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Expression of the "4.2 ka event" drought in the southern Rocky Mountains, USA

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https://doi.org/10.5194/cp-2021-149
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Abstract

The use of the climatic anomaly known as the "4.2 ka event" as the stratigraphic division between the mid- and late Holocene has prompted debate over its impact, geographic pattern, and significance. The anomaly has primarily been described as abrupt drying, but evidence of hydroclimate change at ca. 4 ka is inconsistent among sites globally, and few sites in North America document a major drought. Climate records from the southern Rocky Mountains demonstrate the challenge with diagnosing the extent and severity of the anomaly. Dune-field chronologies and a pollen record in southeast Wyoming reveal several centuries of low moisture at around 4.2 ka and prominent low stands in lakes in Colorado suggest the drought was unique amid Holocene variability, but detailed carbonate oxygen isotope ($\delta^{18}O_{carb}$) records from Colorado do not record it. We find new evidence from $\delta^{18}O_{carb}$ in a small mountain lake in southeast Wyoming of an abrupt reduction in effective moisture or snowpack from approximately 4.2–4 ka that coincides in time with the other evidence from the southern Rocky Mountains and the western Great Plains of regional drying at around 4.2 ka. We find that the $\delta^{18}O_{carb}$ in our record may reflect cool-season inputs into the lake, which do not appear to track the strong enrichment of heavy oxygen by evaporation during summer months today. The modern relationship differs from some widely applied conceptual models of lake-isotope systems and may indicate reduced winter precipitation rather than enhanced evaporation at ca. 4.2 ka. Inconsistencies among the North American records, particularly in $\delta^{18}O_{carb}$ trends, thus show that site-specific factors can prevent identification of the patterns of multi-century drought. However, the prominence of the drought at ca. 4 ka among a growing number of sites in the North American interior suggests it was a regionally substantial climate event amid other Holocene variability.

https://doi.org/10.5194/cp-2021-149
Preprint. Discussion started: 19 November 2021


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1. Introduction

Rapid climate changes are well documented in the late Pleistocene and early Holocene, such as during the Younger Dryas chronozone and at 8.2 ka (thousands of years before present; Alley et al., 1997; Clark et al., 1999; Von Grafenstein et al., 1998), but mid- to late-Holocene changes are less well understood (Wanner et al., 2008, 2011). One potential abrupt change during this time, a multi-century climatic anomaly known as the "4.2 ka event," has been used as the benchmark for the stratigraphic division between the mid- and late Holocene (Walker et al., 2019). Consequently, the 4.2 ka event has become a topic of scrutiny with debate over its impact, geographic pattern, and significance (Bradley & Bakke, 2019; Weiss, 2016, 2019). The ostensibly global event has primarily been described as a dry episode at low and mid-latitudes (Booth et al., 2005; Nakamura et al., 2016; Di Rita & Magri, 2019; Scuderi et al., 2019; Xiao et al., 2018). Consistent with spatial variation expected from climate variability that shifts atmospheric waves and dynamics, however, some regions show increased precipitation (Huang et al., 2011; Railsback et al., 2018) or no change (Roland et al., 2014).

Despite the widespread examination of the 4.2 ka event, its cause and significance amid other millennial-to-centennial climate variability during the Holocene remain unknown. Recent simulations have produced similar patterns of extended drought in the northern hemisphere without external forcings such as insolation changes or volcanism (Yan & Liu, 2019), and others confirm that multi-decadal megadroughts can arise through internal climate variability without changes in boundary conditions (Ault et al., 2018). Internal climate dynamics and feedbacks could also interact with stochastic variability and external forcing to produce such events without consistent or linear relationships to the forcing; forcing may only have a modest probability of

https://doi.org/10.5194/cp-2021-149
Preprint. Discussion started: 19 November 2021


Climate of the Past Discussions

triggering rapid climate changes (Renssen et al., 2006). Less clear is how unusual or frequent prolonged 'megadroughts' may be within the Holocene across different regions.

That such droughts can occur stochastically indicates the 4.2 ka event could be an example of typical late-Holocene climate variability at multi-century time scales (Shuman & Burrell, 2017), but in at least some regions, the event may be exceptional within the spectrum of Holocene variability. Evidence for a major hydroclimate change at ca. 4 ka has been growing in the North American midcontinent (Booth et al., 2005; Carter et al., 2018; Dean, 1997; Denniston et al., 1992; Halfen & Johnson, 2013; Jiménez-Moreno et al., 2019), and adjacent regions, such as the northeastern United States, where it is recorded as part of a series of Holocene wetting and drying events (Newby et al., 2014; Shuman et al., 2019; Shuman & Burrell, 2017). The event's significance or uniqueness has been difficult to verify, however, in North America where few sites document the anomaly compared to other regions of the mid-latitudes globally (Ran & Chen, 2019; Zhang et al., 2018).

Records from the southern Rocky Mountains demonstrate the challenge. In the mid-latitude Rocky Mountains, only dune and pollen records have been explicitly interpreted to show the 4.2 ka event. Initial recognition in North America derived from the timing of the reactivation of the Ferris, Seminoe, and Casper Dune Fields in east-central Wyoming (Fig. 1; Booth et al., 2005; Halfen et al., 2010; Stokes & Gaylord, 1993), but the extent of the drought has been unclear because other dune-field chronologies in the adjacent western Great Plains do not clearly document the drought (Dean, 1997; Halfen & Johnson, 2013; Mason et al., 1997). More recently, Carter et al. (2013, 2017a, 2018) used fossil pollen from Long Lake in the Medicine Bow Mountains, south of the Wyoming dune fields (Fig. 1), to identify a 150-year interval of increased temperature and decreased precipitation centered at 4.2 ka. The inferred precipitation

https://doi.org/10.5194/cp-2021-149
Preprint. Discussion started: 19 November 2021

93 reductions were largest in springtime (Carter et al., 2018), when snowfall in the southern Rocky
94 Mountains is highest today (Mock, 1996). Consistent with this interpretation, prominent
95 stratigraphic evidence of lake-level changes in Colorado and Wyoming lakes could indicate that
96 low water phases at ca. 4.2 ka were one of the most prominent hydrologic changes during the
97 Holocene (Jiménez-Moreno et al., 2019; Shuman et al., 2009; Shuman et al., 2014, 2015). It
98 stands out as one of the only multi-centennial features in a summary of low lakes in the Rocky
99 Mountains during the late-Quaternary (Shuman and Serravezza 2017).
100 By contrast, the 4.2 ka event does not appear in stable oxygen isotope records from lakes
101 in the same region, such as detailed carbonate-$\delta^{18}O$ ($\delta^{18}O_{carb}$) records from Bison and Yellow
102 lakes, Colorado (Fig. 1; Anderson, 2011, 2012). Widely applied conceptual models of lake-
103 isotope systems indicate that hydrologic controls on isotope budgets and the timing of carbonate
104 formation should play an important role in how the event was recorded, but that the isotopic
105 response should vary predictably by hydrologic setting (e.g., Anderson et al., 2016; Leng &
106 Marshall, 2004; Talbot, 1990). According to such models, long lake-water residence times and
107 high rates of evaporation cause hydrologically closed lakes (i.e., terminal basins) to record shifts
108 in effective moisture (precipitation – evaporation) because endogenic carbonates will typically
109 precipitate in evaporated, ^{18}O-rich water during the warm summer months. Drought could affect
110 such a lake-isotope system by both increasing evaporation and changing seasonal precipitation,
111 such as by reducing snowpack. In hydrologically open lakes with short residence times, the
112 continual replacement of evaporated water creates isotopic sensitivity primarily to the seasonal
113 balance of precipitation without a strong evaporation effect. Many lakes fall somewhere between
114 fully hydrologically open and closed and additional site-specific influences may also override
115 such expectations. Consequently, not all stable oxygen isotope records from lakes may have been

https://doi.org/10.5194/cp-2021-149
Preprint. Discussion started: 19 November 2021


Climate of the Past Discussions

116 sensitive to the specific climate variables that changed at 4.2 ka. Modern lake-water
117 measurements can help to identify the relative influences of different controls (Fig. 2; Anderson
118 et al., 2016).

119 Here we present a new $\delta^{18}O_{carb}$ record from Highway 130 Lake (HL) in southeast
120 Wyoming near where other Holocene paleohydrological and paleoecological records have been
121 developed (Mensing et al., 2012; Minckley et al., 2012; Brunelle et al., 2013). HL is an
122 intermittently closed subalpine lake in the Medicine Bow Mountains, within 20 km of Long Lake
123 where fossil pollen indicates a prolonged 'megadrought' at 4.2 ka (Fig. 1; Carter et al., 2018).
124 The lake is also <60 km from Upper Big Creek Lake, Colorado, where a prominent
125 paleoshoreline detected in geophysical surveys and cores indicates low water after 4.7 ka (Fig. 1;
126 Shuman et al., 2015). Previous work at HL indicates a strong influence of evaporation on the
127 lake and its water isotopes, which we compare with Bison and Yellow lakes in Colorado (Fig. 2;
128 Liefert et al., 2018). We discuss how dissimilarities in $\delta^{18}O_{carb}$ among lakes, possibly driven by
129 non-climatic factors, could complicate interpretations of the patterns of past hydroclimate
130 changes including megadroughts and Holocene trends. Together these outcomes may clarify the
131 timescales on which drought operates within a critical headwater area of North America, but also
132 confirm that interpretations of stable isotope records of past hydroclimate changes may depend
133 heavily on site-specific dynamics.
134
135 2. **Site description**

136 HL (41°21'05" N, 106°15'50" W; 3,199 m a.s.l. (above sea level)) fills a shallow
137 depression in the uneven terrain covering the Libby Creek watershed (12 km^2 surface area) in the
138 Snowy Range, a southwest trending subsection of the Medicine Bow Mountains in southeast

https://doi.org/10.5194/cp-2021-149
Preprint. Discussion started: 19 November 2021


Climate of the Past Discussions

Wyoming (Fig. 1). Around HL, subalpine coniferous forests interspersed with open meadows grow on thin glaciated soils and tills between the frequent outcroppings of the underlying siliceous metadolomite (Houston & Karlstrom, 1992; Musselman et al., 1992). Southeast Wyoming has a semi-arid climate, but high elevations in the Medicine Bow Mountains receive about 1,000 mm of precipitation each year, with approximately 70% of annual totals falling as snow from October to June (Mock, 1996). Local average wind speeds are high (~5 m/s) and minimum winter and maximum summer temperatures typically reach -23°C and 21°C, respectively.

The surface watershed around HL occupies ~0.45 km^2, while the lake has a surface area of ~0.02 km^2, a maximum (spring) water depth of ~200 cm, and declines in water level by ~30 cm from July to late October (Liefert et al., 2018). Ice covers HL from approximately October to May and stream connections shut off in June following spring flooding. Measurements reveal no thermal stratification because of the shallow water depth, flat-bottom bathymetry, and high average wind speeds, which promote mixing throughout the water column (Bello & Smith, 1990; Stewart & Rouse, 1976). Liefert et al. (2018) found that evaporation could account for as much as 83% of the seasonal water loss at HL, though the stable water level and temperature compared to nearby lakes of similar size and depth indicates shallow groundwater flow-through driven by seasonal precipitation and infiltration (Rautio & Korkka-Niemi, 2011; Rosenberry & LaBaugh, 2008).

3. **Methods**

To measure the modern oxygen and hydrogen isotope compositions of the lake water ($\delta^{18}O$ and δD, respectively) and specific conductance, water samples were collected at

https://doi.org/10.5194/cp-2021-149
Preprint. Discussion started: 19 November 2021

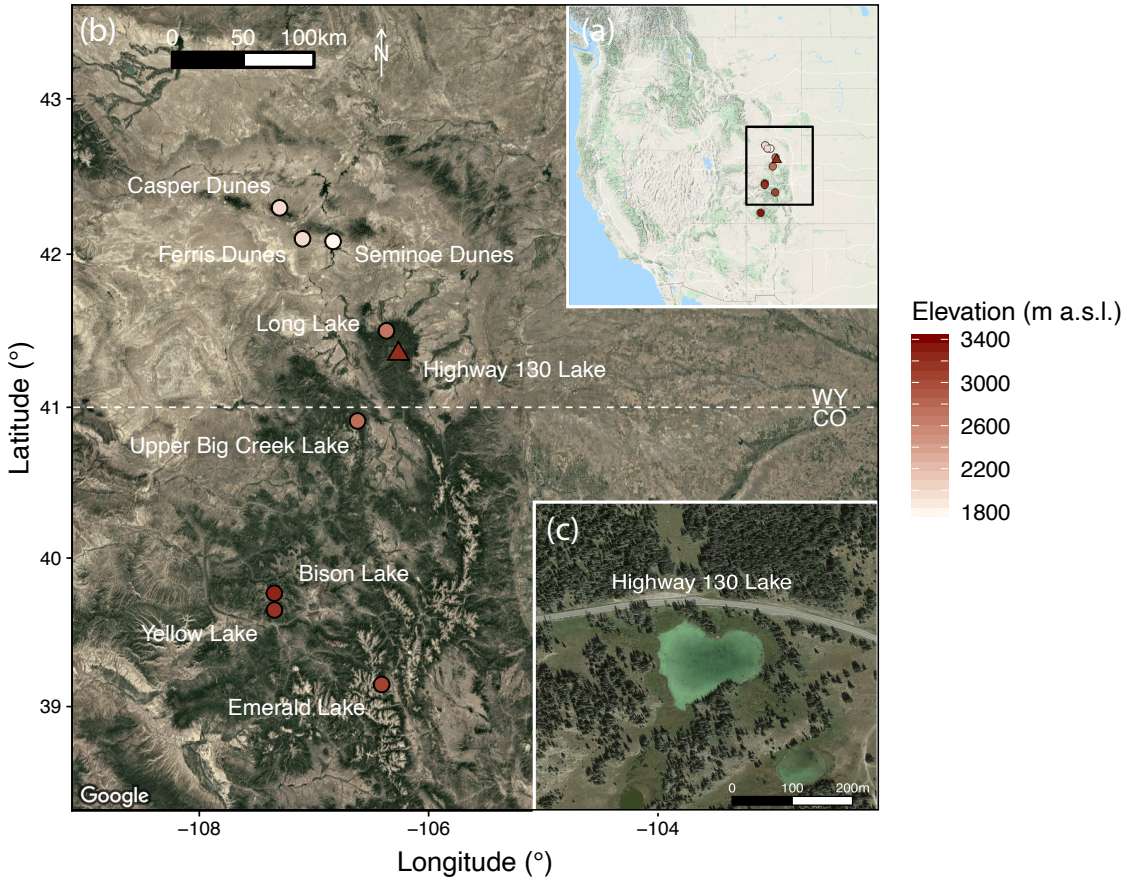

Figure 1. Locations of study site and related climate records. (**a**), Highway 130 Lake (triangle) and related climate records (circles) lie within the southern Rocky Mountains, a critical headwater area in the western United States that contributes snowmelt to the Colorado and North Platte Rivers. (**b**), Study site locations in the Colorado Front Range and Medicine Bow Mountains, southeast Wyoming. Little Molas Lake, a comparator site in the inset map (a), lies south of the focal region (b). (**c**), Highway 130 Lake lies within the Snowy Range, a subsection of the Medicine Bow Mountains. Google images (© Google Maps 2021) were acquired using the ggmap package in R (Kahle & Wickham, 2013).

162 approximately biweekly intervals from June to October in 2017. Additional samples of snowfall,
163 snowpack, rain, and groundwater (from springs and wells) were collected to measure the range in

https://doi.org/10.5194/cp-2021-149
Preprint. Discussion started: 19 November 2021

water isotope values of the watershed's hydrologic components. Isotopic ratios were measured at the University of Wyoming Stable Isotope Facility using a Picarro L2130-I Cavity Ring Down Spectrometer and specific conductance was measured using a YSI Multiparameter Water Quality Meter. We acquired meteorological data from SNOTEL stations near HL at Brooklyn Lake, Wyoming (ID 367; 3,121 m a.s.l.; 41.36 °N, -106.23 °W), and at Bison Lake, Colorado (ID 345; 3,316 m a.s.l.; 39.76 °N, -107.36 °W), to compare the modern ratios of snow/rain that control the seasonal balance of precipitation at the lakes.

In October 2016 we installed a pressure transducer (Onset HOBO U20 Level Data Logger) to measure the water level of HL at 30-min intervals; freezing conditions required that we secure the transducer to the lakebed inside a bladder filled with antifreeze. To compensate for barometric pressure changes we adjusted the transducer data using pressure measurements from the nearby Glacier Lakes Ecosystem Experiments Site Brooklyn Tower Ameriflux site (GLEES Tower; US-GLE: https://ameriflux.lbl.gov/sites/siteinfo/US-GLE; 41°21'57" N, 106°14'23" W; 3,191 m a.s.l.). In late January 2017, we installed a conductivity data logger (Onset HOBO U24 Conductivity Data Logger) at the same location and water depth as the pressure transducer to measure the range in conductivity (converted to specific conductance at 25 °C) at 30-min intervals of the unfrozen water underlying the ice cover to examine the seasonal patterns of water chemistry that influence carbonate formation.

At the same time, we collected a 70-mm diameter sediment core with a modified Livingston piston corer from the center of HL where the combined water and ice depth reached approximately 90 cm; we used this depth to calibrate the pressure transducer. The organic and carbonate content of contiguous 1-cm intervals of the sediment core were measured by weighing the residual sediment after burning the samples at 550 and 1000 °C, respectively. One-cm^3 sub-

https://doi.org/10.5194/cp-2021-149
Preprint. Discussion started: 19 November 2021


Climate of the Past Discussions

187 samples were isolated from each interval after the 550 °C burn for isotopic analysis to remove
188 organic matter; comparison with organic removal using oxidizing agents indicated no additional
189 fractionation. The sub-samples were also sieved using a 63-μm mesh to isolate the fine fraction
190 for isotopic analysis using a Thermo Gasbench coupled to a Thermo Delta Plus XL isotope ratio
191 mass spectrometer at the University of Wyoming Stable Isotope Facility. X-ray powder
192 diffraction confirmed that the samples contained only calcite. Ostracod tests were present in less
193 than 10 of the 300 samples. We report $\delta^{18}O_{carb}$ in the per mil (‰) notation relative to the Vienna
194 Pee Dee Belemnite (VPDB) standard.
195 We isolated sedimentary charcoal (>125 μm) and conifer needles from the sediment core
196 for radiocarbon analyses to estimate sedimentation rates and oxygen isotope chronology
197 calibrated to radiocarbon years using intcal13 (Reimer et al., 2013) and the age-depth model was
198 generated using Bchron (Parnell et al., 2008). Radiocarbon samples were analyzed at the
199 University of California Irvine Keck Carbon Cycle facility.
200
201 4. **Results**
202 **4.1** ***Modern water-chemistry and level measurements***
203 Lake-water $\delta^{18}O$ and δD in HL increased during the ice-free season from -17.8‰ and -
204 132‰ (sampled in late June) to -10.8‰ and -94.2‰ (sampled in late October), respectively
205 (black circles, Fig. 2). The local evaporation line (LEL) defined by the HL samples (thick black
206 line, Fig. 2) traces the LEL defined by samples from lakes in the Colorado Front Range (red
207 dashed line, Fig. 2; Henderson & Shuman, 2009). Several consecutive years of measurements
208 reveal that isotope values at HL are consistent from year to year. The LEL's deviation from both
209 the global meteoric water line (GMWL; Fig. 2) and isotope composition of the hydrologic inputs

https://doi.org/10.5194/cp-2021-149
Preprint. Discussion started: 19 November 2021


Climate of the Past Discussions

(open symbols, Fig. 2) indicates a strong evaporative influence. $\delta^{18}O$ and δD values at HL also indicate stronger fractionation by evaporation compared to representative warm-season isotope compositions measured at Bison and Yellow lakes (thick purple and yellow lines, Fig. 2; Anderson, 2011, 2012), which remained closer to the composition of meteoric waters. Longer lake-water residence time and higher evaporation in HL thus appear to produce a greater range of warm-season isotope compositions compared to Bison and Yellow Lakes.

The different lake-water-$\delta^{18}O$ values among the lakes contrasts with their similar seasonal precipitation patterns. The modern ratio of snow/rain, which can determine the mean precipitation and lake-water $\delta^{18}O$, is comparable in the watersheds of HL and Bison Lake (inset plot, Fig. 2). Other modern differences among the lakes, which all have surface areas of < 0.1 km^2, include that the maximum water depth of HL is several meters shallower than Bison and Yellow Lakes (Anderson, 2012) and that the summer lake-water temperatures in HL typically range from 8–12 °C, which is cooler than the epilimnion at Yellow Lake (Anderson, 2012). HL is also several degrees cooler than nearby lakes also without thermal stratification (Liefert et al., 2018).

Continuous measurements of specific conductance began in early February when the combined water and ice depth reached approximately 90 cm (Fig. 3). Specific conductance increased from 700 µS/cm to 1,115 µS/cm by early April while the lake surface was frozen. The specific conductance fell below 500 µS/cm as the lake flooded with snowmelt in early May. Specific conductance ranged from 250–300 µS/cm after the conductivity data logger was removed in late June and before the lake froze over in the fall, and the water depth stayed between 100–150 cm, which was low compared to previous years.

https://doi.org/10.5194/cp-2021-149
Preprint. Discussion started: 19 November 2021

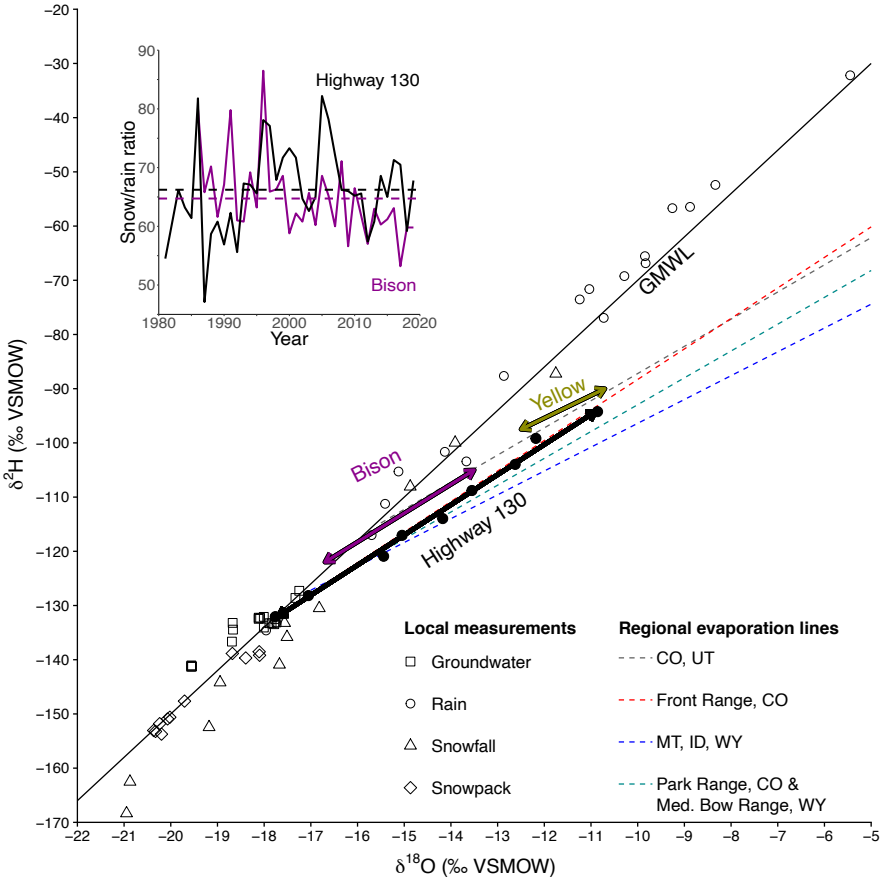

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Figure 2. Modern measurements of $\delta^{18}O$ and δD. Regional evaporation lines (dashed lines) intersect the global meteoric water line and represent the linear regression of lake-water isotope compositions in a region (Henderson & Shuman, 2009; Anderson et al., 2016). Isotopic measurements from the study watershed (open symbols) show the range in isotope compositions of hydrologic inputs to Highway 130 Lake from the watershed. Arrows represent the range in modern $\delta^{18}O$ and δD values of Highway 130 Lake (black), Bison Lake (purple; Anderson, 2011), and Yellow Lake (yellow; Anderson, 2012) throughout the ice-free season, and black dots show the individual measurements at Highway 130 Lake. The inset plot shows the modern annual ratio of snow/rain for the SNOTEL stations nearest Highway 130 Lake (black line) and Bison Lake (purple line) and the dashed lines show the means.

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Preprint. Discussion started: 19 November 2021

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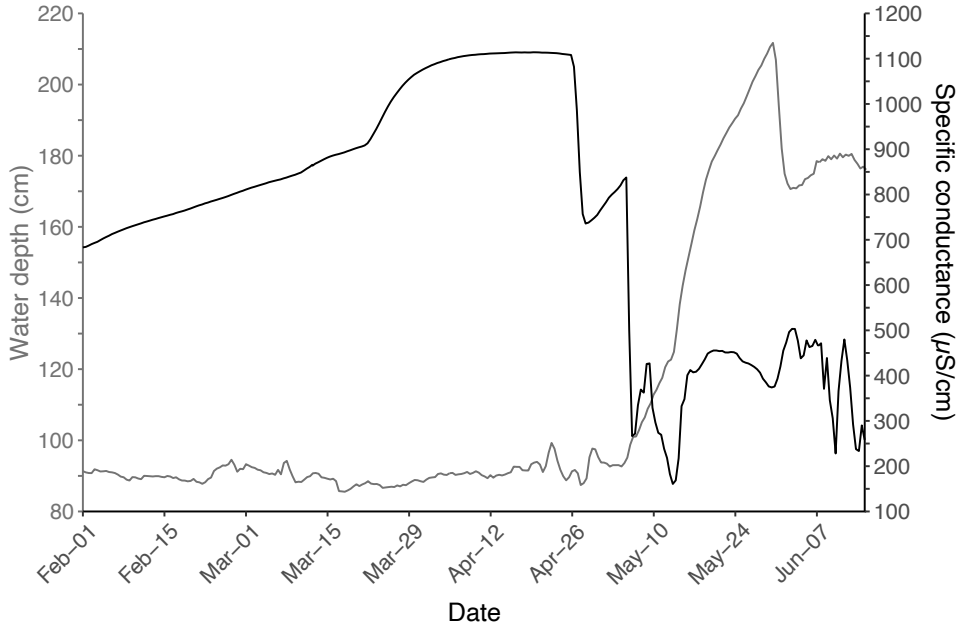

Figure 3. Measurements of water depth (gray line) and specific conductance (black line) at Highway 130 Lake in 2017.

234

235

236 **4.2 *Sediment characteristics***

237 The 333-cm core from HL extends to at least the early Holocene and contains
238 predominantly carbonate sediment underlain by silicate clays (Fig. 4). The upper 303 cm
239 contains from 5–55% in organics and 5–90% in carbonate; the core above the basal 30 cm has a
240 mean carbonate content of 65%. In the basal unit, the carbonate content drops below 5%, which
241 was too low for isotopic analysis. The age-depth model (black line with 2-sigma gray uncertainty
242 band, Fig. 4, Table 1) reveals average net sediment accumulation rates of 18 cm/kyr (thousand
243 years) from 11.7–4.4 ka and 45.5 cm/kyr from 4.4 ka to present. High rates of net sedimentation

https://doi.org/10.5194/cp-2021-149
Preprint. Discussion started: 19 November 2021

correspond with intervals of high carbonate flux into the lake, indicating that carbonate production may largely control sedimentation rates. The carbonate content and carbonate flux, representing the mass of carbonates deposited per unit area per year, increased simultaneously with the sedimentation rate at 4.4 ka (Fig. 4), but the percent carbonate content subsequently declined until 4.0 ka. The radiocarbon age at 119-cm depth (3.072 ± 0.03 ka) has an age similar to the date at 67-cm depth (3.031 ± 0.02 ka), which may indicate a reworked upper age (black dots in Fig. 4). However, high total sediment and carbonate accumulation rates are inferred even if the upper age was excluded from the age-depth model.

Table 1. Calibrated radiocarbon ages used for the age-depth model.

Lake	Core	Depth (cm)	Material	Lab number	Age (^{14}C yr BP)	Uncertainty (1 σ,^{14}C yr BP)	Calibrated age ranges (1 σ, cal yr BP)		
							Median	Maximum	Minimum
Highway 130 Lake	2A	18	Charcoal	UCIAMS-194167	850	30	748	783	726
		67	Charcoal	UCIAMS-194168	2,900	20	3,033	3,070	2,996
	2B	119-121	Charcoal	UCIAMS-194169	2,925	15	3,073	3,144	3,004
		154-156	Charcoal	UCIAMS-194170	3,660	35	3,986	4,081	3,921
		193-195	Charcoal	UCIAMS-194171	3,840	20	4,241	4,290	4,157
		204	Charcoal	UCIAMS-194172	3,965	20	4,438	4,508	4,412
		239	Charcoal	UCIAMS-194173	6,210	60	7,096	7,132	7,007
		302	Charcoal	UCIAMS-194174	9,580	25	10,927	11,074	10,781

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Preprint. Discussion started: 19 November 2021

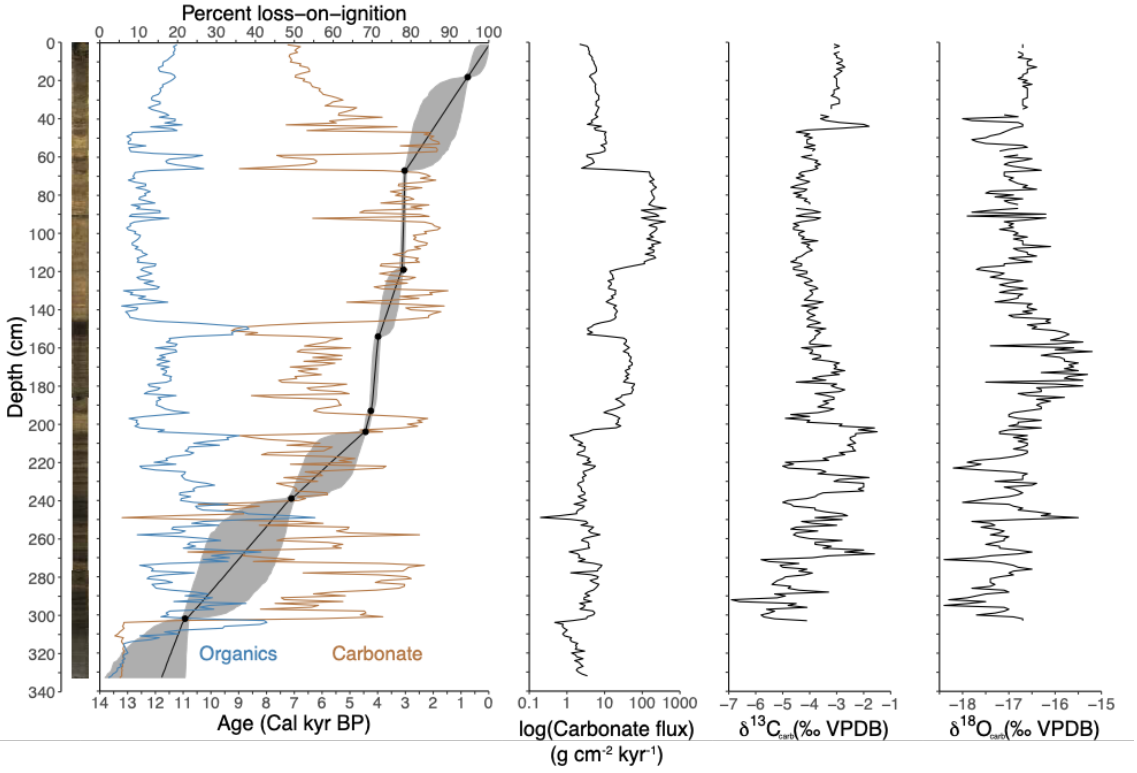

Figure 4. Percent organics, percent carbonate, carbonate flux, $\delta^{13}C_{carb}$, and $\delta^{18}O_{carb}$ are shown by depth alongside an image of the 333-cm-long sediment core from Highway 130 Lake. Radiocarbon ages (black dots) were used to create the age-depth model and gray uncertainty band (2 sigma).

4.3 Sedimentary oxygen and carbon isotopes

$\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ in the upper 303 cm of sediment range from -6.9 to -1.5‰ and -18.4 to -15.2‰, respectively, and the mean isotope compositions become more positive over the record, but the long-term trend of $\delta^{18}O_{carb}$ is not statistically significant (Fig. 4). Variance in $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ values is highest before 4.4 ka (below 200-cm depth) and lowest since 1.5 ka

https://doi.org/10.5194/cp-2021-149
Preprint. Discussion started: 19 November 2021

Climate of the Past Discussions
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262 (above 40-cm depth; Fig. 5). Isotope excursions appear in both the slow and fast sedimentation
263 intervals and when the carbonate flux is both low and high (Fig. 4).

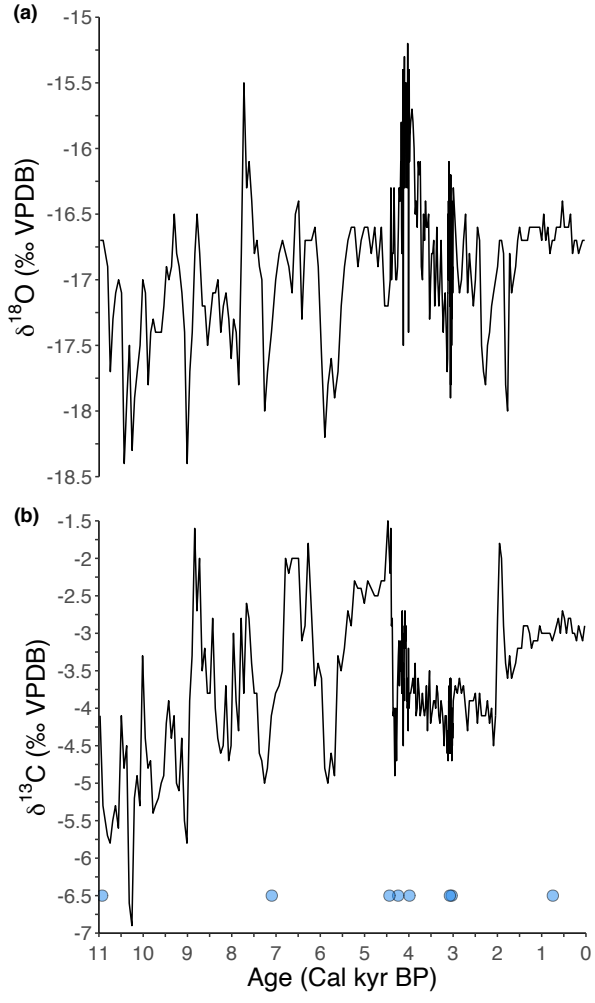

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Figure 5. $\delta^{18}O_{carb}$ (a) and $\delta^{13}C_{carb}$ (b) from Highway 130 Lake. The blue dots indicate the calibrated radiocarbon ages used for the age-depth model (refer to Table 1 for calibrated age uncertainties).

265 $\delta^{18}O_{carb}$ peaks from approximately 4.2–4 ka, where four calibrated radiocarbon ages
266 constrain the timing and indicate a fast sedimentation rate (Fig. 5). The carbonate flux is high,

https://doi.org/10.5194/cp-2021-149
Preprint. Discussion started: 19 November 2021


Climate of the Past Discussions

but the carbonate content is low (~55%) during this interval relative to the mean (Fig. 4). A positive excursion of similar magnitude also occurred from 7.8–7.3 ka, but aligns with high organic content and low $\delta^{13}C_{carb}$, carbonate content, carbonate flux, and total net sediment accumulation.

Compared to the records for Bison and Yellow Lakes in Colorado (Anderson, 2011, 2012; Fig. 1), $\delta^{18}O_{carb}$ values of HL are several per mil lower with higher variance for most of the Holocene (Fig. 6). This pattern changes in the late Holocene as carbonate in Bison Lake becomes isotopically lighter than before and approaches the oxygen isotope composition of HL, which maintains a relatively constant mean $\delta^{18}O_{carb}$ value. After approximately 1.5 ka, $\delta^{18}O_{carb}$ variability in HL drops to near the analytical uncertainty (± 0.2‰) while the other records show increased variably (Fig. 6).

5. Discussion

5.1 Evidence of the 4.2 ka drought in the southern Rocky Mountains

Peak $\delta^{18}O_{carb}$ in HL indicates an abrupt decline in effective moisture or at least a decline in the ratio of snowfall to rain in the Medicine Bow Mountains from approximately 4.2–4 ka (Fig. 5) when evidence from additional climate records shows that aridity affected the southern Rocky Mountains and portions of the Great Plains (Carter et al., 2013; Halfen & Johnson, 2013; Stokes & Gaylord, 1993). The highest $\delta^{18}O_{carb}$ values at HL coincide with the pollen-inferred precipitation and temperature changes at 4.2 ka at Long Lake, which records two centuries of severe drought (Long Lake, Fig. 1; Carter et al., 2013). The excursion also aligns with the

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Preprint. Discussion started: 19 November 2021

Climate of the Past Discussions
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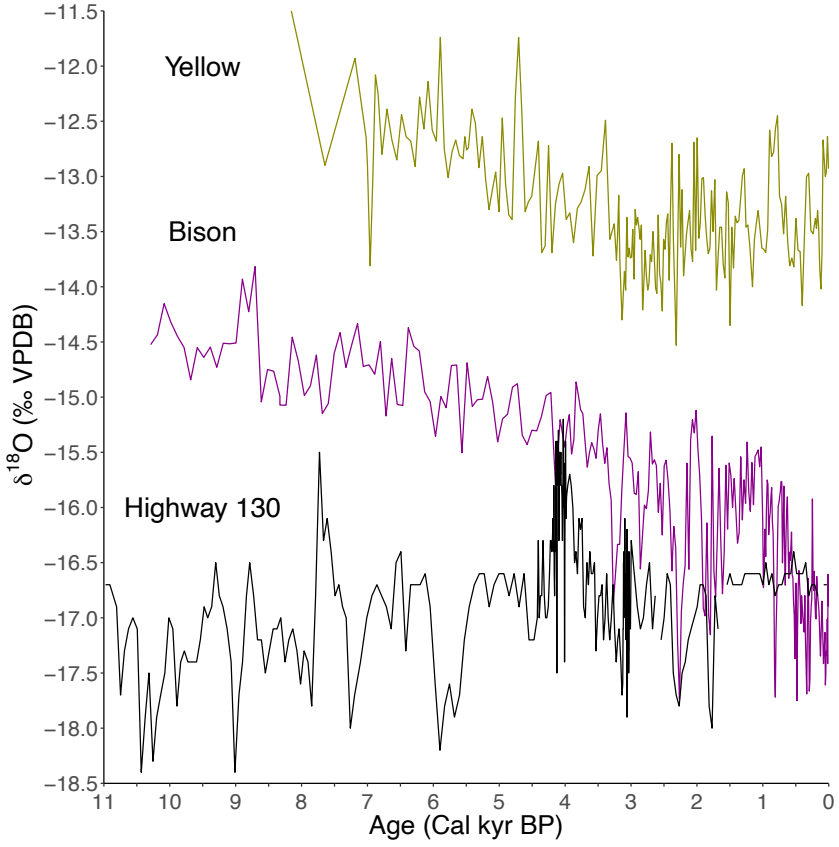


Figure 6. $\delta^{18}O_{carb}$ records from Highway 130 Lake (black), Bison Lake (purple; Anderson, 2011), and Yellow Lake (yellow; Anderson, 2012) vary despite their similar locations and elevations.

288 longstanding evidence of drought in the Great Plains and eastern southern Rocky Mountains
289 (dune fields, Fig. 1), where a rapid loss of grain-trapping vegetation likely triggered several
290 centuries of increased aeolian transport documented across multiple dune fields (Booth et al.,
291 2005; Forman et al., 2001; Halfen et al., 2010; Stokes & Gaylord, 1993).

292 Taken together, the records suggest that rapid drying at around 4.2 ka was an important
293 climatic event in the Medicine Bow Mountains even if the drought is not a prominent feature in
294 other paleoclimate studies from the mid-latitude Rocky Mountains (Anderson et al., 2008;

https://doi.org/10.5194/cp-2021-149
Preprint. Discussion started: 19 November 2021

295 Brunelle et al., 2013; Feiler et al., 1997; Johnson et al., 2013; Mensing et al., 2012; Minckley et
296 al., 2012; Shuman et al., 2010; Thompson et al., 1993; Whitlock & Bartlein, 1993), including the
297 nearby $\delta^{18}O_{carb}$ records from Bison and Yellow Lakes (Anderson, 2011, 2012). The spatial
298 patterns of late-Holocene hydroclimate changes in North America may have been complex
299 compared to other regions, such as the European continent where late-Holocene climate
300 variability appears more coherently in climate records (e.g., Deininger et al., 2017). Still, the
301 inconsistent evidence complicates interpretations of the 4.2 ka anomaly here and elsewhere
302 (Bradley & Bakke, 2019).
303 Some paleohydrologic evidence indicates, however, that the event may have been
304 extensive in the southern Rocky Mountains. Sedimentological changes in high-elevation lakes in
305 Colorado show substantial hydrological transformation at around 4 ka matching the timing and
306 scale of drought inferred from HL's record (Fig. 1 & 7; Shuman et al., 2009a, 2014, 2015). The
307 sediment stratigraphies record low water levels that shifted shoreline sands to the locations of
308 cores collected in 1–5-m water depth today and thus indicate reduced effective moisture at 4.2–
309 3.9 ka (gray shaded regions, Fig. 7). The sand layers coincide in time with the elevated carbonate
310 accumulation rate and $\delta^{18}O_{carb}$ values at HL and with dune activity in southeast Wyoming
311 (Halfen & Johnson, 2013; Stokes & Gaylord, 1993); the high-elevation lake locations and
312 geophysical site surveys confirm that the shallow-water sands were not deposited by aeolian
313 activity. A second prominent sand layer in Emerald and Upper Big Creek Lakes at ca. 3.1 ka
314 (gray shaded regions, Fig. 7) indicates low effective moisture and overlaps with the maximum
315 rate of carbonate accumulation at HL, but a second sand layer does not appear in Little Molas
316 Lake and the $\delta^{18}O_{carb}$ values in HL are lower than at 4.2 ka. It took several centuries for $\delta^{18}O_{carb}$
317 values to rise and fall before and after the peak from 4.2–4 ka, but the excursion at 3.1 ka

https://doi.org/10.5194/cp-2021-149
Preprint. Discussion started: 19 November 2021

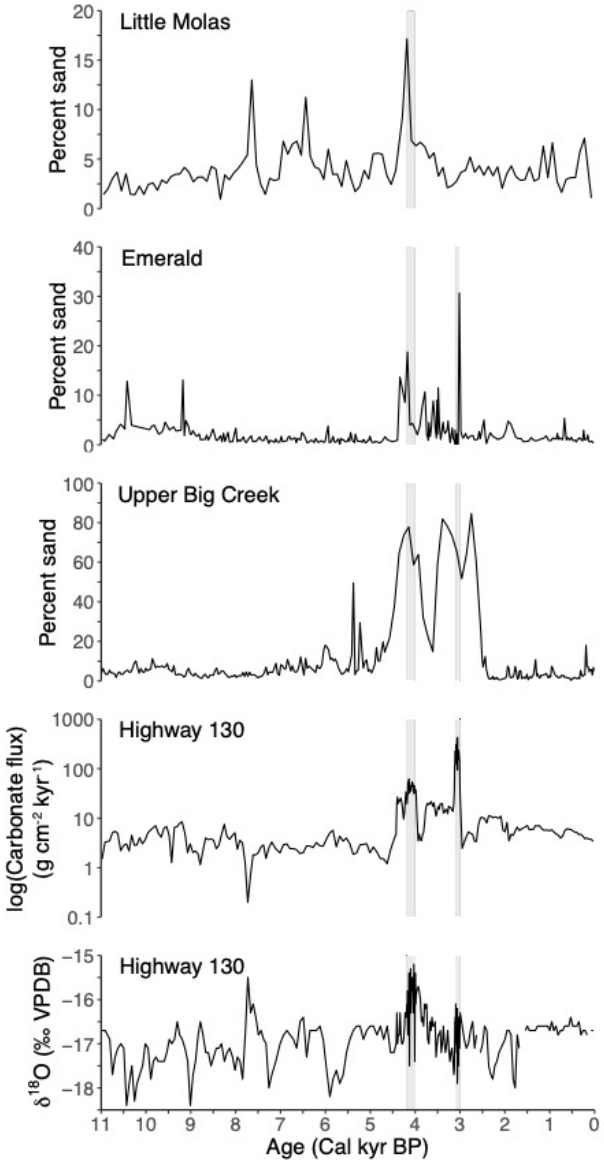

Climate of the Past Discussions

318 occurred within a century. Multiple radiocarbon ages constrain the interval of high carbonate
319 accumulation from approximately 4.4–3 ka, but the sedimentation rate in the interval is sensitive
320 to removal of one of the ages; if the age at ca. 3 ka is out of sequence, we could bias the peak
321 rate of sediment and carbonate accumulation toward high values.

322 The rapid transition from deep-water muds to shallow-water sands as water levels
323 dropped in the Colorado lakes at around 4.2 ka corresponds with changes in pollen assemblages
324 in central Colorado (Jiménez-Moreno et al., 2019) and southeast Wyoming (Carter et al., 2013),
325 as well as with other evidence for drought in North America (Booth et al., 2004). Similar
326 sedimentological features found in lakes along the Atlantic margin from Maine to Pennsylvania
327 date to around 4.2 ka, for example, where the drought appears as one of multiple events linked to
328 circulation changes over the North Atlantic (Li et al., 2007; Marsicek et al., 2013; Newby et al.,
329 2014; Nolan, 2020; Shuman et al., 2019; Shuman & Burrell, 2017). The sequences in the
330 southern Rocky Mountains, however, include uniquely prominent isotopic and sedimentological
331 changes from 4.2–3.9 ka.

332 Given the growing evidence of drought within the southern Rocky Mountains associated
333 with the widespread climatic anomaly at 4.2 ka, a lack of $\delta^{18}O_{carb}$ records of the event in the
334 region, or in North America entirely, is surprising (Anderson et al., 2016b; Konecky et al., 2020).
335 However, individual sites respond to a varying mixture of local and regional factors. The
336 stratigraphic evidence of lake-level change in the region is not entirely consistent either and may
337 indicate interactions with different directions of hydroclimate change across seasons, elevations,
338 and latitudes. Stratigraphic features in Hidden Lake, located in northern Colorado just south of
339 Upper Big Creek Lake (Fig. 1) but several hundred meters lower in elevation, document a rapid
340 increase in effective moisture at around 4 ka (Shuman et al., 2009)—the opposite response of the

https://doi.org/10.5194/cp-2021-149
Preprint. Discussion started: 19 November 2021



341

Figure 7. Spikes in the sand content of Little Molas, Emerald, and Upper Big Creek Lakes, located in high-elevation watersheds in Colorado (Fig. 1), align with the positive $\delta^{18}O_{carb}$ excursion at Highway 130 Lake and indicate low water from approximately 4.2–4 ka (gray shaded areas) resulting from low effective moisture (Shuman et al., 2009a, 2014, 2015). Another positive $\delta^{18}O_{carb}$ excursion at ca. 3.1 ka (gray shaded areas) aligns with intervals of low water at Emerald and Upper Big Creek Lakes.

https://doi.org/10.5194/cp-2021-149
Preprint. Discussion started: 19 November 2021

surrounding lakes at higher elevations (Fig. 7). The wet phase was abrupt in onset and termination and lasted from around 4.4–3.7 ka, which stands out in the otherwise gradual trend towards higher water levels since 6 ka without any major intervals of low water in the late Holocene (Shuman et al., 2009).

The low-elevation location of Hidden Lake may indicate an important role for increased summer or fall rainfall when high-elevation sites declined in response to low winter snowfall. The combined effects could have favored the unusually high $\delta^{18}O_{carb}$ at HL. Low winter snow can create favorable surface-energy conditions for strong summer convective precipitation (Zhu et al. 2005). Alternatively, the reversed hydrologic response of Hidden Lake could indicate antiphased hydroclimate changes in the southern Rocky Mountains between high and low elevations, which is consistent with modern responses to the El Niño-Southern Oscillation (Preece et al., 2020). The active dune fields in east-central Wyoming, however, confound a simple interpretation of the elevational and seasonally antiphased hydrologic changes. Latitudinal hydroclimate variability could be an additional complicating factor and has previously been described across the area due to transient climatic boundaries with the northern and southern Rocky Mountains (Shinker, 2010; Wise, 2010). The comparison of the radiocarbon age uncertainties of the 4.2 ka paleoshoreline sands at lake-level sites, including Emerald Lake in central Colorado (Fig. 1 & 7), indicates a late-Holocene north-south moisture dipole extending across much of the area described here (Shuman et al., 2014).

Given the potential prominence of the 4.2 ka drought at HL and other southern Rocky Mountain records, it may have been uniquely severe in this region even if it had a complex regional expression at broader spatial scales. The lake-level reconstructions from Colorado contain evidence of other Holocene hydrologic changes (Fig. 7) and HL shows another positive

https://doi.org/10.5194/cp-2021-149
Preprint. Discussion started: 19 November 2021


Climate of the Past Discussions

365 excursion at 7.8 ka (Fig. 5), but the records lack evidence for multiple recurrent, multi-century
366 hydroclimate changes recorded with the 4.2 ka event in places like the Atlantic margin (Shuman
367 et al., 2019). Elsewhere, aridity at 4.2 ka may represent just one of several repeated drying events
368 consistent with climate records and simulations from around the world that show drought as a
369 regular feature of late-Holocene climate variability (Arz et al., 2006; Bradley & Bakke, 2019;
370 Mayewski et al., 2004; Wanner et al., 2008; Wanner et al., 2015; Yan & Liu, 2019). The mid-
371 latitude Rocky Mountain records may suggest that the midcontinent was insulated from some of
372 the abrupt late-Holocene climate changes, possibly due to its isolation from the ocean-
373 atmosphere dynamics proposed to play key roles in Holocene variability (Arz et al., 2006;
374 Deininger et al., 2017; Jalali et al., 2019; Yan & Liu, 2019).
375
376 **5.2 *Varying $\delta^{18}O_{carb}$ trends in the southern Rocky Mountains***
377 The marked sensitivity of lake-water $\delta^{18}O$ to hydroclimate changes may make lacustrine
378 carbonates ideal indicators of past droughts like the 4.2 ka event, as documented by $\delta^{18}O_{carb}$
379 records outside of North America (e.g., Bini et al., 2019; Dean et al., 2015) and by our record at
380 HL (Fig. 5), but site-specific hydrologic conditions could complicate the signals. They may
381 generate inconsistent trends among records over both short (seasonal) and long (millennial)
382 timescales (Gibson et al., 2016; Mark D. Shapley et al., 2008; Steinman & Abbott, 2013; Tyler et
383 al., 2007). Indeed, we observe such inconsistency in the southern Rocky Mountains (Fig. 6).
384 The hydrologic controls, such as groundwater fluxes and basin morphology, can vary
385 based on a lake's geohydrological setting (Anderson et al., 2016; Dean et al., 2015). Modern
386 lake-water hydrogen and oxygen isotope measurements reveal stronger fractionation by
387 evaporation in HL (thick black line, Fig. 2) compared to Bison and Yellow Lakes (purple and

https://doi.org/10.5194/cp-2021-149
Preprint. Discussion started: 19 November 2021


Climate of the Past Discussions

388 yellow lines, Fig. 2), which exhibit a narrower range in modern water isotope values and smaller
389 deviation from the global meteoric water line (Anderson, 2012). However, the carbonate at HL is
390 isotopically lighter than the other two (Fig. 6), which is antithetical to the expectation based on
391 evaporatively enriched summer waters (Fig. 2). The pattern differs from the interpretation that
392 the Bison Lake $\delta^{18}O_{carb}$ was not strongly influenced by evaporation because it was isotopically
393 lighter than other sites like Yellow Lake (Anderson, 2011, 2012). HL lacks the prominent
394 $\delta^{18}O_{carb}$ trend observed at these other sites (Fig. 6). Given the modern water isotope values, we
395 had anticipated that $\delta^{18}O_{carb}$ from HL would be isotopically heavy compared to Bison and
396 Yellow Lakes, but track similar trends (Anderson, 2012).
397 Because the modern water isotope values poorly predicted $\delta^{18}O_{carb}$ trends, the different
398 lakes may record past changes in different ways. HL may be a better indicator of winter
399 snowpack than evaporation. $\delta^{18}O_{carb}$ values were below the mean at 6 ka (Fig. 5) when simulated
400 estimates of evaporation rates in the Medicine Bow Mountains were up to 30% higher than today
401 (Morrill et al., 2019), which would indicate that such enhanced summer evaporation did not
402 affect $\delta^{18}O_{carb}$ at HL. We also find lower-than-expected $\delta^{18}O_{carb}$ values in the uppermost
403 sediments. Despite an increase in summer lake-water $\delta^{18}O$ from -17.8 to -10.8‰ today (thick
404 black line, Fig. 2), $\delta^{18}O_{carb}$ values since 1.5 ka only reached a maximum of -16.4‰ and the core-
405 top value is -16.7‰ (Fig. 5), which is closer to the composition of groundwater (open squares,
406 Fig. 2) than the mid- to late-summer lake-water-$\delta^{18}O$ values (black circles, Fig. 2).
407 Previous studies have shown that the deposition of endogenic carbonate occurs
408 predominantly in the warm summer months when photosynthesis optimizes carbonate production
409 by modifying dissolved CO_2 concentrations and pH (Leng & Marshall, 2004), but the
410 isotopically light carbonate at HL may contradict this expectation. For comparison, the

https://doi.org/10.5194/cp-2021-149
Preprint. Discussion started: 19 November 2021


Climate of the Past Discussions

411 uppermost $\delta^{18}O_{carb}$ value of -14.9‰ in Bison Lake (purple line, Fig. 6) falls within the range in
412 modern summer lake-water-$\delta^{18}O$ values of -16.7 to -13.5‰ (purple line, Fig. 2; Anderson,
413 2011). $\delta^{18}O_{carb}$ in HL, therefore, may not integrate the range of summer lake-water $\delta^{18}O$ as is
414 assumed for Bison and Yellow Lakes (and carbonate lakes in general), but early springtime
415 deposition of carbonate at HL could capture the signature of isotopically light lake water without
416 modification by warm-season evaporation.
417 The year-round measurements of specific conductance show that conditions favorable for
418 carbonate precipitation may indeed be highest during late winter and spring. In 2017, specific
419 conductance of the water below the surface ice rose from 700 µS/cm in early February to 1,115
420 µS/cm by early April, and it remained above 1,000 µS/cm throughout April (Fig. 3). These high
421 values would favor carbonate precipitation, whereas the summertime waters are more dilute.
422 Specific conductance of HL and other lakes within the watershed during the summer typically
423 does not exceed 300 µS/cm. Melting of lake ice and snowpack rapidly lowers the specific
424 conductance by early May and it remains between 250–300 µS/cm for the remaining ice-free
425 months. The conductance likely remains lower than in winter despite evaporative enrichment of
426 the oxygen isotopes because of groundwater discharge into the lake (Rautio & Korkka-Niemi,
427 2011), which geophysical surveys, water temperatures, and stable summer water levels at HL
428 support (Liefert et al., 2018). If so, ions exsolved from overlying ice raise the conductance of the
429 lake water and pore water within the bottom sediments beyond the concentration in groundwater
430 in winter (Adams & Lasenby, 1985), and create favorable conditions for the rapid deposition of
431 endogenic carbonate in early spring when the isotopic signal would not reflect evaporation or
432 isotopically heavy summer rainfall (open circles, Fig. 2). Spring carbonate formation could also
433 yield a different temperature-dependent effect on the $\delta^{18}O_{carb}$ in HL compared to the other lakes,

https://doi.org/10.5194/cp-2021-149
Preprint. Discussion started: 19 November 2021


Climate of the Past Discussions

but the cold spring waters at HL should favor an increase, not decrease, in $\delta^{18}O_{carb.}$ Indeed, all of the readily expected process that could complicate a carbonate isotopic record should drive the $\delta^{18}O_{carb}$ in the positive (not negative) direction and underscore the significance of the difference between HL and the other lakes.

As an alternative explanation, the $\delta^{18}O_{carb}$ values could reflect changes in total precipitation rather than seasonality effects because the inflow of Ca-bearing groundwater (which should rise with precipitation) can increase carbonate production and lower $\delta^{18}O_{carb}$ values in alkaline lakes in both ice-free and ice-covered conditions (Shapley et al., 2005), but the weak covariance of weight percent carbonate and $\delta^{18}O_{carb}$ suggest that rates of groundwater inflow did not strongly influence $\delta^{18}O_{carb}$ (Fig. S1a). A weak covariance of $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ indicates short lake-water residence times throughout the lake's history (Fig. S1b; Drummond et al., 1995; Talbot & Kelts, 1990), which could be consistent with rapid flowthrough that reduced evaporative enrichment; removing values from 4.2-4 ka only marginally improves the correlation.

A strong negative correlation of weight percent organics and carbonate ($R^2 = –0.79$) suggests that carbonate abundance depends primarily on biological productivity that promoted carbonate dissolution by releasing CO_2 and lowering pH (Fig S1c; Dean, 1999). Carbonate content from 4.2–4 ka was below the mean despite low organic content (red dots, Fig. S1c) and a high flux of carbonate (Fig. 4), which may represent a shift in HL's water levels and chemistry that favored both acidic conditions and isotopically heavy carbonate (red dots, Fig. S1). Down-core shifts in $\delta^{18}O_{carb}$ produced by seasonal changes in the timing and rate of carbonate formation have been proposed as potential sources of variability within individual records (Fronval et al., 1995; Lamb et al., 2007; Steinman et al., 2012; Steinman & Abbott, 2013; Tyler et al., 2007) and

https://doi.org/10.5194/cp-2021-149
Preprint. Discussion started: 19 November 2021

457 could play a role in the record at HL, but such differences could also generate the variability in
458 the long-term trends observed among records from the southern Rocky Mountains and elsewhere
459 (Bini et al., 2019; Konecky et al., 2020; Roberts et al., 2008).
460 Other factors that could affect $\delta^{18}O_{carb}$, such as precipitation patterns and biological
461 disequilibrium effects, are unlikely sources of variability among the regional $\delta^{18}O_{carb}$ records.
462 The seasonal balance of precipitation today is broadly similar among the sites (inset plot, Fig. 2)
463 and the calculated annual precipitation-$\delta^{18}O$ value is approximately 1‰ lower at HL
464 (http://waterisotopesDB.org). Annual temperature ranges are also similar for the watersheds,
465 making it unlikely that temperature dependence of fractionation could explain the range in
466 $\delta^{18}O_{carb}$ values recorded across the three records unless the different water depths and
467 groundwater influences altered the seasonal temperature progression among lakes. The
468 difference in temperature would need to be large (~12 °C) to explain the offset in $\delta^{18}O_{carb}$
469 between HL and Bison Lake (and larger for the offset between HL and Yellow Lake), which is
470 unrealistic given the sites' comparable locations and elevations and the relatively small
471 temperature changes at mid-latitudes since 11 ka (Marsicek et al., 2018). We also find no
472 evidence in the sediment core or modern lake setting to indicate that biologically mediated
473 precipitation of calcite substantially altered $\delta^{18}O_{carb}$ at HL (e.g., by the accumulation of ostracod
474 tests that precipitate carbonates in disequilibrium with lake water). Disequilibrium effects
475 associated with biogenic carbonates generally increase $\delta^{18}O_{carb}$ (Holmes & Chivas, 2002; Leng &
476 Marshall, 2004), which would be difficult to reconcile with the surprisingly negative mean and
477 core-top $\delta^{18}O_{carb}$ values at HL. Down-core carbonate phase changes are also unlikely as we
478 identified that only calcite was present using x-ray diffraction (XRD).
479

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6. Conclusions

$\delta^{18}O_{carb}$ from HL indicates an abrupt hydroclimate change in the southern Rocky Mountains from approximately 4.2–4 ka that reduced effective moisture or caused less snow to fall than today at high elevations in southern Wyoming. Other $\delta^{18}O_{carb}$ records from the region do not document the drought (Fig. 6; Anderson, 2012), but the event's timing overlaps with evidence of multi-century drought from pollen, lake stratigraphies, and dunes in the southern Rocky Mountains (Carter et al., 2013; Halfen & Johnson, 2013; Shuman et al., 2009a, 2014, 2015; Stokes & Gaylord, 1993), the western Great Plains (Booth et al., 2005; Dean, 1997; Halfen & Johnson, 2013; Mason et al., 1997; Stokes & Gaylord, 1993), and elsewhere around the world (Nakamura et al., 2016; Di Rita & Magri, 2019; Scuderi et al., 2019; Xiao et al., 2018).

The timing and magnitude of hydroclimate change in our record agrees with the perspective of a widespread megadrought at around 4.2 ka (Weiss, 2016), but inconsistencies among climate records suggests that (1) site-specific factors can prevent identification of the patterns of abrupt hydroclimate changes, particularly in $\delta^{18}O_{carb}$ records; (2) the hydrologic response in North America and likely elsewhere around the world was spatially complex; and (3) the abrupt hydroclimate changes in the North American midcontinent were more pronounced against background Holocene variability than in many regions such as the Atlantic margin. Consequently, a prolonged 'megadrought' at 4.2 ka was likely a significant feature of the hydroclimate history in the mid-latitude Rocky Mountains even if that is not true globally.

Data availability

Data related to this paper will be made available through the National Centers for Environmental Information on the National Oceanic and Atmospheric Administration website:

503 https://www.ncdc.noaa.gov/data-access/paleoclimatology-data. The analyses were performed in
504 R.

505

506 **Author contributions**

507 D. Liefert and B. Shuman contributed to the design and implementation of the research, to the
508 analysis of the results, and to the writing of the manuscript.

509

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517

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