflects the overall drier surface conditions during the full glacial, but it also suggests an additional mechanism for limiting deep-soil moisture levels. The simulated annual average precipitation at OCNM is 2.2 mm d\(^{-1}\) and 3.6 mm d\(^{-1}\) for the LGM and present, respectively (Fig. 2.5 F,G). In contrast to the 39\% reduction in precipitation, LGM runoff is 74\% greater than that of the present. This counterintuitive result reflects the interplay of snow melt dynamics and the presence of frozen soils at the LGM. During the LGM, the uppermost soil-zone layers begin to freeze in November and continue to cool into the early winter. Subsequent winter snow, which is roughly double that of the present, accumulates and insulates the frozen ground well into the early spring. The snow also plays a role in suppressing the rise of spring air temperatures in that the bright snow surface reflects solar radiation that otherwise would be available for soil heating (TS2 in Fig. 2.5 F, G). As the snowpack melts, the relatively impermeable soil layer limits infiltration and most of the melt water runs off (Fig. 2.5 F). This dynamic is illustrated by the time series of the ratio of runoff to precipitation and snow water equivalent which peaks at \(~1.0\). In contrast, the control simulation exhibits no seasonal freezing of the soil zone, less winter snowpack, much smaller runoff ratios (maximum of \(~0.2\)), and thus more infiltration into the soil.

Together with general LGM drying, the seasonal cycle of the surface water balance thus provides a feedback mechanism to limit recharge to the deeper soil and reduce or eliminate drip water in the cave. Warm-season precipitation associated with the Mediterranean climate of the region (wet winter and dry summer) is insufficient to overcome the soil moisture deficit during the winter and spring.

The computed LGM temperature of the cave based on the thermal advection-diffusion model is 2.5\(^{\circ}\)C, a 4.5\(^{\circ}\)C cooling relative to present for a 4\(^{\circ}\)C depression in air temperature. Under our assumptions, this represents the plausible extreme in LGM cooling. The absence of freezing temperatures at the LGM is robust to a range of specified changes in radiation intensity. The temperature computations and climate
model simulations thus support the assertion of ice-free conditions in the cave, and the analysis of the simulated thermal and hydrologic dynamics in the climate model output provide a mechanism and support for limiting the supply of drip water and thus of speleothem growth at the LGM. Similar but less extreme dynamics likely operated during times of intermediate ice volume, atmospheric CO$_2$ concentrations, and SSTs, thus explaining the occurrence of some growth at other times during the LGI (Fig. 2.2).

2.7 Discussion

2.7.1 Temperature influence on speleothem growth

Turgeon and Lundberg (2001; 2004) argued that permafrost formation above OCNM during glacial intervals prevented water percolation in the cave and thus stopped speleothem formation. For permafrost to form, the mean annual temperature of air (MAT) should be at most 0°C, but usually a MAT limit of -3 to -5°C is required for discontinuous permafrost formation and even lower temperatures are necessary for continuous permafrost (French, 1976; Washburn, 1980). Given that the average MAT at OCNM is 8.8 °C ± 0.7°C, a minimum temperature depression of 9°C and more likely of at least 10°C would be necessary to sustain permafrost. Neither the results of our cave-temperature modeling or our climate modeling, however, support this magnitude of temperature depression required to form permafrost.

Proxy data from OCNM and the surrounding region support the modeled temperatures. Pollen data suggest that the MAT during MIS 2 was lower by 3-5 °C in southern Oregon and northern California (Briles et al., 2005; Heusser, 1998; Thompson et al., 1993). Furthermore, a survey of alpine periglacial features indicate that permafrost during the LGI only occurred in limited high-altitude areas of the Cascade Range in Oregon and northern California (Péwé, 1983).

2.7.2 Precipitation and vegetation influence on speleothem growth

Climate model simulations suggest that precipitation rates decreased significantly during the LGM relative to modern (Fig. 2.5), indicating that water balance was likely an important control on speleothem growth. In particular,
maximum reduction in LGM precipitation occurs during the winter months, which is the wet season today when dripwaters enter the cave. Because the LGM changes are largely induced by topographic and thermal effects from the ice sheet, modulation of precipitation by ice-sheet size may be an important factor in controlling speleothem growth rates.

In addition to the direct effect of water availability on speleothem growth, moisture supply, combined with temperature, also induces changes in biologic productivity, thus influencing soil pCO$_2$ and carbonate dissolution rate. For example, a pollen sequence from Carp Lake, WA (45°55’ N, 120°53’ W) indicates that for the past 125 ka, vegetation changes at this site were closely related to changes in ice volume, solar insolation, and atmospheric CO$_2$ (Whitlock and Bartlein, 1997), while a pollen record from Bolan Lake, ~10 km SW of OCNM, indicates that the surrounding area was covered by parkland vegetation during late MIS 2, similar to the modern subalpine environment (Briles et al., 2005). Such a change in vegetation would reduce biomass, and thus soil respiration rates, over the cave site, inducing a decrease in soil pCO$_2$. However, because there are no vegetation records near the cave that cover the entire glacial cycle and because our regional model does not explicitly resolve changes in soil pCO$_2$, we cannot quantify the effects of vegetation changes on speleothem growth.

We thus expect that changes in ice volume, solar radiation, and atmospheric CO$_2$ influence speleothem growth at OCNM through their influence on water balance and vegetation type. This relation is supported by our data, which demonstrate that greatest growth occurred during interglaciations (minima in ice volume, maxima in solar radiation and atmospheric CO$_2$), and little or no growth occurred during glaciations (Fig. 2.2).

2.8 Conclusions

Speleothem growth at OCNM provides a first-order indication of past climate conditions because it requires the presence of liquid water in the karst system, and vegetation cover above the cave. The combination of 27 previously published $^{230}$Th ages (Turgeon and Lundberg, 2004; Vacco et al., 2005) and 46 new ages from five
stalagmites indicates that maximum speleothem growth at OCNM occurred during interglaciations, and little growth occurred during glacial intervals.

To evaluate whether speleothem growth was interrupted by a temperature decrease sufficient to establish permafrost, by changes in water availability, or by some combination of these factors, we used a regional climate model to evaluate the magnitude and direction of changes in environmental variables (temperature, precipitation, soil moisture), and an energy balance model to assess the probability of freezing temperatures in the cave. The energy balance model results suggest that for the past 100 ka, cave temperature remained above 0°C. Similarly, the regional climate model simulation of LGM climate indicates that while surface soils froze seasonally, deeper soils remained unfrozen throughout the year. Accordingly, these results do not support the presence of permafrost as a mechanism to reduce speleothem growth during glaciations. On the other hand, the regional climate model suggests that a combination of drier LGM conditions combined with a change in the seasonal cycle of the surface water balance provides a feedback mechanism that limited recharge to the deeper soil and reduced or eliminated drip water in the cave. Warm-season precipitation associated with the Mediterranean climate of the region (wet winters and dry summers) is insufficient to overcome the soil moisture deficit during the winter and spring. These modeling results, which are supported by proxy data, thus indicate that OCNM climate and speleothem growth are sensitive to changes in precipitation and runoff associated with changes in ice-sheet size and radiative forcing.

2.9 Acknowledgments

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2.10 References


Figure 2.1. Sea-level pressure climatology (1948-2007) for western North America and eastern Pacific Ocean. A. Winter season (December to January). B. Summer season (June to August). The location of Oregon Caves National Monument is marked by an X in both graphs. Data retrieved and plotted from the NCEP Reanalysis website available at http://www.cdc.noaa.gov/.
Figure 2.2. Age versus depth plot for the stalagmites discussed in text (left plot). Most error bars in the present study are smaller than the symbol size (Table A1). Dashed lines are shown to facilitate comparison between dated stalagmites, and are not intended to indicate continuous growth between individual ages. Samples with high $^{232}$Th concentrations that resulted in age inversions (Table A1) are not plotted. Also shown (from left to right) are the $^{230}$Th ages obtained by Turegon and Lunderg (2004) (TL (2004)), the global benthic $^{18}$O stack (marine isotopes stages labeled) (Lisiecki and Raymo, 2005), the atmospheric CO$_2$ record from the Vostok ice core (Petit et al., 1999), and summer insolation at 45°N (Laskar et al., 2004). Shaded areas correspond with MIS 1, 5e 7c and 9.
Figure 2.3. Results of cave temperature modeling for the last 100 ka. The stippled line represents the adjusted SST record (see text for explanation) and the continuous line shows the average temperature of the model runs that produced modern cave temperatures of 7 to 8.8°C. The shaded gray area represents one standard deviation of the average modeled temperatures.
Figure 2.4. Monthly anomalies of 2-meter air temperature and precipitation for the LGM and present as simulated by the RegCM regional climate model. The anomalies are for 5-yr averages of each simulation. The location of the cave site is shown by an X on December maps. The location of the LGM North American ice sheets is outlined on each map.
Figure 2.5. Simulated seasonal cycle of soil temperature and moisture at OCNM obtained from the 6-layer soil representation in the LSX surface physics model. The values plotted were obtained by inverse-distance weighting of the four model grid cells nearest to the OCNM location. The total depth of the soil zone is 4.5 m; the depth scale on the ordinate axes is scaled relative to the mid-layer depths (0.025, .10, .25, .55, 1.25, and 3.0 m). Seasonal cycle of soil temperature (°K) for the LGM (A), present (B) and the 21ka-0ka anomaly (C). Anomalies of fractional volume of liquid soil water (D) and frozen soil water (E, note different soil depth axis). Monthly time series of 2-meter temperature (TS2), precipitation (Precip), snow water equivalent (SWE), surface runoff (Runoff), and a runoff coefficient (R/(P+SWE)) for the LGM (F) and present (G). Temperature is plotted on the left ordinates, SWE, runoff, and the runoff coefficient are plotted on the inside right ordinate and precipitation is plotted on the outside right ordinate. The error bars are one standard deviation (1σ) of the time series over the length of the model runs (8 years).
Chapter 3

Temporal variations of oxygen isotopes in rainwater from southwestern Oregon

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3.1 Abstract

We examine the relationship between climate variables and rainwater δ¹⁸O in event-based precipitation samples collected in SW Oregon. We derive back-trajectories for storms reaching our collection site for a period of 48 hours using the HYSPLIT model and we find that storm and local temperatures explain ~30% of δ¹⁸O variability. The relationship between temperature and δ¹⁸O is 0.7 ‰/°C and we conclude that this high slope is due to the dominance of winter rainfall at our site. Changes in specific humidity along the storm track suggest that continuous condensation and entrainment of moisture along the storm path may influence the δ¹⁸O composition of rainfall events in SW Oregon. The storms span over ~30° of latitude and a large gradient in sea surface temperatures and δ¹⁸O, but we find no significant relationship between trajectory origin or path and the isotopic composition of rainfall events sampled.

3.2 Introduction

Oxygen isotopes preserved in ice cores, speleothems, lake sediments and other geological archives have been used in the past three decades as indicators of past climate changes. They proved particularly useful at high latitudes where the ratio of oxygen-18 to oxygen-16 (hereafter δ¹⁸O) is positively correlated with temperature, with a slope of ~0.9 ‰/°C and ~0.69 ‰/°C for high latitudes in the Antarctic and Greenland respectively (Dansgaard, 1964; Peel et al., 1988).

The observed isotopic composition of rainfall at low-latitudes is inversely correlated with rainfall rate, a process that was called the “amount effect” by Dansgaard (1964). The amount effect was also identified at mid-latitudes where it is found especially during intense convective storms in the summer time (Dansgaard, 1964), although winter precipitation is more likely to correlate with temperature. Studies of mid-latitude event-based precipitation noticed only a weak correlation of δ¹⁸O with temperature and a strong correlation with precipitation amount (Gedzelman et al., 1987; Lawrence et al., 1982; Treble et al., 2005). However, long-term averages...
of temperature and rainwater $\delta^{18}O$ suggest a close correspondence between these two parameters, and a slope of $\sim 0.5 \% \text{c/o} ^\circ \text{C}$ for many mid-latitude locations (Rozanski et al., 1993).

Paleo-temperature or paleo-rainfall reconstructions based on stable isotopes preserved in different environments can be complicated by changes in transport pathways of the water vapor, changes in source regions, or changes in the seasonality of precipitation (Áraguas-Áraguas et al., 2000). For example, modeling studies that include predictions of stable isotopes in water showed that a large portion of the variance in rainwater $\delta^{18}O$ cannot be explained by either temperature or precipitation (Noone and Simmonds, 2002) and factors such as the regional balance between precipitation and evaporation have a major influence on the isotopic composition of rainwater (Lee et al., 2007). Other models also suggest non-stationary local relationships between water isotopes and climate variables (Schmidt et al., 2007).

The degree to which these changes affect the isotopic composition of rainfall in the western U.S. is difficult to evaluate because of the low density and short record of the Global Network of Isotopes in Precipitation (GNIP) stations, although there are substantial recent efforts to increase the number of monitoring stations for the entire U.S. (Welker, 2000). In the present study we investigate the effects of rainfall amount, temperature and changes in storm trajectory on the isotopic compositions of rain events sampled at Oregon Caves National Monument (OCNM) in western United States ($42^\circ \text{N}, 123^\circ \text{W}$). Our study site is located at an altitude of 1200 m in the Klamath Mountains, approximately 75 km east of the Pacific coast (Fig. 3.1). The mean annual temperature at OCNM is 8$^\circ \text{C}$, the mean annual precipitation is 1600 mm and approximately 85% of precipitation falls between November and April (PRISM Group, Oregon State University, http://www.prismclimate.org). The Pacific Northwest region of the United States is strongly influenced by large-scale atmospheric circulation patterns in the Pacific Ocean that create a significant seasonal cycle in surface winds. This is driven by the migration of the Pacific subtropical high-pressure cell which is displaced southward during winter months by the Aleutian low-pressure system, resulting in primarily southwesterly winds and abundant precipitation. The opposite situation occurs during summer months when the Aleutian low is weakened.
and the northward migration of the subtropical high causes predominantly northeasterly winds and dry conditions (Taylor and Hannan, 1999).

3.3 Methodology

We collected 49 rainfall samples between 2005 and 2008 and each sample represents precipitation accumulated during an interval of 2-5 hours. The water was collected in 40 ml glass bottles with airtight screw-cap seals and plastic vapor barriers, which were subsequently stored in a refrigerator for periods of 1 to 12 months before measurement. Rainwater $\delta^{18}O$ was measured using established H$_2$O-CO$_2$ equilibration methods (Epstein and Mayeda, 1953) on a ThermoQuest Finigan Deltaplus XL mass spectrometer at Oregon State University equipped with an automated equilibrator of local design. The isotopic composition of waters is reported relative to VSMOW based on analyses of the standards NIST-8535 (VSMOW), WAIS3 (West Antarctic Ice Sheet from INSTAAR Stable Isotope Laboratory), HOT (500 m seawater filtered at 1µm) and LROSS (calibrated Lab Reverse Osmosis water). The reproducibility of local reference waters ran at the same time with the samples was ±0.03‰ (n=23) on this system. The trajectories of air masses that delivered precipitation at our study site were derived using the Hybrid Single-Particle Lagrangian Integrated Trajectory Model (HYSPLIT), version 4 (Draxler and Hess, 1997) from the National Oceanic and Atmospheric Administration’s (NOAA) Air Resources Laboratory. This method provides insight into the history of the air masses that were sampled in our study and has been used in other similar studies that related air trajectories with rainfall isotopic composition (Burnett et al., 2004; Lee et al., 2003). The trajectories were visually checked against satellite imagery obtained from NOAA’s Geostationary Operational Environmental Satellites (GOES).

All trajectories were started at 500 m above the ground and were traced back 48 hours before arrival. Temperature, precipitation, altitude, pressure and relative humidity were calculated hourly for each trajectory using information from the EDAS 40 km meteorological dataset.
3.4 Results

The values for rainwater $\delta^{18}O$ are negatively skewed, with a mean of $-10.8 \%_o \pm 4.1 \%_o$ (1σ), and range from $-22.6 \%_o$ to $-3.8 \%_o$ (Fig. 3.2, 3.3). Because of the precipitation seasonality in SW Oregon, our sampling is biased towards the cool-rainy season when most of precipitation falls at our site. Skewness is associated with a small population of highly depleted values with $\delta^{18}O$ of less than $-19 \%_o$, a surprising level of depletion in this mid-latitude site near the ocean. These highly depleted values were analyzed in replicate and confirmed. At least one such value was measured in each of four winters.

3.5 Effect of storm trajectories

Previous studies that assessed the influence of storm track trajectories on rainfall $\delta^{18}O$ showed a strong connection between the path of the storm and the isotopic composition of precipitation (Barras and Simmonds, 2008; Burnett et al., 2004; Gedzelman and Lawrence, 1982). The analysis of back-trajectories shows that storm tracks span an area from 26°N to 55°N and from ~130°W to 150°W. In Fig. 3.4 we show the differences in the path of storm track trajectories for rainfall events with low $\delta^{18}O$ compared with high $\delta^{18}O$ rainfall events. Back-trajectories for high $\delta^{18}O$ events ($\delta^{18}O$ values $\geq -8 \%_o$) are spanning over ~20 degrees of latitude, from 26°N to 46°N and the gradient for seawater $\delta^{18}O$ over this latitude range is ~1.2 \%_o. Depleted trajectories (defined here as having $\delta^{18}O$ values $\leq -18 \%_o$) span over ~10 degrees of latitude (32°N to 42°N) and an area of ~1 \% change in seawater $\delta^{18}O$, and tend to spend more time over land compared with enriched trajectories. Overall, the trajectory analysis shows that the source and direction from which the storm tracks approach the west coast of Oregon does not have a major impact on the final $\delta^{18}O$ composition of rain that falls at OCNM (Fig. 3.4). It would therefore appear that, at least on the synoptic timescales analyzed here, the large differences in $\delta^{18}O$ of rainwater for depleted and enriched events cannot be attributed to systematic changes in air mass trajectory. However, there are several uncertainties in our trajectory analysis including errors associated with the representation of the atmospheric flow field with gridded
data points, uncertainties related to model representation of air flow across the complex topographic features of the Klamath Mountains, and the tracking time of 48 hours before arrival may not be representative for the overall storm trajectory.

3.6 Effect of temperature

The linear relationship observed in mid- and high-latitudes between the isotopic content of precipitation and surface air temperature has been explained in terms of a gradual rainout of atmospheric water vapor due to changes in solubility of water vapor in air with lower temperatures (Dansgaard, 1964). However, recent studies that included water isotopologue tracers in general circulation models, highlighted the complexity of the controlling factors that can affect the isotopic composition of rainwater (Lee et al., 2007; Schmidt et al., 2007).

Our study area is a relatively simple setting where the water source remains constant (the Pacific Ocean) throughout the year and the precipitation falls overwhelmingly only during one half of the year. On the west coast of Oregon and northern California, seasonality is a minor component of annual rainfall $\delta^{18}O$ variability because the majority of precipitation falls during the cool season, and a sensitivity study of seasonally changing $\delta^{18}O$ in rainfall from North America (Vachon et al., 2007) indicates that western Oregon has one of the lowest seasonal $\delta^{18}O$ ranges in the United States (1-2.5‰). In many other areas where the precipitation is more evenly distributed throughout the year, or where there are several different sources of atmospheric moisture, the interpretation of rainfall $\delta^{18}O$ is often more complex due to seasonally changing temperatures, changes in evapotranspiration fluxes or shifts in the source areas of water vapor (Rozanski et al., 1993).

Fitting a linear model to describe the relationship between rainwater $\delta^{18}O$ and temperature at the time of sample collection (Fig. 3.5) indicates that there is a significant statistical relationship between these two parameters (P-value = 0.0001, $r=0.54$). This moderately strong correlation indicates that local temperature explains 29% of the variability in $\delta^{18}O$ with a slope of 0.7‰/°C. It can be seen from Fig. 3.4, however, that enriched events also correspond with warmer storm temperatures while
depleted events are characterized by lower temperatures along the storm track. The relationship between average storm temperature and rainwater $\delta^{18}O$ is strongest when we consider storm temperatures for the last five to twelve hours before arrival (Fig. 3.5). Over this period the storm temperature explains 31% of rainwater $\delta^{18}O$ variability with a slope of 0.7 ‰/°C. Performing a multiple linear regression indicates that if both storm and local temperatures are included, the regression model explains 37% of the variance in rainwater $\delta^{18}O$. Even though local and average storm temperature explain a similar amount of variance in the $\delta^{18}O$ record, their mutual correlation is not particularly strong ($r = 0.6$) suggesting either that their influence on the isotopic composition of rainwater is distinct or it could related to the averaging of temperatures over the entire grid size used in the model.

In general, for depleted trajectories temperatures are rising for the period between 48 and 8 hours before arrival, while for enriched trajectories temperatures tend to remain approximately constant or vary within a narrow range over the same time interval. Both enriched and depleted trajectories show a sharp decrease in temperature during the last 8-4 hours before arrival as the air masses are raised over the high altitudes of the Klamath Mountains (Fig. 3.6).

The slope of 0.7 that we observe for the regressions of temperature versus precipitation $\delta^{18}O$ is higher than the average of 0.5 commonly observed for the mid-latitudes (Dansgaard, 1964), especially considering that our site is in a near-coastal setting where slopes are generally lower (Rozanski et al., 1993). For example, rainfall $\delta^{18}O$ data from Vancouver Island (British Columbia), which is the nearest GNIP station that has both $\delta^{18}O$ and temperature information, was collected from 1975 to 1982 (IAEA/WMO, 2006). Here, the monthly precipitation $\delta^{18}O$ is well correlated with monthly temperature ($r = 0.8$) with a slope of 0.2 ‰/°C. However, if we consider only the cool half of the year (October-March) the slope (and correlation) increases to 0.5 ‰/°C ($r = 0.9$) while during the warm season (April-September) the slope is 0.1 ‰/°C ($r = 0.5$). It is likely that the lower slope during the warmer half of the year results from a combination of lower fractionation coefficients for the liquid-water transition vs. solid-vapor phase change, the mixing of both temperature and mixing ratios during convection (Jouzel, 1986) and higher evapotranspiration during the plant
growing season (Rozanski et al., 1993). These complicating effects are not important in southern Oregon, where the summers are dry.

### 3.7 Effect of rainfall amount

The amount effect has been explained by a high degree of removal of $^{18}$O from the cloud layer during intense precipitation events, an incomplete equilibration of large rain drops with water vapor near the ground, evaporation below the cloud base causing lower $\delta^{18}$O values of rainwater due to evaporative loss of lighter isotopes (Dansgaard, 1964), or continuous isotopic exchange of the water vapor with raindrops below the cloud base (Miyake et al., 1968; Rozanski et al., 1993).

The regression analysis of the relationship between rainfall accumulated 48 hours prior to sample collection and $\delta^{18}$O indicates no statistically significant association between these two parameters at 95% confidence ($r^2=-0.06$, $P=0.909$) regardless of the time interval for which precipitation amount is considered. Rainfall rates along the storm tracks are more heterogeneous for events with high $\delta^{18}$O values and can be divided into a group that has low precipitation amounts throughout the 48 hours before arrival and a group that has highly variable precipitation rates (Fig. 3.7). In the case of depleted trajectories, precipitation starts to increase only during the last ~8 hours before arrival as the airmasses rise and cool over the Klamath Mountains.

Large or rapid changes in specific humidity along the storm tracks can indicate condensation or entrainment of moisture (Barras and Simmonds, 2008). The calculated specific humidities along the trajectories indicate that depleted trajectories start picking up moisture 30-16 hours before arrival, while enriched trajectories have ~constant humidities indicating water vapor exchange of the airmass for the much of the 48 hours before arrival (Fig. 3.8). This prevents the continuous depletion of airmasses through Rayleigh distillation and contributes to the higher $\delta^{18}$O values measured in rainwater.

To obtain a semi-quantitative distinction between the factors that influenced the most depleted and most enriched rainwater samples we conducted a Q-mode factor analysis (Klovan and Imbrie, 1971; Miesch, 1980) using the parameters obtained for
storm trajectories in the HYSPLIT model. Included in this analysis are latitude, longitude, altitude, pressure, temperature, precipitation and specific humidity. The first three extracted varimax factors in the Q-mode analysis for depleted storms explain 98% of the variance in rainfall $\delta^{18}$O. Factor 1 explains 64% of the variance and is dominated by temperature (Fig. 3.9), factor 2 has the highest scores for altitude and explains 24% of the variance, while factor 3 explains 10% of the variance and is clearly dominated by precipitation. In the case of enriched storms the first three factors explain 98% of the rainfall $\delta^{18}$O variance. Factor 1 has 76% of the total variance and shows similar factor scores for all of the variables, except rainfall and altitude, factor 2, which explains 12% of the variance, is dominated by altitude, and factor 3 is clearly dominated by precipitation, explaining 10% of the variance. It would therefore appear that while storm temperature is the dominant control over $\delta^{18}$O values in depleted storms, the situation is more complex in warmer storms, although temperature still plays a major role.

3.8 Summary

We conducted an analysis of event-based rainfall $\delta^{18}$O variability in SW Oregon which has a climate characterized by a strong seasonality of precipitation and where the moisture source is entirely derived from the Pacific Ocean. We assessed the influence of changes in storm trajectories, temperature and rainfall amount on $\delta^{18}$O values and found that storm and local temperature play an important role in determining the isotopic composition of rainfall. We find that the local slope of temperature vs. rainfall $\delta^{18}$O is 0.7 ‰/°C and conclude that this high slope (compared with other mid-latitude locations) is due to the dominance of winter rainfall at our site. Temperature explains ~30% of the variance in rainfall $\delta^{18}$O, and therefore a large amount of variability cannot be explained by either temperature or precipitation. Changes in specific humidity along the storm track suggest that continuous condensation and entrainment of moisture along the storm path, particularly in the case of storms with high $\delta^{18}$O values, may influence the $\delta^{18}$O composition of rainfall events in SW Oregon. Storms reaching OCNM span over ~30° of latitude and a large
gradient in sea surface temperatures and \(\delta^{18}O\). However, we found no significant relationship between trajectory origin or path and the isotopic composition of rainfall events sampled at OCNM. Similarly, there is no significant association between rainfall amount and rainfall \(\delta^{18}O\) values. Based on the results presented above, paleoclimate studies using stable isotopes of oxygen (e.g. speleothems) in western Oregon and the Pacific Northwest are likely to capture past temperature changes in this area.

3.9 Acknowledgements

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3.10 References


Figure 3.1. Location of the OCNM (black arrow) in the Klamath Mountains.
Figure 3.2. Histogram showing the distribution of δ18O values for the rainwater samples collected at OCNM between 2005 and 2008.
Figure 3.3. Timeseries of precipitation (upper panel) and temperature (lower panel) plotted together with rainwater $\delta^{18}$O of samples collected at OCNM.
Figure 3.4. Back-trajectories for low $\delta^{18}$O (A) and high $\delta^{18}$O (B) rain events arriving at OCNM at an altitude of 500 m above ground level. The color scale of the storm tracks represents the hourly temperatures as calculated by the HYSPLIT model, and the colored background represents the gridded sea surface $\delta^{18}$O from LeGrande and Schmidt (2006).
Figure 3.5. Plot of average storm temperature for the last 12 hours before arrival versus precipitation $\delta^{18}\text{O}$ (A), and local temperature versus $\delta^{18}\text{O}$ (B).
Figure 3.6. Calculated temperatures for the last 48 hours before arrival at OCNM for depleted and enriched trajectories.
Figure 3.7. Rainfall amount versus hours before arrival at OCNM for depleted and enriched events. Colors are the same as in Fig. 3.5. It can be seen that depleted events have low precipitation rates between 48 and 8 hours before arrival after which there is an abrupt increase in rainfall amount.
Figure 3.8. Plot of specific humidity versus hours before arrival suggesting the continuous entrainment of water vapor for enriched trajectories and a gradual increase in specific humidity for depleted trajectories.
Figure 3.9. Score loading for the first three varimax factors from Q-mode factor analysis. A. Factor scores for depleted events; B. Factor scores for enriched events. The labels for the x-axes are latitude (La), longitude (Lo), altitude (Al), pressure (Ps), temperature (Te) precipitation (Pr) and specific humidity (Sh).
Chapter 4

A high-resolution Holocene and Last Interglacial Period climate reconstruction in mid-latitude western North America

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4.1 Abstract

We present a high-resolution, absolutely dated winter climate reconstruction in western North America using speleothems from a cave in Oregon that spans the entire Holocene and the last interglacial period. We use speleothem δ¹⁸O as an indicator of temperature change, while the δ¹³C is interpreted as reflecting mainly changes in precipitation and soil biomass. Holocene winter climate at our site was influenced by long-term changes in solar insolation, but also exhibits shorter, millennial and centennial climate variability. The phase relationship between speleothem δ¹⁸O and δ¹³C suggests a delayed response of vegetation to climate change at the millennial time band, and synchronized changes in temperature and precipitation on multi-centennial timescales. Time series analyses identify significant, but non-stationary spectral power at millennial and centennial timescales, some of which might correspond with solar cycles. We identify maximum climate variability corresponding with the peak rate of warming which by analogy implies that global warming may make future climate in the Pacific Northwest more unstable. Compared with the Holocene, the last interglacial period was characterized by similar average temperatures, but was possibly wetter and/or had higher soil productivity.

4.2 Introduction

Studies of past climate variability in the western U.S. and the eastern Pacific highlighted important changes in temperature, precipitation, vegetation and ocean circulation on decadal to orbital timescales. Even though climate variability during the Holocene had lower amplitudes than during the last glacial period these changes had a significant impact on ecological communities and perhaps human society. Constraining the timing of these changes is important for documenting how coupled atmosphere-ocean-terrestrial interactions have affected the climate of western North America and requires long, high-resolution and well-dated paleoclimate datasets. Water resources in western U.S. are largely dependent on winter climate when the majority of precipitation falls and when the impact of changes in the large-scale Pacific Ocean coupled ocean-atmosphere system are the strongest. Here we present
an absolutely-dated, high-resolution winter climate reconstruction that spans the entire Holocene and the middle of the last interglacial period using speleothems from Oregon Caves National Monument (OCNM) in southwestern Oregon. We first introduce the regional physical and climatological setting of our study area followed by a description of the methods and the results obtained. We then discuss the main findings of this study and the broader implications of our results first for the Holocene and then for the last interglacial period.

4.3 Regional Setting

OCNM is situated in the Klamath Mountains at an altitude of ~1200 meters and consists of several sub-horizontal levels with a total surveyed length of ~5.6 km. The cave is developed in a Triassic marble lens which is surrounded by metavolcanics and metasediments. Approximately 85% of precipitation falls between November and April (PRISM Group, Oregon State University, http://www.prismclimate.org). The cave temperature is constant through the year at 7°C while the driprates show seasonal variations with active drips during the cool season and dry passages during the warm season. As a result of these seasonal changes in drip rates, which are associated with changes in precipitation regime outside the cave, the speleothems at OCMN are likely to reflect cool season climate variations. Since OCMN is located only 75 km inland from the Pacific coast, climate variability at our study site is strongly coupled with the Pacific Ocean basin, which has been one of the strongest sources of regional-to-global climate variability during the Holocene.

The climate of the Pacific Northwest is dominated by seasonal changes in the strength of the Aleutian Low and of the Subtropical high pressure system, resulting in cool-wet winters and warm-dry summers. The major mode of interannual climate variability is associated with El Niño-Southern Oscillations (ENSO) which affects mainly the cool season by causing cooler and wetter condition during La Niña years and warm-dry anomalies associated with El Niño. Decadal-scale variability is influenced by the Pacific Decadal Oscillations where the positive (negative) PDO phase is associated with cooler (warmer) sea surface temperatures (SSTs) in the
central North Pacific, and warmer (cooler) SSTs in the Gulf of Alaska and along the 
Pacific coast of North America. Increased frequencies of El Niño years tend to 
correspond to times of a positive PDO, whereas increased frequencies of La Niña 
years correspond with the negative phase of the index. Climate anomalies in the PNW 
during a positive (negative) PDO are thus similar to those associated with warm (cool) 
phases of ENSO. Climate mechanisms affecting the Pacific Ocean and western North 
America on longer timescales are less well understood, but geologic records clearly 
demonstrate significant climate variability on millennial and orbital timescales.  

4.4 Methods

We used four stalagmites from OCNM (OCNM02-1, OCNM05-1A, OCNM05-1B 
and OCNM07-1) that were cut in half, parallel to the growth direction, and polished 
(Figure A1). Stalagmites OCNM05-1B and OCNM07-1 are used to obtain a low-
resolution replication of the high-resolution Holocene record from stalagmite 
OCNM02-1. Samples for U-Th dating were obtained by drilling ~200 mg of calcite 
powder parallel to the growth layers which were processed following procedures 
described in Edwards et al. and Dorale et al. by inductively coupled plasma mass 
spectrometry (ICP-MS) using a Thermo-Finnigan Element at the University of 
Minnesota. Stalagmites were micro-sampled using a New Wave MicroMill and stable 
isotopes were measured at Oregon State University on a Finnigan-MAT 252 mass 
spectrometer interfaced with a Kiel III online acid digestion device. Results are 
reported relative to the VPDB standard and external precision on replicate samples 
(NBS 19, NBS 20, and a local carbonate standard) run daily on this system was 0.06‰ 
and 0.02‰ for δ¹⁸O and δ¹³C respectively.

4.5 Age Model

Various strategies can be employed in constructing age-depth relationships for 
paleoclimatic timeseries. Linear models provide a simple and straightforward 
approach and commonly have narrow confidence intervals, but assume that growth 
rates change sharply at the depths corresponding with dated locations. On the other
hand, polynomial models imply that there are no abrupt changes in growth rates but
the interpolation model does not necessarily go through the dates. For stalagmite
OCNM02-1, a polynomial trend of any order violates the constraints imposed by the
2σ dating error of some U-Th ages and results in a reverse slope in the older part of
the record (Fig. 4.1). We construct instead three different linear age models for our
Holocene timeseries: age model 1 (AM1) includes the nominal U-Th dates listed in
Table A1; age model 2 (AM2) uses the ages corresponding with the younger end of
the errors for the U-Th dates; and age model 3 (AM3) includes the older extreme of
the dating errors (Fig. 4.1). The average analytical error (142 years) is on the same
order of magnitude as the error associated with the width of the channels where
samples for U-Th dates were collected (160 years).

4.6 Relationship between δ¹⁸O and climate

A survey of event-based precipitation samples collected at the cave site from 2005 to
2008 indicates a moderately strong relationship ($r = 0.6$) between temperature and
δ¹⁸O, with a slope of 0.7 ‰ /°C (Ersek et al., in prep.). This positive relationship
between temperature and rainwater δ¹⁸O is further supported by long-term (1975-
1982) monthly data from the GNIP station on Vancouver Island (Canada) where the
slope during the cool-wet half of the year (October through March) is 0.5 ‰ /°C ($r=
0.9$). Since these slopes are higher than the water-calcite isotope fractionation of
-0.2 ‰ /°C, and because of the seasonality of precipitation, we interpret the changes in
speleothem δ¹⁸O variability as reflecting predominantly winter temperatures. We
recognize, however, that other factors, such as changes in sea surface temperature, in
precipitation-evaporation balance over the Pacific Ocean, and in seawater δ¹⁸O or
processes internal to the karst system may also influence the δ¹⁸O composition of
OCNM speleothems through time.

4.7 Relationship between δ¹³C and climate

The δ¹³C composition of speleothems is derived from the dissolved inorganic
carbon of dripwaters, which in turn is related to processes that occur both internally
and externally to the cave system. Increased (decreased) soil productivity and
vegetation density leads to more negative (positive) δ¹³C values, but the relationship
between soil productivity and climate varies by region. Due to the seasonality of precipitation with respect to the plant growing season and water infiltration in the cave, at OCNM we expect soil productivity to be higher during wetter climate conditions compared with drier climates, and the effect of changes in soil biomass on speleothem $\delta^{13}C$ to be biased towards the spring-early summer season.

Changes in speleothem $\delta^{13}C$ can also depend on whether the groundwater flows under open- or closed-system conditions. Under open-system conditions the water is in contact with the soil CO$_2$ and has more depleted $\delta^{13}C$ values, while in a closed system the water does not maintain exchange with the soil CO$_2$, resulting in more positive $\delta^{13}C$ values because the carbon is consumed during dissolution. Open-system conditions are more likely to occur during wet periods while closed-system conditions are favored by drier conditions. Additional factors may also influence speleothem $\delta^{13}C$ towards lower values with an increase in precipitation, including less prior calcite precipitation and/or increases in drip rates. However, increases in precipitation rate may limit the extent of soil-water interaction resulting in more positive $\delta^{13}C$ values. Evaporation of groundwater in the epikarst or in the cave represents a non-equilibrium process that may also increase the $\delta^{13}C$ values of the speleothems and obstruct the paleoclimate interpretation, but it is more likely to occur under drier conditions.

Because of large differences in the isotopic composition of C$_3$ (-26‰) versus C$_4$ (-12‰) plants, changes in their distribution through time are likely to be captured by speleothems forming at the same time as vegetation changes. However, the present biomass above the cave is formed of C$_3$ vegetation, and pollen records from the Pacific Northwest showed that this was also true for at least the past 125 ky, so changes between C$_3$ and C$_4$ plants cannot explain the speleothem $\delta^{13}C$ variability at OCNM. We therefore expect the OCNM $\delta^{13}C$ values to be more negative during wetter-than-average conditions due to a combination of increased soil productivity, more open-system conditions and/or higher drip rates in the cave.
4.8 Results

An important first step in speleothem studies is to establish if calcite was formed in isotopic equilibrium with cave dripwaters. Hendy proposed two criteria to determine equilibrium fractionation of oxygen and carbon isotopes in speleothems: $\delta^{18}$O should be constant along a single growth layer and the covariance of $\delta^{18}$O and $\delta^{13}$C along the growth axis should be low. While the latter criterion can be readily applied to speleothems, sampling along a single growth layer is difficult because of the slow speleothem growth rates, irregular layer shapes and thinning of the layers from the growth axis of the stalagmite towards the edges. We find that the longitudinal correlation between $\delta^{18}$O and $\delta^{13}$C is low for both OCNM02-1 ($r^2 = 0.17$, n=2600) and OCNM05-1 ($r^2 = 0.008$, n=318). An initial sampling along a growth layer in OCNM02-1 and OCNM05-1 using a hand-held dental drill had nearly constant $\delta^{18}$O values (standard deviation = 0.06 for OCNM02-1 and 0.04 for OCNM05-1). For stalagmite OCNM02-1 we also sampled along a growth layer using a New Wave Micromill by first drilling a trench above the layer of interest and gradually milling in small increments until we reached the target layer. The samples collected away from the growth axis showed no systematic $\delta^{18}$O enrichment towards the edges of the stalagmite and had low $\delta^{18}$O variability (standard deviation = 0.04).

To further test the presence of equilibrium conditions we obtained a low-resolution record from two additional Holocene stalagmites (OCNM05-1B and OCNM07-1) in order to replicate the first-order characteristics of the Holocene record. This test is useful because it is improbable that speleothems formed under non-equilibrium conditions will exhibit similar variability through time.

Insofar as stalagmites OCNM02-1 OCNM05-1B and OCNM07-1 were collected in different locations in the cave, it is likely that they experienced different growth conditions (e.g. different flow pathways of dripwater, variable drip rates, different carbonate concentration, etc). The $\delta^{18}$O and $\delta^{13}$C values from these three stalagmites are shown in Fig. 4.2. The $\delta^{18}$O in all stalagmites have an increasing trend from lower values in the early Holocene to higher values during the late Holocene. Furthermore, both OCNM02-1 and OCNM05-1B share a pronounced isotopic depletion during the Younger Dryas cold interval. The $\delta^{13}$C values for the Holocene are very similar for
OCNM02-1 and OCNM07-1, but, while sharing similar long-term trends with OCNM02-1 and OCNM07-1, stalagmite OCNM05-1B is characterized by values that are ~1‰ more depleted. However, between 11 and 12.5 ky OCNM05-1 has more positive values than OCNM02-1. These differences are likely related to process affecting the chemical evolution of water in the soil and epikarst region such as the degree of open/closed chemical systems along the water path or degassing of CO₂ within the epikarst zone.

While more detailed comparisons between the three records are not feasible at this time due to the poorer age control and potential aliasing of the signal as a result of lower resolutions in OCNM07-1 and OCNM05-1B, the general agreement between the stalagmites together with the Hendy criteria discussed above, further supports the presence of equilibrium conditions during speleothem formation at our site and preservation of the paleoclimate signal in OCNM speleothems.

4.9 The Holocene

The age control for stalagmite OCNM02-1 is given by 10 U-Th dates with an average error of 140 years (Fig. 4.1). The average Holocene growth rate was 13 mm/ky with the fastest growth occurring during the mid-Holocene (between 5.7 and 7.3 ky) when the growth rate was twice as fast (26 mm/ky) than the Holocene average. The mean spacing of the OCNM stable isotope record is 3 years for the period between 0.2 k.y and 6.49 ky, after which the resolution increases to an average of 19 years between 6.5 and 7.3 ky, and an average of 105 years for periods after 7.3 ky

Because the total amplitude of the Holocene δ¹⁸O record is relatively small (1.5 ‰) we evaluated the degree to which we can identify fine structure in the isotope record by comparing it’s spectral characteristics with several realizations of white noise time series spectra that have the same standard deviation as the analytical error for δ¹⁸O measurements. We conclude that at periods less than ~10 years the isotope signal cannot be reliably distinguished from white noise and therefore we resample the entire Holocene record at 10-year intervals using Gaussian interpolation to minimize the aliasing of the time series.
One feature that stands out in the OCNM isotope record is the long-term trend of increasing δ\(^{18}\)O values from an average of -9.3‰ ± 0.2‰ in the early Holocene to an average of -8.8‰ ± 0.1‰ for the last 1000 years of growth. For comparison, two soda-straw stalactites that were growing in the cave in 2007 have δ\(^{18}\)O values of -8.7‰ and -8.8‰ (Fig. 4.3). The long-term trend in stalagmite OCNM02-1 can be attributed to a rise in winter insolation through the Holocene and our δ\(^{18}\)O record is well correlated with winter insolation (r = 0.7). The early Holocene was characterized by higher values of Earth’s tilt and the occurrence of perihelion in summer, a combination that induced a 10% reduction of winter insolation and an 8% increase of summer insolation over North America\(^{31}\). The winter insolation values rose from 115 Wm\(^{-2}\) in the early Holocene to 132 Wm\(^{-2}\) at present\(^{32}\), which implies a δ\(^{18}\)O sensitivity to orbital solar forcing of ~0.03‰ / Wm\(^{-2}\) for this stalagmite. The control of solar insolation on climate and vegetation is also readily apparent in many paleoclimate records from northwest North America and elsewhere\(^{1,33,34}\).

Superimposed on the long-term trend is significant millennial to multi-decadal climate variability characterized by both gradual and abrupt shifts. There is a step-wise transition between ~6.7 and 5.7 ky from the lower δ\(^{18}\)O values of the early Holocene to the higher overall δ\(^{18}\)O values that characterize the mid-late Holocene. Several prolonged periods of low δ\(^{18}\)O values are apparent in the speleothem record for the past 9 ky with inferred century-scale cooling periods occurring near 6.8, 5, and 3 ky. In general, warm periods were of shorter duration than cool intervals, but a broad interval with high δ\(^{18}\)O values appears around 5.7 ky. The highest δ\(^{18}\)O values occur near 1.4 ky and last about one century. This period was followed by a gradual transition to lower δ\(^{18}\)O values that lasted from ~0.8 ky to 0.3 ky, but this interval was interrupted by several abrupt transitions towards more positive δ\(^{18}\)O values.

In Fig. 4.4 we smooth the δ\(^{18}\)O record with a moving average of 600 years after which we calculate the standard deviation of the residual δ\(^{18}\)O values over a window of 600 years. We also calculate the first derivative of the 600-year moving average. The smoothed record identifies maximum δ\(^{18}\)O variability associated with periods of warming, with four clear cycles over the last 9 ky. A sensitivity test with moving
windows of 600 ± 200 years shows that this relationship is robust on multi-centennial timescales.

To test for the presence of cyclic components in the OCNM time series we calculate the multitaper spectrum\textsuperscript{35} using the SSA-MTM Toolkit\textsuperscript{36} and we assess how the choice of age model impacts the dominant frequencies of the speleothem timeseries. Prior to this analysis, the long-term trend attributed to insolation was removed. The age model has the largest impact on the dominant millennial period which varies between 1707 in AM1 and AM3 to 1463 years in AM4 (Table 4.1), resulting in a mean period of 1626 ± 141 years. The centennial variability is characterized by significant peaks near ~640, ~233 ± 33 and ~131 ± 23 years, while significant multi-decadal periodicities centered of ~96 ± 3, ~88 ± 2, ~73 ± 5, and ~63 ± 3 years. AM2 does not have the ~640- and the ~126-year cycles, while AM3 does not have the ~73 year cycle.

To evaluate the non-stationary nature of these cycles we used the Morlet wavelet transform in which the significance of the peaks in the power spectrum is tested against an autoregressive red noise background spectrum\textsuperscript{37}. The wavelet spectral density for AM1 shows that millennial-scale variability is most pronounced during the mid-early Holocene with significant nonstationary power at the centennial and decadal frequencies identified in the MTM analysis (Fig. 5B). The absence of decadal variability in the early Holocene is due to the lower sampling interval for that period, but the periods between ~1.5 ky and ~3 ky and between 4.5 and 5.5 ky are characterized by reduced variability at both decadal and centennial timescales.

The $\delta^{13}$C values of speleothem OCNM02-1 have an average of -5‰ and a standard deviation of 1‰ (Fig. 4.6). The total amplitude ranges from -2.5 to -8.1‰, and the $\delta^{13}$C record is marked by several large events of isotopic depletion or enrichment of 3-4‰. At 1.6 ky there is a major and abrupt transition of ~3‰ from the lower values that characterized most of the Holocene to more positive values that characterized the most recent ~1600 years of growth. Other prominent and rapid changes in the $\delta^{13}$C record occurred near 0.8, 1.3, 3.1, 4.7, 5, 6.1 and 6.8 ky. The wavelet spectrum of the $\delta^{13}$C record shows a prominent ~2000-1700 year cycle which
is most clearly pronounced between 2 and 6 ky, as well as significant power at centennial timescales centered near ~300 and ~600 years.

In Fig. 4.7 we examine the relationship between the $\delta^{13}C$ and $\delta^{18}O$ variability during the Holocene by comparing the standard deviations of their residuals over a window of 600 years. Both records share similar long-term trends with high variability in the $\delta^{18}O$ record being accompanied by high variability in the $\delta^{13}C$ values. More insight can be gained by examining their cross wavelet (Fig. 4.8), which indicates that they are out of phase at millennial timescales where the $\delta^{18}O$ leads $\delta^{13}C$ by 80-90° (~400 years) but they are in-phase in the multi-centennial time band. Since both $\delta^{13}C$ and $\delta^{18}O$ are measured at the same time on the same sample, the phase relationships are not dependent on the errors associated with U-Th dating or the choice of age model.

4.10 The last interglaciation

Based on six U-Th dates with an average error of 850 years, stalagmite OCNM05-1A grew from 118.6 ky to 124.8 ky or during marine isotope stage (MIS) 5e (Fig 9). The average growth rate was 92.6 mm/ky with the fastest growth occurring between 121.7 ky and 122.7 ky when the growth rate was on average 175 mm/ky. The sampling interval varies from 5-6 years in areas of fast growth to 95 years in the oldest part of the record. The total amplitude of the $\delta^{18}O$ record is higher by 0.4‰ than during the Holocene but the average $\delta^{18}O$ value for the Holocene is nearly identical to the last interglacial period (-8.92‰ for MIS 5e and -8.99‰ for Holocene). Nevertheless the maximum $\delta^{18}O$ value for the last interglaciation (-7.84‰) is 0.52‰ higher than the maximum Holocene value (-8.36). The average $\delta^{13}C$ value is -8.7‰, which is 3.7‰ more depleted than the average $\delta^{13}C$ during the Holocene in stalagmite OCNM02-1 and 2.2 ‰ more depleted than the average Holocene $\delta^{13}C$ in stalagmite OCNM05-1B (which is adjacent to OCNM05-1A). The highest $\delta^{13}C$ values are seen in the youngest portion of the record and occur after an abrupt isotopic enrichment of ~5‰. For most of the rest of the record the $\delta^{13}C$ variability is on the order of 2 - 4‰.
Because of the relatively large differences in the time spacing between samples we divide our time series analysis into a low-frequency section in which we interpolate the entire time series at 100-years resolution, and a high-frequency component in which we use short, high-resolution intervals during the late and middle parts of the record to assess decadal climate variability. The MTM spectrum for the entire record is plotted in Fig. 4.10 and shows one significant peak at 385 years. Even though our time series is long enough (5.9 ky) to capture millennial scale variability, there are no significant millennial spectral peaks detected in the MTM analysis. The MTM spectra for the high-resolution sections of the stalagmite (Fig. 4.11) shows no significant peaks for the youngest portion of the record, while the period between 121.7 ky and 122.6 ky has two significant spectral peaks at 52 and 32 years. It would thus appear that even though the MIS 5e record shows significant variability, it was not characterized by the same cyclicity as the present interglacial period.

4.11 Discussion

The relation between modern rainwater δ\(^{18}\)O and temperature at OCNM suggests that there is a strong temperature component to the δ\(^{18}\)O. This temperature control is identifiable in the δ\(^{18}\)O record from OCNM02-1, in which there is a clear expression of the Younger Dryas cold interval \(^{40}\) and of the orbitally induced increase in winter insolation through the Holocene (Figure 12a). We thus interpret the Holocene centennial-and millennial-scale variability identified from OCNM02-1 as similarly recording winter-season temperature changes.

The variability in Holocene δ\(^{13}\)C at OCNM is controlled largely by precipitation, which in turn influences soil biomass and open- versus closed-system behavior. We suggest that at millennial timescales the δ\(^{13}\)C lag relative to δ\(^{18}\)O (Fig. 4.8) is most likely caused by a delay in vegetation response to climate change. This is supported by pollen studies in the Klamath Mountains \(^{28,34}\) as well as in the Coast and Cascade Ranges \(^{38}\) showing that vegetation can lag long-term climate change by centuries to millennia. In contrast, synchronous changes between δ\(^{18}\)O and δ\(^{13}\)C at the 300-600 year periods may reflect simultaneous changes in temperature and precipitation.
because the cave response to changes in moisture supply is rapid. Since the impact of ENSO on modern climate of the Pacific Northwest is generally associated with simultaneous changes in temperature and precipitation, this centennial variability could be modulated by ENSO.

We next compare our results with paleo-proxies that are in the vicinity of our site, which include an alkenone-based sea surface temperature (SST) reconstruction from offshore northern California (ODP 1019) and the organic carbon and charcoal influxes from Bolan Lake (~20 km SE of OCNM). The SST cooling in ODP 1020 during the Younger Dryas cooling period is consistent with the cooling inferred at OCNM, and the Bolan Lake record shows a corresponding abrupt reduction in organic productivity which lags the temperature changes by a few centuries (Fig. 4.12a). This may reflect a delayed adjustment of the lake system to climate changes, or could be associated with dating uncertainties. Similar millennial-scale variability suggests that SST warming is generally associated with higher winter temperatures at OCNM and lower organic productivity at Bolan Lake (Fig. 4.12a). Organic productivity in lakes and preservation of organic matter through time is controlled by multiple factors that can be affected by climate, including nutrient supply, summer temperature, length of the growing season, water level and pH. It is unclear which of these factors is the most important at Bolan Lake, but variations in organic productivity displays similar millennial-scale variability as the OCNM isotope record. Further evidence of synchronous climate changes in the Klamath Mountains is suggested by the record of fire frequency from Bolan Lake. The comparison between the variability of the charcoal influx into Bolan Lake, as a proxy of fire frequency in the area, and the variability of the OCNM δ¹³C record (Fig. 4.12b) shows that increases in δ¹³C variability are associated with increases in the variability of fire frequency. Since the δ¹³C record at OCNM is at least in part a reflection of changes in precipitation, this relationship suggests that higher rainfall variability is associated with increased variability in fire occurrences at Bolan Lake. Furthermore, the highest variance of fire frequency tends to occur during peak rates of warming as inferred from the OCNM δ¹⁸O record (Fig. 4.12a and b). Based on the arguments discussed above, it is apparent that these three records (ODP 1019 SST, OCNM and Bolan)
indicate coherent regional climate changes on millennial timescales in the Pacific Northwest and the eastern North Pacific during the late Pleistocene and Holocene.

The results of the spectral analysis (Fig. 4.5A, Table 4.1) reveal a series of frequencies that are similar to known cycles of solar variability such as the Gleissberg cycle (1/88 years) and the Suess (de Vries) cycle (1/210 years), which are also present in the atmospheric Δ¹⁴C record ⁴². However, the frequency with the most spectral power is at ~1/1600 years, but the origin of this cycle is not widely accepted, although a solar connection has been proposed in concert with ocean-atmosphere feedbacks ⁴³. Performing a cross-wavelet analysis of the paleo-record of Δ¹⁴C, which is interpreted to reflect changes in solar activity ⁴⁴, and our speleothem δ¹⁸O time series reveals significant coherency centered near ~2000 years that is above the 95% significance level between ~4.5 ky and ~8.5 ky (Fig. 4.13). This is also the interval where there is well defined millennial-scale variability in our record. Significant coherencies are also detected at centennial and decadal timescales that are particularly strong during the mid-Holocene, but the phasing at any of the millennial or centennial frequencies is not constant.

In North America, millennial-scale cycles of similar length as at OCNM (~1600 years) have been identified in several studies including a compilation of North American pollen sequences ¹¹,₄⁵, lacustrine sediments from Alaska ⁴⁶, and in a speleothem record from the U.S. southwest monsoon region ⁴⁷, with the latter two studies invoking solar forcing amplified by the climate system to explain the observed millennial scale variability.

There is no clear evidence that changes in solar luminosity can directly drive significant climate variations on centennial to millennial timescales ⁴⁸. However, modeling studies suggest that small changes in solar irradiance can be amplified by ocean-atmosphere interactions which then influence large-scale climatic patterns. For instance, increased solar activity may cause an increase in UV radiation in the upper stratosphere which leads to a differential warming through UV absorption by ozone, resulting in changes in atmospheric circulation that are propagated in the lower stratosphere and in the troposphere ⁴⁹. These changes affect large-scale atmosphere-ocean systems such as the Arctic Oscillation (AO)/North Atlantic Oscillations (NAO)
causing significant (but spatially heterogeneous) wintertime cooling (warming) associated with a decrease (increase) in solar irradiance. Models that studied the response of the tropical Pacific to solar forcing indicate a solar modulation of ENSO variability whereby an increase (decrease) in radiative forcing enhances (decreases) the thermal gradient between the eastern and western Pacific resulting in a “La-Niña-like (El Niño-like)” state of the tropical ocean. Simulations of the response of ENSO to solar and volcanic forcings indicate that the El Niño-like pattern is characterized by increased variability while La-Niña-like conditions correspond to a decrease in ENSO variance.

Insofar as the Pacific Northwest winter temperatures are influenced by ENSO variability, one would expect that solar modulated changes in ENSO will be present in our speleothem reconstruction. To this end, we compare the response of ENSO to solar and orbital forcing through the Holocene as derived by Emile-Geay et al. using a simplified model of the tropical Pacific. While the OCNM timeseries shares some common features with the modeled ENSO variability (e.g. the period around 3 ky) the overall agreement between the two datasets is poor in that one would expect higher probabilities of El Niños corresponding with warmer intervals at OCNM. The cross-wavelet analysis indicates coherency over a broad and irregular area near ~500 years, but with no consistent phasing, and indicates that at millennial frequencies they are nearly antiphased (although this is outside the areas with significant common power), which is the opposite of the modern climatology. We tried to improve the correlation between these two time series using a cross-correlation maximizer algorithm to obtain an objective tuning, but the correlation cannot be improved without severely violating the constraints imposed by the U-Th dating with the 2σ dating errors. These inconsistencies between our results and the modeling experiment could be related to errors in our age model, uncertainties related to the use of Δ14C as a proxy of solar variability in the model simulation or related to processes and feedbacks that were not considered in the modeling experiment.

Solar forcing may also be manifested at our site through changes in high-latitude atmospheric circulation as expressed in the AO/NAO. It has been suggested that solar-modulated ENSO variability may have an impact on atmospheric circulation in...
the North Atlantic whereby increased solar irradiance causes more La Nina-like conditions in the tropical Pacific, which drives the NAO towards a more positive phase\textsuperscript{53}. Positive phases of the NAO are associated with cooling over North Atlantic\textsuperscript{57}, and one would therefore expect episodes of high influx of ice-rafted debris (IRD) into the North Atlantic to be associated with a La Nina-like state of the tropical Pacific. By comparing a composite record of IRD fluxes in the North Atlantic\textsuperscript{43} with the modeled response of the tropical Pacific to solar and orbital forcing\textsuperscript{53} we observe that they are antiphased at millennial timescales (Fig. 4.13). We also note that at similarly long timescales, the episodes of high IRD fluxes into the North Atlantic correspond to higher winter temperatures at OCNM (Fig. 4.15), which combined with the antiphasing of ENSO and IRD fluxes is inconsistent with the ENSO forcing mechanism of North Atlantic climate described above.

Comparison of our reconstruction with other records in the Northern Hemisphere, (Fig. 4.15) indicates that on millennial timescales the winter climate in southwestern Oregon is most similar with the North Atlantic, although opposite in sign. Within dating uncertainties, cooler winters at our site are also correlated with decreases in precipitation in southwestern U.S.

\textbf{4.12 Last Interglacial Period}

Compared with the Holocene, relatively little information exists about climate variability in western North America and the Pacific Ocean during the last interglaciation. A pollen record from Carp Lake in the Cascade Range indicates that MIS 5e was characterized by open forest conditions and the pollen assemblage suggests warmer and drier conditions compared with modern climate\textsuperscript{1}. A low-resolution alkenone record off the coast of northern California shows warmer SSTs during the last interglaciation, reaching a peak at 122 ky\textsuperscript{3}, while a sediment core from the Santa Barbara basin reveals significant millennial climate variability during the last interglaciation\textsuperscript{58}.

To the extent that precipitation seasonality during the last interglacial was the same as during the Holocene (i.e. cool-wet winters and warm-dry summers), the $\delta^{18}$O
values in speleothem OCNM05-1 suggest overall similar winter conditions in the Pacific Northwest during the last interglacial period, except for a short warmer interval centered around 120.8 ky. However, the pattern of variability differs from that of the Holocene as highlighted in the MTM analyses. Lower $\delta^{13}C$ values than during the Holocene characterized most of the last interglaciation, which suggests wetter conditions and/or more biomass relative to the present interglacial period.

4.13 Conclusions

We present the first absolutely dated Holocene winter climate record in the western U.S. This reconstruction indicates that climate in southwestern Oregon is sensitive to long-term changes in solar insolation. We also identify significant millennial and centennial climate variability and note that the highest variability is associated with warming trends. Similarly, fire frequency in the Klamath Mountains appears to have the highest variance during peak rates of warming and greatest variability in temperature and moisture, as inferred from our reconstruction. To the extent that the past can serve as an analog to the future, this implies that global warming may be associated with higher climate variability in the future in the Pacific Northwest. The phase relationship between speleothem $\delta^{18}O$ and $\delta^{13}C$ suggests a delayed response of vegetation to climate change at the millennial time band, and synchronized changes in temperature and precipitation on multi-centennial timescales. On millennial timescales, winter temperatures at OCNM are antiphased with influxes of IRD in the North Atlantic, particularly during the early- and mid-Holocene.

Unlike during the Holocene, there is no significant millennial climate cycllicity during the last interglacial. Comparison of $\delta^{18}O$ and $\delta^{13}C$ values suggest similar overall temperatures, but possibly wetter conditions and/or higher soil productivity compared with the present interglaciation.
4.14 Acknowledgements

We thank John Roth and Elizabeth Hale for facilitating our research at OCNM and Xianfeng Wang for help with U-Th dating. This research was funded by the National Science Foundation.
4.15 References


Figure 4.1. Distance-age relationship for dated horizons in speleothem OCNM02-1 showing the three age models discussed in text. Dashed gray line shows a 4th order polynomial fit through the data. The upper panel indicates changes in growth rate through time.
Figure 4.2. Replication test for the low-resolution $\delta^{18}$O and $\delta^{13}$C record of stalagmite OCNM02-1. Data older than 11 ky in OCNM02-1 are from Vacco et al.\textsuperscript{40} Shown at the bottom of the graph are the corresponding U-Th dates with their respective $2\sigma$ errors. The effect of lower sea-level on speleothem $\delta^{18}$O values for the period between 11 and 13 ky was removed as described in Vacco et al.\textsuperscript{40}. 
Figure 4.3. December insolation at 45°N (red-dashed line) plotted against the OCNM δ¹⁸O record. The lower panel shows the δ¹⁸O record with the effect of insolation removed. Smooth lines show the average δ¹⁸O values over a 600-year moving window, and the stars indicate the δ¹⁸O values of two soda-straw stalactites that were actively growing in the cave in 2007.
Figure 4.4. The raw and smoothed $\delta^{18}$O record (upper panel) plotted together with the residual $\delta^{18}$O values after the mean was removed (see text). Also shown are the standard deviation of the residual and the first derivative of the smoothed $\delta^{18}$O record over a 600 year moving window.
Table 4.1. Significant frequencies above the 99% significance level in the MTM spectrum identified in the three age models discussed in text.

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Figure 4.5. A. Adaptively weighted multitaper (MTM) spectrum of the OCNM $\delta^{18}$O time series. Black line shows the reshaped spectrum in which the contribution from harmonic signals has been removed based on a 90% F test significance criterion. Smooth-dashed gray curve shows the robustly estimated red-noise background at 99% significance level and the numbers on the graph correspond with significant periods that are above this 99% threshold. B. Continuous wavelet power spectrum of the $\delta^{18}$O Holocene record. Thick black contour lines enclose regions of greater than 95% confidence against red noise and the lighter shades define the cone of influence where edge effects might distort the signal. Both the x and y-axes are in years before present (B.P.)
Figure 4.6. The raw OCNM $\delta^{13}$C record (black line) and the smoothed $\delta^{13}$C values over 600 year moving window. The lower panel shows the continuous wavelet power spectrum of the $\delta^{13}$C record indicating significant power at ~2000-1700 years as well as at ~300 and ~600 years. Note that in the upper panel time is expressed in years x 1000.
Figure 4.7. Comparison between the smoothed $\delta^{18}$O and $\delta^{13}$C records and the standard deviation of their residuals.
Figure 4.8. Cross wavelet transform of the Holocene $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ records. Relative phase relationships between the two time series are shown by arrows, with in-phase pointing right, anti-phase pointing left, $\delta^{18}\text{O}$ record leading $\delta^{13}\text{C}$ by 90° pointing straight up, $\delta^{18}\text{O}$ record lagging $\delta^{13}\text{C}$ by 90° pointing straight down. The areas that are above the 95% confidence threshold are outlined by the thick black contours. X and Y axes labeled in years.
Figure 4.9. Age model for the stalagmite OCNM05-1 (top) plotted together with the δ¹³C and δ¹⁸O time series.
Figure 4.10. A. The raw last interglacial period $\delta^{18}O$ time series (gray) and the interpolated time series (black) used in the MTM calculation shown in B. The two gray circles on the left show the $\delta^{18}O$ values of two soda straw stalactites that were actively growing in the cave in 2007. B. MTM spectrum of the 100-year interpolated time series showing one significant peak above the 95% red noise level. Spectrum procedure is the same as in Fig. 4.5A.
Figure 4.11. High-resolution δ¹⁸O time series for two periods during the last interglaciation: A. period between 118.6 ky and 119.5 ky; B. period between 121.7 ky and 122.6 ky Gray lines show the raw δ¹⁸O record and black lines indicate the interpolated time series used in the MTM calculation. The MTM spectrum is shown below each time series (panels C and D) and the MTM procedure is the same as in Fig. 4.2B.
Figure 4.12a. The smoothed $\delta^{18}O$ Holocene record presented in this paper (lower panel) combined with December solar insolation, 600-year smoothed record of organic productivity in Bolan Lake$^{28}$, July insolation at 45°N$^{32}$, the SST reconstruction from offshore northern California$^4$ and the raw percent organic carbon from Bolan Lake$^{28}$. Also shown are the winter (dashed line) and summer (dash-dotted line) insolation at 45°N$^{32}$ while the age control points and their associated errors are shown under each record. Note that the y-axis for Bolan Lake is reversed. The inset shows the response of these three records during the Younger Dryas and the late glacial period in which solid-red lines represent the smoothed records over a moving window of 600 years.
Figure 4.12b. Comparison between the OCNM $\delta^{13}$C residual with the residual of charcoal influx into Bolan Lake. The lower panel shows the standard deviation of the OCNM $\delta^{13}$C residual (solid line) and the standard deviation of Bolan Lake charcoal influx, both calculated over a moving window of 600 years.
Figure 4.13. Cross-wavelet transform of the OCNM $\delta^{18}O$ timeseries with the $\Delta^{14}C$ record, the modeled response of ENSO to orbital and solar forcing, and the flux of HSG in the North Atlantic. Also shown is the cross-wavelet transform between the HSG flux and the modeled ENSO behavior. Prior to the analysis the $\Delta^{14}C$ record was detrended by subtracting a six-order polynomial fit from the data. Relative phase relationships between the two time series are shown by arrows as in Fig. 4.8. Periods for both the x and y axes are in years B.P.
Figure 4.14. Comparison of the OCNM isotope timeseries with the probability of a large ENSO event over a 200-year time window. The long-term trend of winter insolation has been removed from both timeseries.
Figure 4.15. Comparison of OCNM $\delta^{18}$O with other records from the North Atlantic and western North America. From the top down these are the biogenic silica from Arolik Lake in Alaska\textsuperscript{46}, percentage of hematite stained grains (HSG) into the North Atlantic\textsuperscript{43}, OCNM $\delta^{18}$O record, SST record from the Santa Barbara basin\textsuperscript{60}, and speleothem-based precipitation reconstruction in southwestern U.S.\textsuperscript{47}. All timeseries have been detrended and resampled at equally spaced 100 year intervals.
Chapter 5

Conclusions

We used the growth of stalagmites in a cave from southwestern Oregon as a first-order indicator of past climate conditions over the last 380 000 years. Stalagmite growth requires the presence of liquid water as well as vegetation on top of the cave, and both of these parameters will be influenced by climate. Because the growth of any one stalagmite can be interrupted due to localized factors along the dripwater path that are unrelated to climate, we used five stalagmites collected from different locations within the cave to robustly constrain the times in the past when climate conditions were favorable to speleothem growth. The majority of speleothem growth occurred during interglacial climates corresponding to Marine Isotope Stages (MIS) 1, 5e, 7c and 9, indicating that at these times climate conditions allowed sufficient aquifer recharge, and the soil development and vegetation cover on top of the cave were extensive enough to promote calcite dissolution in the epikarst zone and calcite deposition in the cave. Speleothems also grew during past glacial periods near the penultimate glacial-interglacial transition and during MIS 3. The start of speleothem growth near 135 ± 1.2 ka suggests an early start of the last interglaciation, consistent with other records from the northern hemisphere (e.g. Stirling et al., 1998, Henderson and Slowey, 2000). Speleothem growth occurred during MIS 3 and 4, suggesting milder climate conditions during some portions of the last glacial period. However, in general glacial periods were characterized by non-deposition or speleothem dissolution.

To investigate the mechanisms responsible for growth cessation during glaciations we evaluated the magnitude of climate change during the peak of the last glaciation (LGM), 21 000 years ago, using a regional climate model. The modeling results indicate a depression of mean annual temperature of 4°C, insufficient to allow permafrost to form on top of the cave. Instead, our simulation indicates that
speleothem growth cessation during the LGM was likely caused by a drastic reduction in aquifer recharge induced by the overall drier LGM conditions and the presence of frozen topsoil from late fall to early spring.

To simulate the cave temperature history during the last 100 000 years we used a thermal advection-diffusion model and the results are consistent with the regional climate model in indicating an LGM temperature depression of $\sim 4^\circ$C. The temperature for the rest of the last glaciation did not drop at any time below 0°C, suggesting that permafrost was not a likely controlling factor on speleothem growth.

In an effort to investigate the environmental controls on oxygen isotopes in precipitation, chapter 2 presents a study of isotopic variability in rainfall from southwestern Oregon. Because the relationship between temperature and rainfall is variable and site dependent, interpretation of oxygen isotopes in speleothems and other environmental archives requires an evaluation of local controls on stable isotopes in precipitation.

The influence of changes in storm trajectories on $\delta^{18}$O composition of rainfall was assessed with a Hybrid Single Particle Lagrangian Integrated Trajectory (HYSPLIT) model from NOAA’s Air Resource Laboratory. Storms reaching southwestern Oregon originated from a wide area of the North Pacific Ocean, stretching from the subtropical to subpolar areas. However, on the synoptic timescales analyzed in this study, there was no consistent relationship between storm trajectory and the isotopic composition of rainfall. Temperature at the time of sample collection explains 30% of $\delta^{18}$O variability, and the air temperature along the storm track averaged over the last 8-12 hours before arrival shows the highest correlation with rainwater $\delta^{18}$O and explains a similar amount of variance as the local temperature. We find that the local slope of temperature vs. rainfall $\delta^{18}$O is 0.7 ‰/°C and conclude that this high slope is due to the dominance of winter rainfall at our site.

We did not find a statistically significant relationship between precipitation amount and rainfall $\delta^{18}$O. However, changes in specific humidity along the storm track indicate some level of condensation and entrainment of moisture along the storm paths, especially in the case of storms with high $\delta^{18}$O values.
Because the slope of temperature vs. rainfall $\delta^{18}O$ is 0.7 ‰/°C is larger than the fractionation factor between cave dripwater $\delta^{18}O$ and speleothem calcite (-0.2 ‰/°C), speleothems in southwestern Oregon are likely to preserve in their oxygen isotope composition past changes in temperature.

Chapter 3 presents a study of climate variability in southern Oregon during the Holocene and the last interglacial period. To evaluate the use of speleothems as paleoclimate archives we first established that speleothem calcite was deposited in equilibrium with cave dripwaters by performing Hendy tests and by replicating a low-resolution record in three separate stalagmites. The age control for our high-resolution Holocene reconstruction was given by 10 U-Th dates and the timeseries of $\delta^{18}O$ variation, which we interpret as reflecting mainly temperature, shows that winter temperatures were sensitive to long-term changes in solar insolation. The highest winter temperatures occurred around 1400 years before present, while the coolest winter conditions occurred during the early Holocene. Temperatures varied on millennial timescales in a quasi-cyclical fashion, with a significant spectral peak centered near 1700 years. This millennial-scale variability was particularly pronounced during the early- and mid-Holocene. Significant centennial and decadal variability also characterized our temperature reconstruction, with spectral peaks that might correspond to known cycles of solar variability. Since changes in solar irradiance are too small to have a direct effect on temperature in southwestern Oregon, and because ENSO variability has a measurable impact on Pacific Northwest temperatures, we compare our climate reconstruction with a model simulation of the tropical Pacific climate which was forced by changes in solar irradiance and insolation during the Holocene. The overall agreement between the two timeseries is poor and may be caused by uncertainties in our temperature reconstruction or could be related to processes and feedbacks not represented in the modeling experiment or uncertainties in the use of $\Delta^{14}C$ as a proxy for solar variability.

The highest temperature variability in southern Oregon was consistently associated during the Holocene with periods when the climate was warming, implying that, in the context of the present warming trend, future climates may be characterized by more unstable conditions.
By comparing the results of our Holocene reconstruction with data obtained from a stalagmite that grew during the last interglaciation we suggest that lower overall $\delta^{13}$C values that characterized MIS 5e indicate more soil development and vegetation cover and/or wetter climate. The average $\delta^{18}$O values during the last interglaciation are similar to the average Holocene values, indicating similar temperature conditions, but the millennial and centennial variability was significantly different that during the Holocene.

References


Rozanski, K., Áraguas-Áraguas, L. and Gonfiantini, R., 1993. Isotopic patterns in modern global precipitation. In: P.K. Swart (Editor), Climate Change in


Appendix
Figure A1. Images of speleothems discussed in text. From left to right, these are OCNM05-1A and 1B, OCNM02-2, OCNM02-1 and OCNM07-1. Black lines indicate the locations of samples for $^{230}$Th dating, with the exception of the bracket on stalagmite OCNM02-1 which shows the short growth interval during the last glacial period where we had samples at closely spaced intervals (Table A2).
Table A1. $^{230}$Th dating results for OCNM speleothems

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$\lambda_{230} = 9.1577 \times 10^{-6} y^{-1}$, $\lambda_{234} = 2.8263 \times 10^{-6} y^{-1}$, $\lambda_{238} = 1.55125 \times 10^{-10} y^{-1}$, $^{234}U = \left(\frac{^{234}U/^{238}U}{\text{activity}} - 1\right) \times 10000$. ** $^{234}U_{\text{initial}}$ was calculated based on $^{230}$Th age (T), i.e., $^{234}U_{\text{initial}} = \left(\frac{^{234}U}{^{238}U}\right)_{\text{activity}} \times e^{\lambda_{234}T}$. Corrected $^{230}$Th ages assume the initial $^{230}$Th to $^{232}$Th atomic ratio of 4.4 ± 2.2 x 10^{-6} except for stalagmite OCNM02-1 where based on the isochron technique (Vacco et al., 2005) the ages were calculated with a ratio of 12 ± 3 x 10^{-6}. The errors are arbitrarily assumed to be 50%. The error is 2σ error. *** Ages reported in Vacco et al., (2005).