1   **Stromatolites in Walker Lake (Nevada, Great Basin, USA) record climate and lake**
2   **level changes ~35,000 years ago**

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17

18   **Abstract**

19   Walker Lake is a closed-basin remnant of the large Pleistocene glacial Lake Lahontan
20   system that has experienced multiple high amplitude (100-200 m) changes in water level
21   over the past ~40,000 years in response to changes in climate. A laminated carbonate
22   stromatolite composed of varying proportions of calcite fans and micrite was collected
23   from a paleoshoreline located at approximately 58 meters above present lake level.
24   Radiocarbon dating revealed that the stromatolite spans approximately 2,000 years of
25   growth, from 35,227 to 33,727 calibrated years before present (YBP), a time period
26   during which paleolake-level is not well-constrained. Distinct laminae were drilled along
27   the growth axis, and the resulting powders were collected for clumped isotope analyses to
28   generate formation temperatures (lake water temperatures) during stromatolite formation
29   from which $\delta^{18}\text{O}_{\text{water}}$ was calculated.

30   Results indicate that the stromatolite experienced an initial increase in temperature and
31   water $\delta^{18}\text{O}$ values followed by a decrease in both during the course of accretion. The
32   resulting temperature and isotopic data were input into a Rayleigh distillation model for
33   lakewater evaporation in order to estimate the magnitude of lake level and volume
34   fluctuations over the course of accretion. Modeling results reveal a lake level decrease of
35   between 8.1 and 15.6 m, followed by an increase of between 4.3 and 8.8 m during the
36   course of stromatolite growth.

37   The results of this study indicate that Walker Lake experienced significant lake volume
38   change over the course of 2000 years, perhaps as a response to precipitation changes

39 driven by fluctuations in the polar jet stream and accompanying changes in regional
40 climate, and/or evaporation-induced changes in lake level. These results add to a growing
41 body of research indicating that stromatolites and other lacustrine tufas represent a
42 detailed and extensive terrestrial archive that can potentially be used to reconstruct the
43 timing and magnitude of climate change.

44 **Keywords: Stromatolite, Clumped Isotope, Walker Lake, Polar Jet Stream,**
45 **Pleistocene**

46

47   **Introduction**

48   Accurate terrestrial reconstructions of paleotemperature and paleohydrology are
49   critical for understanding surface environments throughout Earth history. Terrestrial
50   paleoclimate data can aid in building reliable climate models, and can help us understand
51   the links between continental and oceanic climate data (e.g. Rowley and Garzione, 2007;
52   Hren and Sheldon, 2012). Such data are also relevant to constraining the sensitivity of the
53   Earth system to climate forcing (MARGO, 2009).

54   Various archives have been used to reconstruct climate parameters in terrestrial
55   settings, (e.g. leaf margin analysis, palynology, glacial deposit distributions, and stable
56   isotopes of lacustrine deposits). In many cases, terrestrial signals yield qualitative, but not
57   high-resolution quantitative results because of the non-continuous nature of such
58   archives. While overall trends (wetter/drier, cooler/warmer) may be discernable, absolute
59   values cannot be determined with a high degree of confidence. Speleothems can provide
60   insights into terrestrial climate, although in some cases the proxy data derived from
61   speleothems are subject to multiple influences (e.g., Affek et al., 2008; Daëron et al.,
62   2011) and therefore can be challenging to interpret (Johnson and Ingram, 2004; Dayem et
63   al., 2010). Lake sediment cores represent another useful terrestrial archive of climate
64   (Rosenthal and Broccoli, 2004), however not all hydrologic proxies are easy to interpret
65   (Yuan et al., 2006A), and sediment cores are limited to the geologically recent.

66   This study dates stromatolites from Walker Lake, Nevada, and applies carbonate
67   “clumped” isotope thermometry to reconstruct paleo-lake temperatures, shifts in lake

68 level and terrestrial paleoclimate that occurred ~30,000 years ago (30 ka), during which
69 the Lahontan lake level record is spotty and unresolved. Clumped isotope thermometry,
70 despite being a relatively young proxy, has an extensive record of use in lacustrine
71 systems (e.g, Huntington et al., 2010; Quade et al., 2011; Hren and Sheldon, 2012;
72 Lechler et al., 2014; Frantz et al., 2014; Petryshyn et al., 2015), discussed below.

73 *Traditional Paleothermometry vs. Clumped Isotope Paleothermometry*

74 Changes in $\delta^{18}\text{O}$ ratios in lacustrine carbonates can be a useful tool for the study of
75 temperature, vapor sources, and changes in evaporation/precipitation as they pertain to
76 lake level fluctuations (e.g. Benson et al., 1996, Solari et al., 2010). The traditional, most
77 widely-used method of determining the temperature of formation of a carbonate mineral,
78 the oxygen isotope paleothermometer, requires knowledge of both the oxygen isotope
79 ratio of the fluid of formation and the oxygen isotope ratio of the carbonate (e.g. Urey,
80 1947; Epstein et al., 1953, Emiliani, 1966). The fractionation between these two phases is
81 temperature-dependent (Kim and O'Neil, 1997). Unfortunately, the isotopic composition
82 of the fluid from which the carbonate precipitated is nearly always unknown and must be
83 estimated/guessed to solve the temperature equation with varying degrees of confidence.
84 Recent work suggests this fractionation can also be affected by pH and growth rate
85 (Watkins et al., 2014). All of these factors can further complicate the interpretation of
86 oxygen isotope ratios in carbonate minerals and the paleotemperatures derived from
87 them. Here, we apply the clumped isotope paleothermometer to lacustrine stromatolites
88 in order to surmount some of the drawbacks of traditional $\delta^{18}\text{O}$ paleothermometry. With

89 the temperature of formation and the isotopic composition of the mineral known, the $\delta^{18}\text{O}$
90 of ancient lake water can be uniquely calculated.

91 *Stromatolites as terrestrial climate archives*

92 Several studies have explored the use of lacustrine carbonates, and stromatolites in
93 particular, to reconstruct past environment and climate (e.g. Petryshyn et al., 2012; Ibarra
94 et al., 2013; Frantz et al., 2014; Petryshyn et al., 2015). Stromatolites, laminated
95 sedimentary structures that are accretionary away from a point or surface (Semikhatov et
96 al., 1979), typically accrete through the interplay of sediment, environment, and biology
97 (e.g., Shapiro, 2007; Riding, 2011; Bosak et al., 2013, Frantz et al., 2015). As the
98 structures form, they record chemical and physical information about their surroundings.
99 Frantz et al. (2014), coupling trace element analysis, clumped isotope paleothermometry,
100 and oxygen isotope modeling of stromatolites from the Eocene Green River Formation,
101 was able to document several meter-scale changes in paleolake level, corresponding to
102 several tens of kilometers of shoreline change over 30 cm of stromatolite accretion. This
103 study demonstrated that stromatolites record the physical changes in their surroundings as
104 lake level, water temperature, and environment fluctuates around them.

105 Petryshyn et al. (2015) also utilized clumped isotope thermometry while studying
106 the carbonate tufas (microbialites) of Pavilion Lake, British Columbia. The stromatolites
107 in this study, although thought to be mostly modern, were actually found to be up to 9 ka.
108 The stromatolites also contained evidence of a historic drought in the region prior to 7 ka,
109 followed by a period of lake level rise. These studies (and several others mentioned

110 above) have demonstrated the utility of lacustrine stromatolites as a high-resolution
111 climate archive.

112 **Walker Lake**

113 Walker Lake is a closed-basin, sodium bicarbonate lake with an area of $\sim 100 \text{ km}^2$
114 and a maximum depth of 28 m in Summer 2015 (Figure 1A). The lake was once part of
115 the Pleistocene Lake Lahontan complex (e.g., Russel, 1885), and has been isolated from
116 the remaining Lake Lahontan system for the last 12,000 years (Benson and Thompson,
117 1987; Benson et al., 1995). Walker Lake has seen a historic drawdown (over 50 m) since
118 1882 due to anthropogenic diversion of the Walker River, its main tributary (Benson et
119 al., 1991; Yuan, et al., 2006B; Adams, 2007). Currently the lake surface is at an elevation
120 of 1198 m above sea level.

121 Presently in Walker Lake, calcium carbonate precipitation is limited by the
122 availability of dissolved calcium. While it is replete with carbonate ($\sim 20 \text{ mM}$; pH 9.5),
123 the overall concentration of calcium is less than 0.2 mM (Berelson et al., 2009).
124 Petryshyn et al. (2012) analyzed a number of Holocene carbonate stromatolites found
125 currently exposed on the southwest corner of the lake. The stromatolites were dated, and
126 laminations were counted to produce a record of lamination frequency. Surprisingly,
127 Petryshyn et al. (2012) found that lamination rate was not yearly, or even seasonal,
128 contrary to long-standing assumptions that stromatolite laminae represent biologically-
129 relevant diurnal, seasonal, or annual cycling. When lake level was high, laminae
130 precipitated on average every four to six years with ENSO-like periodicity. This larger-

131 scale climatic forcing was required to deliver enough calcium to the lake to produce a
132 single lamination. When lake level was low (near present day levels), yearly runoff was
133 able to deliver enough calcium to promote precipitation. These structures hint that high-
134 resolution terrestrial climate signals are captured by Walker Lake stromatolites.

135   In addition to the stromatolites studied by Petryshyn et al. (2012), several other
136 discrete horizons of carbonate accretions of varying ages are found above current lake
137 level (Newton and Grossman, 1988; Benson et al., 1996). The stromatolites studied here
138 are found 55-60 m above the eastern shoreline (38°38'30"N, 118°40'21"W). The
139 stromatolites crop out in large, meter-tall ‘heads’, with arms or ‘petals’ that protrude
140 laterally out from the sides (Figure 1B-C); that is, the structures grew upward and
141 outward simultaneously.

142 *Lake Level and Climate History*

143   The history of Pleistocene Lake Lahontan and its remnants, including Walker
144 Lake, has been documented and refined in many studies (e.g., Newton and Grossman,
145 1988, Benson et al., 1991; 1995 and 1996; Beutal et al., 2001; Yuan et al., 2006A; Yuan
146 et al., 2006B; Adams 2007; Berelson 2009; Benson et al., 2013; Reheis et al., 2014). The
147 majority of the carbonates (tufas) used to construct the record were deposited in the last
148 35 ka (Benson et al., 1995).

149   Generally, it is thought that Lake Lahontan was at a lowstand between 27,500 and
150 23,500 ybp (Benson et al., 1995; 2013). At this time, lake surface elevation was ~1156 m
151 above sea level. Walker Lake would have been a separate, closed basin at this time, given

152 the height of the sill that separates it from the rest of the Lahontan basin (Benson et al.,
153 1995; 2013). Prior to ~30 ka, there are few constraints on lake level, though recently, the
154 use of volcanic tephra (interpreted at the outcrop scale) has been implemented in
155 reconstruction of the Lake Lahontan record (Reheis et al., 2014). While the two proxies
156 (tufa and tephra) produce similar lake level curves from 22 ka to the present, they diverge
157 (often by tens of meters) at older dates (see Reheis et al., 2014, Figure 12). Thus, the
158 history of Lake Lahontan prior to 30 ka remains unresolved.

159 In this study, we aim to use the stromatolites found on the east shore of Walker
160 Lake to derive a lake level history during this disputed time period by coupling carbon
161 dating with clumped isotope thermometry and an oxygen isotope-based model of lake
162 volume.

163 **Methods**

164 Full stromatolite heads were collected from the east side of Walker Lake (Figure
165 1) at a paleoshoreline at 1288 m elevation, 58 m above the 2014 lake level. Stromatolites
166 were cut, polished, imaged, and thin sectioned. One lateral growth structure (termed a
167 ‘petal’; Figure 2A) was drilled along sequential laminae for ^{14}C dating. That same petal
168 was sampled along the growth transect (from the point of accretion to the top of the
169 structure) and along laminae for clumped isotope analyses. The point of petal inception is
170 termed the ‘proximal’ end of the petal, and the termination of growth is referred to as the
171 ‘distal’ end for the remainder of the manuscript. The drilled slabs were made into thin
172 sections with the drill holes preserved for careful petrographic analysis. Modern tufa

173 towers located ~2 m above the current shoreline on the southeast corner of the lake
174 (Figure 1A; S4) were also collected and treated in the same manner, for comparison.

175 *¹⁴C dates*

176 Powders were analyzed for radiocarbon age at the Keck Radiocarbon Facility at
177 the University of California, Irvine. Radiocarbon ages are calibrated using the IntCal13
178 database (for Northern Hemisphere terrestrial samples), and corrected for the reservoir
179 effect of carbon in the lake. The reservoir effect in Walker Lake has been estimated to be
180 300 years (Broecker and Walton, 1959; Benson et al., 1995), although this could have
181 been different during different time periods, given the fluctuating size of the lake.

182 *Clumped Isotope Analysis*

183 Clumped isotope thermometry is based on the measurement of the rare mass 47
184 isotopic species (isotopologue) of CO₂ produced during phosphoric acid digestion of
185 carbonate minerals. “Clumped” refers to the occurrence of bonds between two heavy
186 isotopes in a molecule (e.g., the bond between ¹⁸O and ¹³C in mass 47 ¹⁸O=¹³C=¹⁶O). In
187 carbonate minerals, the abundance of the mass 47 isotopologue is temperature-dependent
188 (Schauble et al., 2006). When precipitation proceeds at thermodynamic equilibrium,
189 lower temperatures favor the formation of mass 47 isotopologues, while progressively
190 higher temperatures produce “clumped” abundances that approach a stochastic (random)
191 value. These properties of clumped isotopes led to the development of the carbonate
192 clumped isotope paleothermometer, which is independent of the fluid from which the
193 carbonate precipitated, circumventing many of the problems associated with using δ¹⁸O

194 paleothermometry (e.g., Ghosh et al., 2006, 2007; Eiler, 2007; Huntington et al., 2009;
195 Tripathi et al., 2010; Thiagaraian et al., 2011; Eagle et al., 2013; Zaruur et al, 2013;
196 Cummins et al., 2014; Henkes et al., 2013; 2014; Tang et al., 2014; Frantz et al., 2014;
197 Petryshyn et al., 2015).

198   At least 40 mg of carbonate powder was taken from each dated section of the
199 stromatolite petal in order to ensure a sufficient amount of CO₂ for radiocarbon and
200 clumped isotope analysis in triplicate. Samples were measured on a specially modified
201 Thermo Fisher 253 gas source, dual inlet ion-ratio mass spectrometer at the Tripathi Lab at
202 UCLA, dedicated to measuring clumped isotopes in CO₂ (“Chewbacca”). The mass
203 spectrometer has been configured for the analysis of multiply substituted isotopologues
204 of CO₂, and is attached to a custom-built automated system for carbonate sample
205 preparation, purification, and introduction (after Passey et al., 2010). The system is
206 composed of (1) a Costech Zero Blank autosampler made of stainless steel that will pull
207 high vacuum, (2) a common acid bath for phosphoric acid digestion of samples at 90°C,
208 (3) cryogenic traps (dry ice and ethanol, and liquid nitrogen) for the purification and
209 collection of CO₂ and removal of water and other gases with low vapor pressures, (4) a
210 gas chromatograph with a packed column and a cryogenic trap to further purify CO₂
211 through the removal of organic contaminants, with helium being used as a carrier gas, (5)
212 cryogenic traps to remove the helium, and (6) a final set of valves and traps to further
213 purify CO₂ and transfer it into the bellows of the mass spectrometer. Measurements were
214 made to yield a stable 16-volt signal for mass 44, with peak centering and pressure
215 balancing before each acquisition. Background measurements were performed daily and

216 corrected for. Approximately six sample analyses and three standard analyses were
217 performed each day, and every sample was analyzed in triplicate. Both carbonate
218 standards and ‘equilibrated gases’ were analyzed. Equilibrated gases are composed of CO₂
219 that has been equilibrated at either 25 or 1000°C with different bulk $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ ratios.
220 At 1000°C, CO₂ has a stochastic distribution of isotopes between isotopologues. Gases in
221 quartz breakseals are heated to 1000°C for two hours and then quenched at room
222 temperature. These two types of standard gases are then purified and analyzed using the
223 same protocol as sample gases, and used for the calculation of sample Δ_{47} values on the
224 absolute reference frame (ARF; Dennis et al., 2011). Typical precision is 0.009‰ (1 s.e.),
225 equivalent to about 2°C (cf., Eiler, 2007; Guo et al., 2009; Huntington et al., 2009; Eagle
226 et al., 2010; Passey et al., 2010; Tripathi et al., 2010; Dennis et al., 2011; Thiagarajan et
227 al., 2011; Henkes et al., 2013). Modern lake tufas were also analyzed and compared to
228 current summer water temperature records to assess if there is evidence for kinetic
229 isotope effects or isotopic disequilibrium in the lake.

230 *Notation*

231 ¹³C-¹⁸O bond order in carbon dioxide is described using the variable Δ_{47} , which
232 refers to the per mille (‰) deviation in the abundance of CO₂ molecules with a mass of
233 47 amu from the abundance predicted by stochastic (random) mixing. Specifically Δ_{47} is
234 defined as:

235 Eq.1

$$\Delta_{47} = \left[\frac{R^{47}}{2R^{13} \cdot R^{18} + 2R^{17} \cdot R^{18} + R^{13} \cdot (R^{17})^2} - \frac{R^{46}}{2R^{18} + 2R^{13} \cdot R^{17} + (R^{17})^2} - \frac{R^{45}}{R^{13} + 2R^{17}} + 1 \right] \cdot 1000$$

where R refers to the ratio of the minor isotopologue to the major isotopologue of the molecule (or atom) of interest. As sample reactions were carried out at 90°C, we applied a published (Passey et al., 2010 as recalculated in Henkes et al., 2013) empirically-derived acid digestion fractionation correction of 0.092‰. Errors in reported Δ_{47} values and calculated temperatures include the propagated uncertainty in heated gas determination and in sample measurement (Huntington, et al., 2009). For calcite $\delta^{18}\text{O}$ calculations, an acid digestion fractionation factor of 1.00795400 was used (Swart et al., 1991). All clumped isotope data, including standards and heated gases, can be found in the Supplemental Information.

Calculation of temperatures and water isotope values

Temperatures are calculated from clumped isotope data on the ARF by applying the inorganic calibration of Ghosh et al. (2006) as reported in Dennis et al. (2011). This equation is:

$$\text{Eq. 2} \quad \Delta_{47} = (0.0636 \pm 0.0049 \times 10^6)/T^2 - (0.0047 \pm 0.0520)$$

where T is the temperature in degrees Kelvin. In addition, several other calibration equations were applied to the data in order to check for irregularities. Three of the calibrations utilized (Ghosh et al., 2006; Eagle et al; 2013; Zaruur et al., 2013) have a relatively steep slope, while others (Dennis and Schrag; 2010; Tang et al., 2014) have a

shallower slope, though all calibrations show general agreement in the 15-35°C range (Table S1; Figure S1). Δ_{47} temperatures values were used to calculate the $\delta^{18}\text{O}$ of water from which the mineral precipitated using paired measurements of carbonate $\delta^{18}\text{O}$ and the following published equation from Kim and O'Neil (1997):

Eq. 3
$$1000 \ln \alpha(\text{Calcite-H}_2\text{O}) = (18.03\text{E}10^3)/T - 32.42$$

where temperature again is in Kelvin.

Isotopic lake volume model

In order to reconstruct changes in lake volume during the course of stromatolite accretion, we used an oxygen isotope distillation/dilution model for determining lake volume changes in closed lake systems. The model was developed by Frantz et al. (2014) for stromatolites from the Eocene Green River Formation (Wyoming) where it was used to reconstruct major fluctuations in a periodically closed-basin paleolake from isotope measurements of successive stromatolite layers. The model inputs are a minimum lake volume $V_{\min}$, an assumed $\delta^{18}\text{O}_{\text{input}}$ of the average input of waters to the lake δ_i , and the $\delta^{18}\text{O}_{\text{water}}$ and water liquid-vapor fractionation factor α for each stromatolite layer. The $\delta^{18}\text{O}_{\text{water}}$ is calculated from the measured $\delta^{18}\text{O}_{\text{carbonate}}$ and clumped isotope temperature (Equation 3). Three temperature bounds were tested using both the most shallow (Tang et al., 2014) and steepest (the reported ARF temperatures) Δ_{47} calibrations, along with the Ghosh et al., 2006 calibration, in order get a maximum range of possible lake level and volume change.

275 The Frantz et al. (2014) model utilizes a Rayleigh distillation equation (Equation 4)
 276 to determine lake volume changes ($V_1 - V_0$) during periods of net evaporation (defined by
 277 an increase in the water isotope value between consecutive stromatolite layers, i.e., $\delta_1 >$
 278 δ_0), and a mixing equation (Equation 5) during periods of net filling (when the water
 279 isotope value decreases between consecutive layers, i.e., $\delta_1 < \delta_0$):

280 Eq. 4
$$V_1 = V_0 \left(\frac{\delta_1 + 1000}{\delta_0 + 1000} \right)^{\frac{1}{\alpha - 1}}$$

281 where V_1 and V_0 are the lake volumes associated with ^{18}O fractionation values δ_1 and δ_0 ,
 282 respectively; and α is the fractionation factor for the vapor-liquid phase transition of
 283 water calculated using the temperature-dependent equation of Horita and Wesolowski
 284 (1994) using a surface water temperature $T_{\text{surface}} = 25^\circ\text{C}$ (the average modern measured
 285 value).

286 Eq. 5
$$V_1 = V_0 \left(\frac{\delta_0 - \delta_i}{\delta_1 - \delta_i} \right)$$

287 where, again, V_1 and V_0 are the lake volumes associated with ^{18}O fractionation values δ_1
 288 and δ_0 , and δ_i is an assumed freshwater input $\delta^{18}\text{O}$ value (in this model, $\delta_i = -13.8\text{‰}$
 289 VSMOW, the modern Walker River value from Yuan et al. 2006A).

290 Because the model can be very sensitive to both T_{surface} and δ_i , sensitivity analyses were
 291 conducted using reasonable ranges of T_{surface} and δ_i values (Figure S3). Results of the
 292 model were less sensitive to the assumed surface water temperature than to the different

293 clumped isotope paleothermometer calibrations used, however, assuming temperatures
294 lower than 25°C results in slightly smaller lake volume changes. Calculated lake volumes
295 were not sensitive to δ_i for $\delta_i < -8\text{‰}$ VSMOW. Considering the latitude, mid-continental
296 location, the elevation of Walker Lake, and that the basin hydrology is heavily influenced
297 by snowmelt, average input values $> -8\text{‰}$ are unlikely (cf. Gat, 1996).

298 **Results**

299 *Petrography*

300   At the scale of tens of microns, the stromatolites have two main microfabrics—
301 calcite fans and micrite—with very little detrital material save for the very rare diatom or
302 ostracod shell, and no significant secondary cement phases (Figure 2 B-E). The fans
303 clearly grew in place. Four major zones along a proximal to distal transect along the
304 stromatolite are defined by the proportion of fan to micrite and whether the micrite forms
305 discrete layers. The proximal section of the stromatolite is composed of stacked calcite
306 fans approximately 100-500 microns in length, with minor amounts of micrite found
307 between the fans (Figure 2C). Lamination is plainly visible in hand section and is defined
308 by the boundaries of calcite fans, but is discontinuous and difficult to trace across the
309 structure. In the middle portion of the structure, the microtexture grades into distinct
310 bands of crystal fans alternating with layers of micrite (Figure 2C). The fans are smaller
311 (<100 microns), and the laminations are more continuous and traceable. There is also a
312 greater amount of detrital material in the structure. The distal part of the stromatolite
313 contains poorly-developed fans (<100 microns) and far thicker bands of micrite, as well

314 as abundant trapped detrital material, including ostracod shells and diatoms (Figure 2D).
315 At the very distal tip of the petal, micrite decreases and larger, well-formed crystal fans
316 re-appear (Figure 2E).

317 The tufa towers, in contrast, are comprised entirely of micrite with some detrital
318 material. The meter-scale structures consist of a central core, surrounded by clotted
319 carbonate, much like a smaller version of the Mono Lake tufa towers (though far less
320 extensive). They are not laminated, bear little resemblance texturally to the stromatolites,
321 and extremely porous and friable (Figure S4).

322 *Age dating*

323 The inception of the stromatolite petal growth occurred at $35,227 \pm 270$
324 Calibrated Years Before Present (CYBP – IntCal13). The middle section of the
325 stromatolite is dated at $34,477 \pm 250$ CYBP, and the most distal part of the structure is
326 $33,727 \pm 250$ CYBP (Table 1). Radiocarbon dating has previously been used extensively
327 in the Lake Lahontan system, and in Walker Lake specifically, to date numerous deposits
328 and estimate the timing and magnitude of lake level fluctuations (e.g., Benson et al.,
329 1995, 2013; Yuan et al, 2006B; Adams, 2007; Petryshyn et al., 2012). There are two
330 common sources of error in Lahontan carbonates noted in the radiocarbon studies (and
331 outlined in Benson, 1993): (1) Reservoir effect (low initial $^{14}\text{C}/\text{C}$ ratios in lake water),
332 and (2) addition of modern or ‘dead’ carbon after initial precipitation (via precipitation of
333 second generation cements either from younger lake water, or groundwater). The
334 reported ages have been corrected for a 300 yr reservoir effect (after Benson et al., 1995;

2013; Yuan et al., 2006B; Petryshyn et al., 2012), however, scaling reservoir effect to correspond to a larger lake (after Petryshyn et al., 2012) does not significantly impact the age dating results. Petrography analyses (discussed above) reveal no apparent second- or third-generation cements that could artificially bias the age dating. Radiocarbon dating of the stromatolite yields an average growth rate of 0.085 mm/yr for the entire petal structure, although we cannot be certain that growth was continuous or regular. Tufa towers used to compare the clumped isotope paleothermometry results to present-day water temperatures record a raw ^{14}C age of 155 ± 20 years before present (i.e. before 1959). During this time, climate has been stable, and average lake surface temperatures would not have been much different than present day. If we use the assumed reservoir effect of 300 years for this tufa, we return a raw ^{14}C age of -145 years (145 years after 1959, or 2104), which is clearly impossible. This result indicates that the tufa towers are modern, and the current reservoir effect in Walker Lake is 211 ± 20 years. The previous studies that calculated reservoir effect in Walker Lake (e.g., Benson et al., 1995) were performed when lake volume was greater, and thus could account for a larger reservoir effect than is currently observed (results are summarized in Table 1). Given the petrographic and geochemical evidence indicating no significant diagenesis, the fact that other Walker Lake carbonates have been successfully dated via ^{14}C (e.g., Benson et al., 1995; Yuan et al., 2006B; Petryshyn et al., 2012), and the test provided by the tufa towers, we are reasonably confident in the veracity of the ^{14}C dating of the Walker Lake carbonates, but future work could include other radiometric dating schemes to corroborate the ^{14}C dates.

357 *Clumped and Stable Isotope Data*

358   The clumped isotope data suggest that lake water warmed from the proximal to
359 distal ends of the stromatolite petal. The proximal end grew at an average temperature of
360 $19.6 \pm 3.1^{\circ}\text{C}$. The second portion of the petal grew at $23.4 \pm 2.0^{\circ}\text{C}$. The next portion
361 grew at $31.1 \pm 2.9^{\circ}\text{C}$. The most distal portion returned to cooler temperatures of $20.8 \pm$
362 2.1°C (results are summarized in Table 2; additional clumped isotope calibration
363 information can be found in the Supplemental Information). The modern lake tufas grew
364 at an average temperature of $27.9 \pm 2.1^{\circ}\text{C}$, which is a close match to the average summer
365 surface water temperatures of 22-26°C (Beutel et al., 2001).

366   Stable isotope values remained consistent throughout the stromatolite petal. The
367 proximal end of the petal yielded $\delta^{13}\text{C}_{\text{calcite}}$ values averaging $3.3 \pm 0.1\text{‰}$ VPDB, and
368 $\delta^{18}\text{O}_{\text{calcite}}$ values of $-3.1 \pm 0.2\text{‰}$ VPDB. The second portion averaged $\delta^{13}\text{C}_{\text{calcite}}$ values of
369 $3.2 \pm 0.1\text{‰}$ VPDB, and $\delta^{18}\text{O}_{\text{calcite}}$ values of $-3.1 \pm 0.2\text{‰}$ VPDB. The third portion
370 averaged $\delta^{13}\text{C}_{\text{calcite}}$ values of $3.2 \pm 0.1\text{‰}$ VPDB, and $\delta^{18}\text{O}_{\text{calcite}}$ values of $-3.1 \pm 0.3\text{‰}$
371 VPDB. The most distal portion averaged $\delta^{13}\text{C}_{\text{calcite}}$ values of $3.2 \pm 0.1\text{‰}$ VPDB, and
372 $\delta^{18}\text{O}_{\text{calcite}}$ values of $-2.9 \pm 0.1\text{‰}$ VPDB. Modern tufas have a $\delta^{13}\text{C}_{\text{calcite}}$ value of $0.5 \pm$
373 0.3‰ VPDB, and $\delta^{18}\text{O}_{\text{calcite}}$ value of $-3.0 \pm 0.3\text{‰}$ VPDB.

374   Using the Δ_{47} temperatures (ARF) and $\delta^{18}\text{O}_{\text{calcite}}$ ratios, $\delta^{18}\text{O}_{\text{water}}$ was calculated
375 for the ancient lake water using the equations of Kim and O'Neil (1997). The proximal
376 portion of the petal grew in water with a $\delta^{18}\text{O}$ value of $-1.3 \pm 0.5\text{‰}$ VSMOW. The
377 second portion accreted in a $\delta^{18}\text{O}_{\text{water}}$ of $-0.5 \pm 0.4\text{‰}$ VSMOW, the third portion accreted

378 in a $\delta^{18}\text{O}_{\text{water}}$ of $+1.0 \pm 0.3\text{‰}$ VSMOW, and the distal portion of the petal grew in water
379 with a $\delta^{18}\text{O}$ value of $-1.0 \pm 0.7\text{‰}$ VSMOW. Modern tufas yield a reconstructed $\delta^{18}\text{O}_{\text{water}}$
380 of $+0.4 \pm 0.6\text{‰}$ VSMOW. Current Walker Lake water oxygen isotope values are $+2.1\text{‰}$
381 VSMOW (Table 3).

382 **Discussion**

383 Several factors may impact the interpretation of clumped isotope data, including
384 the presence of contaminants, ground water input, kinetic isotope effects, and diagenesis,
385 so it is important to rule out such factors before proceeding with data interpretation.

386 *Clumped isotope data: Contamination?*

387 In order to screen for the presence of organic contaminants, we measured the
388 abundance of CO_2 with a mass of 48. High values in mass 48 that are dissimilar to what is
389 observed in standards would indicate the presence of organic material, which would
390 artificially drive the apparent temperatures cooler. No such irregularities are seen in the
391 data (Figure S2), consistent with uncontaminated samples.

392 *Clumped isotope data: Kinetic isotope effects?*

393 A recent theoretical study has shown that clumped isotope signatures in
394 carbonates can, at least in principle, be influenced by dissolved inorganic carbon (DIC)
395 speciation (Hill et al., 2014). Calcite precipitating rapidly from a DIC pool that is at a
396 high pH (where carbonate ion is dominant over bicarbonate ion) may result in an
397 overestimation of temperature; this effect should be modulated by growth rate (Hill et al.,

2014). To evaluate kinetic isotope effects, we evaluated data for modern tufas. We note that clumped isotope data for measured modern tufas reproduce temperatures similar to long-term average summer temperatures in the lake. Reconstructed $\delta^{18}\text{O}_{\text{water}}$ values from the modern tufas also match the range for modern lake $\delta^{18}\text{O}_{\text{water}}$. This is consistent with similar systematics governing tufas and the calibration materials being used for the reconstruction of temperatures, and indicates that the tufas are growing in isotopic equilibrium with lake water. Furthermore, application of published Δ_{47} temperature calibrations to data for modern lake carbonates from similar systems also accurately record summer lake temperatures, (e.g. Mitsunaga et al., 2014; Petryshyn et al., 2015). While we cannot say for certain from this experiment that stromatolites that grew in the lake ~30 ka are free of kinetic isotope effects, it seems unlikely to be a factor.

Clumped isotope data: Diagenesis?

There are several observations that indicate these carbonates have not been significantly altered through interaction with groundwater or through burial. There has been no significant burial of the structures, although they are covered with loose mud, sand, and alluvium from times when the lake was deeper (the Lahontan highstand is ~40 m above the stromatolite elevation). Groundwater in the region is isotopically light ($\delta^{18}\text{O} = -15.6\text{‰}$ V-SMOW) and very distinct from reconstructed fluid values. Micrite is found above and between growing fan structures, indicating that it formed coeval with the fans. There are no secondary carbonate precipitates such as equant spar or blocky calcite in the little pore space found in the structure. Fossil diatoms and ostracods are present and

419 identifiable in the distal section of the petal, and retain delicate shell structures indicating
420 no major diagenetic effects.

421 *Clumped isotope data: Groundwater?*

422   While there are no obvious second-generation cements present in the stromatolite
423 structure (as discussed above), it may be that the primary stromatolite structure formed
424 under the influence of groundwater. Specifically, the crystal fan microstructure, with the
425 fans growing surface-normal to the locus of accretion, is reminiscent of a spring deposit.

426   In order to discern whether or not the influx of ground/spring water would affect
427 our measurements, we turn to the modern tufa towers that were sampled from the
428 southwestern edge of the present-day shoreline. Given their morphology and placement,
429 there is little doubt that these towers formed from then interaction between calcium
430 rich/carbonate poor groundwater and calcium poor/highly alkaline lake water. Still, the
431 tufa towers faithfully record modern lake temperatures and radiocarbon dates.
432 Additionally, the reconstructed $\delta^{18}\text{O}_{\text{water}}$ of formation matches that of modern lake water
433 (+2.1 ‰ VSMOW), and not the significantly lighter groundwater (-15.6 ‰ VSMOW)
434 (Table 3).

435   Why would this be? The tufa towers are clearly formed as a result of groundwater
436 entering the lake, yet their geochemical signatures are consistent with lake water, not
437 ground water. As discussed above, Walker Lake is a calcium-limited system. Local
438 groundwater is comparatively calcium-rich (by 8 cm depth below current lake level,
439 calcium concentrations rise sixfold), but low in TCO_2 (only 27% as alkaline as lakewater

440 by 8 cm below current lake level; Table 3). Therefore, the calcite is most likely sourcing
441 its calcium from the groundwater, while the carbonate ion is coming from the lake and is
442 therefore in isotopic equilibrium with lake water.

443 The stromatolites, much like the tufa towers, do not have a calculated $\delta^{18}\text{O}_{\text{water}}$ of
444 formation indicative of groundwater carbonate ion input. Thus, while we cannot rule out
445 groundwater influence in the structure of the stromatolite, we find it probable that the
446 isotopic signatures are due to a lake water, not groundwater, carbonate ion source.

447 *Interpretation of clumped and stable isotopes in Walker Lake stromatolites*

448 Having ruled out contamination, kinetic effects, diagenesis, and the effects of
449 groundwater, we are confident in the veracity of the clumped isotope data. There are two
450 clear possible interpretations of this data:

451 1) Lake level remained constant, and the clumped isotopes temperatures appear to
452 change because of the varying proportions of micrite in the structure. The amount of
453 micrite increases dramatically from the proximal to the middle of the petal, and decreases
454 again at the distal end of the petal, a trend that mirrors the changes in temperature. The
455 presence of two different components in the stromatolite (crystal fans and micrite) in
456 differing proportions in the petal may impact the interpreted paleotemperatures, if for
457 instance, the micrite was forming in warm shallow water and trapped and bound in the
458 stromatolite, rather than forming *in situ*. Alternately,

459   2) Lake temperature revealed by clumped isotopes represents lake level change
460 during the ~2000 years of stromatolite accretion, from a deeper, cooler lake to a
461 shallower, warmer lake, and back again. Below we explore both possibilities.

462 *Origin of micrite*

463   If micrite precipitated in the warm surface water, and the fans precipitated
464 coevally *in situ* deeper in the lake below a thermocline, the apparent warming and
465 cooling trend may simply reflect the percent micrite bound into the structure: low at first,
466 increasing towards the middle, then low again in the most distal portion. It is commonly
467 thought that stromatolites form by either the trapping and binding action of microbes
468 through the biologically-mediated *in situ* precipitation of minerals, or by a complex
469 interplay between the two (Shapiro, 2007; Dupraz et al., 2006; Riding, 2011; Tice et al.,
470 2011; Bosak et al., 2013; Frantz et al., 2015). Therefore we must consider the origin of
471 the micrite seen in the structure to understand the significance of the clumped isotope
472 data from the stromatolite. Did micrite precipitate *in situ* or did it form higher in the
473 water column and was then trapped and bound into the structure?

474   While it is very difficult to discern the origins of the micrite given the scale of
475 crystals (<4 microns), the mesostructure of the stromatolite can add insight into the
476 formation mechanisms. The ‘petal’ stromatolite tested here grew laterally from the side
477 of the main structure (Figure 1C). If the micrite formed in surface waters, and was then
478 trapped and bound into the structure, we would expect to observe micrite blanketing the
479 top of the petal (where top is defined as parallel to the base of the stromatolite vs. the

480 distal end of the petal), or at least in higher abundance at the top. Instead, the micrite is
481 present in similar proportions at all angles, including the bottom of the petal, defying
482 expected gravitational relationships for sediment raining out of the water column (Figure
483 2). This indicates that it was not simply trapped and bound into the structure by microbes,
484 but rather precipitated *in situ*.

485 In addition to the petrographic analysis, we can consider the oxygen isotopes of
486 the stromatolite. The fractionation between water and mineral phases is primarily
487 temperature-dependent (Kim and O'Neil, 1997). In an isotopically well-mixed lake with
488 a single $\delta^{18}\text{O}_{\text{water}}$ value, calcite fans precipitating in a deeper, cooler portion of the lake
489 would have a much different $\delta^{18}\text{O}_{\text{carb}}$ than micrite precipitated in the shallow, warm water
490 (Figure 3). For instance, a fan precipitating from lakewater with a $\delta^{18}\text{O}_{\text{water}}$ value of 0‰
491 at 15°C would have a $\delta^{18}\text{O}_{\text{carb}}$ of -0.3‰ VPDB, while micrite precipitating from the same
492 water at 30°C would have a $\delta^{18}\text{O}_{\text{carb}}$ of -3.4‰ VPDB (Table 4). If the clumped isotope
493 trend is merely due to increasing the percentage of allochthonous micrite formed in
494 warmer waters, we would expect a decrease in $\delta^{18}\text{O}_{\text{carb}}$ as micrite proportions increased
495 (Figure 3). Instead, $\delta^{18}\text{O}_{\text{carb}}$ remains constant throughout the structure.

496 Interestingly, the younger stromatolites studied by Petryshyn et al. (2012) also
497 grew during a time of great lake level fluctuation (a lake level fall and rise of ~40m), yet
498 had constant $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}_{\text{carb}}$ values (varying less than 1‰ VPDB; Newton and
499 Grossman, 1988). Carbonate associated sulfate (CAS), however, did vary throughout the
500 structure (Berelson et al., 2009). Given that sulfate concentration behaves conservatively

501 in Walker Lake, Berelson et al. (2009) interpreted a rise in CAS to represent a lake level
502 fall, as evaporation would lead to increased sulfate concentration in the lake.
503 Furthermore, Yuan et al. (2006A) noted that the common proxies (e.g., $\delta^{18}\text{O}_{\text{carb}}$) are
504 difficult to interpret in Walker Lake. The study found that while changes in the lake's
505 hydrologic balance *usually* caused changes in the isotopic composition of lake water and
506 sediments, not all variations were recorded. Thus, the authors concluded that, considering
507 the complex nature of the Walker Lake system, extra caution should be taken when
508 interpreting isotopic results. The new clumped isotope proxy allows for a new and refined
509 view of climate and lake level change in Walker Lake.

510   Based on our petrographic observations and hypothetical vs. observed isotopic
511 trends, and the similarity to trends seen in other Walker Lake stromatolites, we suggest
512 that the clumped isotope data do not solely reflect the proportion of micrite in the
513 structure, but instead reflect changes in lake level/temperature that are also controlling
514 the morphology of the stromatolites (Figure 3). A similar pattern was reported by Frantz
515 et al. (2014) for Eocene stromatolites where an association between petrography and
516 paleolake temperature was found, with crystal fan formation by predominantly abiogenic
517 processes in cooler, deeper waters, and the formation of micritic layers in warmer,
518 shallower waters (presumably by the activities of microorganisms).

519 *Modeling*

520   In order to estimate the amount of lake level and volume change that would be
521 necessary to explain the data, and to see if that change is historically feasible, we applied

522 the Frantz et al. (2014) oxygen isotope/Rayleigh distillation model as described in the
523 methods section and Supplemental Information.

524 Basin closure is a prerequisite for application of the Frantz et al. (2014) isotope-
525 based lake volume model. The stromatolite petal exhibits a covariance of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$
526 values for most of the structure ($R = 0.74$ for the proximal portion; 0.54 for the second
527 portion; 0.87 for the third; and 0.91 for the most distal portion), which is characteristic of
528 a closed-basin lake with low groundwater input (Talbot, 1990). Regardless of the isotopic
529 covariance, lake level reconstructions indicate that Walker Lake would have been
530 separate from the rest of Lake Lahontan and hydrologically closed during the time the
531 stromatolite formed (Benson et al., 1995).

532 The model was used to calculate lake volumes using $\delta^{18}\text{O}_{\text{water}}$ values
533 reconstructed from the ARF temperatures as well as the Ghosh et al. (2006) and Tang et
534 al. (2014) temperature calibrations as end member scenarios. The volume model used a
535 minimum lake depth of 0 m above the stromatolites during stromatolite growth (since
536 stromatolites such as these must form subaqueously), which corresponds to a lake volume
537 of 77.4 km^3 (based lake bathymetry reported by Lopes and Smith, 2007). The model
538 output reveals a large decrease in lake volume over the initial three-quarters of petal
539 growth, followed by a similarly large increase during the final quarter of growth. When
540 using temperatures from the Tang et al. (2014) calibration, lake volume decreases by 39.2
541 km^3 , which, given the bathymetry of Walker Lake (Lopes and Smith, 2007), corresponds
542 to a lake level fall of 15.6 meters. The final section of the stromatolite experienced a lake

543 volume rise of 23.5 km³ and a lake level rise of 8.8 meters (Table 5). Using the Ghosh et
544 al. (2006) calibration, lake volume decreased 22.7 km³ from the most proximal to third
545 section. This corresponds to a lake level drop of 8.5 meters. In the final portion of the
546 stromatolite, lake volume increases 13.0 km³, which corresponds to a lake level rise of
547 4.6 meters. Using the temperatures plotted in the ARF of Dennis et al. (2011), the volume
548 decrease from the most proximal portion of the stromatolite to the third section is 21.8
549 km³, corresponding to a lake level fall of 8.1 meters. The lake volume increase for the
550 last section is 12.1 km³, or a lake level rise of 4.3 meters (Table 5, Figure 4). Historically,
551 this is plausible, as Lahontan Lake level records (Benson et al., 1995; 2013; Reheis et al.,
552 2014) often contain much larger fluctuations over much shorter time periods (Figure 4).

553 *Lake level fluctuation and Pleistocene Climate*

554   Currently in Walker Lake, the depth of the thermocline is located between 10 and
555 15 m water depth (Beutal et al, 2001). Summer water temperatures drop from ~24°C to
556 ~15°C over the span of the thermocline (Beutal et al., 2001). The data, which indicates a
557 lake level decline of ~8-16 m and a temperature increase of ~11-16 °C over the first three
558 zones of accretion, followed by a lake level increase of ~4-9 m with a temperature
559 decrease of ~10 °C over the last zone of accretion, is consistent with this thermocline
560 structure. Petrography shows a decrease in crystal fan abundance and size, and an
561 increase in micrite from the proximal to the middle of the structure consistent with lake
562 level fall, then a return to fan deposition consistent with a lake level rise (Figure 5). This
563 exact relationship between lake level and changes in microstructure has been reported in

564 another lacustrine system (Frantz et al. 2014). Isotopic modeling of the reconstructed
565 water isotopes also indicates an initial decrease in lake volume, followed by a lake
566 volume increase (Figure 5).

567   Major oscillations in lake level are known to have occurred on sub-10³ year
568 timescales in Lake Lahontan (Figure 4). Climate reconstructions of the Western Great
569 Basin over the last 35 ka show fluctuations in temperature and hydrology (Benson et al.,
570 1995) interpreted to reflect movement of the polar jet stream as the extent of the
571 Laurentide ice sheet changed (Antevs, 1948; Benson et al., 1995; Munroe and Laabs,
572 2012; Oster et al., 2015). A deflection of the jet stream over the Lahontan basin would
573 have driven changes in storm tracking that could result in increased rainfall while causing
574 a decrease in air temperature (Antevs, 1948; Clark and Bartlein, 1995; Oster et al., 2015).
575 These environmental changes could have led to abrupt lake level rise and concomitant
576 cooling. Retreat of ice would have driven both warming and aridity in the region, and a
577 decline in lake level (Benson et al., 1995). It is known that Lake Lahontan increased in
578 size rapidly from near present-day lake levels to its highstand at 1311 m above sea level
579 from 15.5 to 14.5 ka. The lake then decreased rapidly from 13 to 12 ka. It is therefore
580 feasible that these structures could record the reported rises and falls in lake volume in
581 ~2,000 years as the jet stream fluctuated. It is also possible that reductions in evaporation
582 could have contributed to a rise in lake level (Kaufman et al., 2003; Ibarra et al., 2014).

583   These data can be used to refine the Lahontan lake-level records of Benson et al.,
584 (1995; 2013) and Reheis et al. (2014) (Figure 4). Previously, only sparse data existed for

585 times >30,000 years, and the reconstructions are often at odds with each other (Figure 4).
586 This new data indicates that Walker Lake would have reached sizes much larger than
587 previously thought, though the lake would not have risen above the Adrien Valley Sill,
588 which would connect Walker Lake back to Lake Lahontan.

589 **Conclusions**

590 The interpretation of deposits from ancient Lake Lahontan has proven at times to
591 be problematic, leading to disputed and un-resolved lake level record beyond 30 ka. This
592 study found novel results for lake level fluctuations by coupling petrography, clumped
593 and traditional stable isotope analyses of a 33-35 ka carbonate stromatolite from Walker
594 Lake. Using the resulting data we were able to reconstruct paleoenvironmental change in
595 Walker Lake during the course of stromatolite growth. Initial stable isotope results would
596 indicate little to no hydrologic variability during stromatolite growth. However, when
597 combined with clumped isotope temperatures, the data are consistent with a lake volume
598 drop of $\sim 22\text{-}39 \text{ km}^3$, followed by a rise of $\sim 12\text{-}23 \text{ km}^3$ over of course of accretion,
599 possibly caused by precipitation changes due to fluctuation in the position of the polar jet
600 stream during a retreat of the Laurentide Ice Sheet, and/or evaporation-induced changes
601 in lake level. These data indicate a much larger volume for Walker Lake than previously
602 reported during this time period. The results of this study demonstrate the usefulness of
603 the clumped isotope paleothermometer when combined with traditional lake level
604 proxies, as well as the utility of stromatolites as a long-term terrestrial climate archive.

605 **Acknowledgements**

606 We wish to thank the 2014 International Geobiology Course, which is funded by The
 607 Agouron Foundation and The Gordon and Betty Moore Foundation. We would especially
 608 like to thank Lewis Ward and Zak Kerrigan for their insights. Much of this analysis was
 609 done at the USC Wrigley Institute for Environmental Studies. Additionally this work was
 610 aided by Ann Close, Amber Brown, Bradley Stevenson, Blake Stamps, Sean Loyd, John
 611 Spear, Russell Shapiro, Hope Johnson, and the entire Tripati Clumped Isotope Lab. AET
 612 acknowledges support from DOE grant DE-FG0213ER16402, ACS grant 51182-
 613 DNI2, and NSF grants OCE-1437166, EAR-1352212, EAR-1325054, and EAR-
 614 0949191. This work was supported by the “Laboratoire d’Excellence” LabexMER
 615 (ANR-10-LABX-19) and co-funded by a grant from the French government
 616 under the program “Investissements d’Avenir”.

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907 **Figure Captions**

908 **Figure 1:** Location, field photo, and hand sample of Walker Lake East stromatolite
 909 structure. A: Location of Walker Lake in Western Nevada, USA. B: Closeup of black box
 910 from A, showing Walker Lake in more detail, between the Wassuk and Gills Ranges near
 911 the town of Hawthorne, Nevada. Black star indicates stromatolite outcrop; Red star
 912 indicates location of tufa towers. C: Field photo of structures with rock hammer for scale.
 913 D: Cut slab of Walker Lake stromatolites. Main growth is in the ‘up’ direction, though
 914 several “petals” grow surface-normal to the main structure. One of these petals was
 915 chosen for analysis.

916 **Figure 2:** Composite microstructure of the Walker Lake East Stromatolite petal. A:
 917 Entire structure is 16.7 cm and has drill holes preserved. Position of zones 1-4 are

918 indicated. B: Photomicrograph of the proximal section of the petal. Crystal fans are >100
 919 microns in size. Micrite is included between fans. Scale bar is 100 microns. C:
 920 Photomicrograph of the middle section of the petal. Crystal fans are smaller in size (<100
 921 microns) and micrite is more abundant. Scale bar is 100 microns. D: Photomicrograph of
 922 the distal section of the petal. Crystal fans are poorly formed and small. Micrite is thicker
 923 and more abundant than in other sections. Detrital material (included a diatom) is bound
 924 into the structure Scale bar is 500 microns. E: Photomicrograph of the very proximal tip
 925 of the petal. Crystal fans become large again and micrite diminishes. Scale bar is 500
 926 microns.

927 **Figure 3:** Hypothetical isotopic results based on the origin of micrite in the
 928 stromatolites. Top Panel: Hypothesis One states that the lake level and $\delta^{18}\text{O}_{\text{water}}$ have
 929 varied over the course of stromatolite accretion and that this variation has been recorded
 930 in the Δ_{47} temperatures, though due to fractionation between the water and carbonate, the
 931 variation will not be seen in the $\delta^{18}\text{O}_{\text{minearal}}$. Middle panel: Hypothesis Two states that the
 932 Δ_{47} temperatures are tracking the relative abundance of micrite in the structure. In this
 933 case, lake level and $\delta^{18}\text{O}_{\text{water}}$ remain constant, but the two phases of calcite are
 934 precipitated at different temperature regimes; the micrite precipitates in warm, shallow
 935 water and is then trapped and bound into the deeper, cooler structure. In this case, the
 936 $\delta^{18}\text{O}_{\text{carb}}$ would vary significantly throughout the structure. Bottom Panel: Compilation of
 937 actual data, showing no variation in $\delta^{18}\text{O}_{\text{carb}}$ throughout the structure, but variations in
 938 clumped isotope temperature and reconstructed water isotopes; corresponding to large
 939 changes in lake level and confirming Hypothesis One.

940 **Figure 4:** Composite of reconstructions of Lahontan Lake surface elevation (in m above
941 sea level) for the time period explored in this study. Black line is a reproduction of the
942 work of Benson et al. (1995; 2013) based on radiocarbon ages of tufas and other
943 materials. Green line represents the work of Reheis et al. (2014) based on outcrop studies.
944 Orange line represents the results of this study. Solid lines represent what the previous
945 authors considered ‘robust’ data. Dashed lines are less robust. The purple line shows the
946 elevation of the stromatolite horizon, which should be a minimum lake level at that time
947 period. Our data indicates a much larger Walker Lake at this time than previously
948 thought, though lake level was still below the Adrien Valley Sill (red line).

949 Figure 5: Proposed model for the formation of Walker Lake stromatolites from 35-33 ka.
950 A: When growth begins, lake level is relatively high. The dominant morphology is calcite
951 fans. Growth temperature is low (between 10-19°C), and the lake water has a relatively
952 light oxygen isotope signature. The prevailing climate is relatively cool and rainy. B: In
953 zone 2, lake level as begun falling as the jet stream moves and conditions become warmer
954 and more arid. Oxygen isotopes become heavier, and the temperature of the water that
955 the stromatolites are forming in increases as they migrate up in the thermocline. Calcite
956 fans are still forming, but so is micrite. C: As the climate becomes even more arid, lake
957 level falls further. Temperatures reach their maximum for this structure (27-31 °C) as the
958 stromatolites are now above the thermocline. Due to evaporation, water oxygen isotopes
959 are heavy (above 0‰ V-SMOW). The amount of micrite in the structure reaches its
960 maximum. D: As the jet stream shifts back, the climate once more becomes cool and
961 rainy, leading to an increase in lake volume. The stromatolites are once more lower in the

962 thermocline. Temperatures and water isotopes decrease, and calcite fans once more
963 become the dominant crystal structure.

964 **Tables**

965 **Table 1:** ^{14}C ages of Walker Lake stromatolite petal sections, and modern tufas.
966 Radiocarbon concentrations are given as fractions of the Modern Standard, $\Delta^{14}\text{C}$, and
967 conventional radiocarbon age; ^{14}C age Before Present (BP), following the conventions of
968 Stuiver and Polach (1977). Ages are further corrected for the 300-year reservoir effect of
969 carbon in Walker Lake, and then calibrated using IntCAL13 terrestrial calibration curve
970 (Calibrated YBP).

971 **Table 2:** Stable and clumped isotope data with standard deviations (stdv) for all
972 stromatolite and tufa samples, including calculated water isotope data. Δ_{47} measurements
973 are reported as both ‰ and as temperatures in the absolute reference frame (ARF), as
974 outlined by Dennis et al. (2011). Standard error for temperature measurements (1 se) is
975 on the order of 2-3 °C. Water oxygen isotopes (‰ V-SMOW) are reconstructed from Δ_{47}
976 temperature and calcite isotopes using the equation Kim and O'Neil (1997). Water
977 isotope errors are propagated from the temperature uncertainty.

978 **Table 3:** Carbonate geochemistry parameters (calcium ion concentration, TCO_2 , and
979 water isotopes) from current lake water and ground water, measured in June of 2014 and
980 2015. Ground water measurements contain the depth from which the measurement was
981 taken. Oxygen isotopes were measured from nearby well water, while all other
982 parameters were measured from roughly 8cm beneath the sediment-water interface.

983 **Table 4:** Hypothetical data from carbonates precipitated at varying temperatures from 0-
984 30°C in an isotopically well-mixed lake ($\delta^{18}\text{O}_{\text{water}} = 0\text{‰}$ VSMOW). The predicted
985 fractionation factors (α) between the water and calcite phases are temperature dependent.
986 Given different temperatures of formation with the same $\delta^{18}\text{O}_{\text{water}}$, the calcite isotopes
987 would vary by as much as 6‰. CO_2 gas produced from dissolution of carbonate in acid
988 during the clumped isotope measurement would also vary in accordance with known
989 fractionation factors. Resulting Δ_{47} measurements would accurately reflect the
990 temperature of formation.

991 **Table 5:** Model output data for each portion of the stromatolite, using calibrations of
992 Ghosh et al., 2006, and Tang et al., 2014, and Dennis et al., 2011. Stromatolites were
993 divided into four sections with average temperatures and $\delta^{18}\text{O}_{\text{water}}$ values. These values
994 were used to calculate the changes in lake volume (km^3) between layers (after Frantz et
995 al., 2014). Using the bathymetry of Walker Lake (Lopes and Smith, 2007), volume
996 changes were converted to changes in depth (m). Changes in depth and volume between
997 successive layers are also listed.

Figure 1 revised

Figure 2 revised

Figure 3 revised

Figure 4

Figure 5 new

Table 1 revised

Sample name	Description	Fraction Modern	±	$\Delta^{14}\text{C}$ (‰)	±	^{14}C age (BP)	±
WLEbottom	Stromatlite	0.0195	0.0006	-980.5	0.6	31640	270
WLEmiddle	Stromatlite	0.0216	0.0006	-978.4	0.6	30820	250
WLEtop	Stromatlite	0.0243	0.0007	-975.7	0.7	29850	250
WLTT A	Modern Tufa	0.9808	0.0023	-19.2	2.3	155	20

Table 2 revised

ID	Sample Type	$\delta^{13}\text{C}$ mineral (VPDB)	$\delta^{13}\text{C}$ mineral stdv(VPDB)	$\delta^{18}\text{O}$ mineral (VPDB)	$\delta^{18}\text{O}$ mineral stdv(VPDB)
WLTTZa (pos 4)	Modern Tufa	0.280	0.006	-3.210	0.015
WLTTZa (pos 5)	Modern Tufa	0.243	0.003	-3.252	0.010
WLTT 1-A	Modern Tufa	0.354	0.006	-3.368	0.015
WLTT 2-A	Modern Tufa	0.316	0.008	-3.098	0.014
WLTT 1-D	Modern Tufa	0.823	0.007	-2.644	0.016
WLTT 2-D	Modern Tufa	0.845	0.007	-2.596	0.016
10. Walker Lake CF2 A4	Proximal (1)	3.293	0.004	-3.098	0.015
WLE-petal 2	Proximal (1)	3.274	0.004	-2.684	0.007
13 Walker lake CF2	Proximal (1)	3.251	0.006	-3.165	0.011
WLE-petal 3	Proximal (1)	3.282	0.005	-3.071	0.016
WLE-petal 4	Proximal (1)	3.335	0.003	-2.957	0.009
WLE-petal 5	Proximal (1)	3.093	0.006	-3.352	0.014
WLE-petal 6	Middle (2)	3.141	0.010	-2.841	0.013
14 Walker lake CF6	Middle (2)	3.181	0.006	-3.131	0.018
11. Walker Lake CF 6 A6	Middle (2)	3.118	0.003	-3.057	0.014
WLE-petal 7	Middle (2)	3.140	0.012	-3.315	0.015
12, Walker Lake CF 9 A8	Middle (2)	3.242	0.006	-2.915	0.015
WLE-petal 10	Middle (3)	3.152	0.008	-3.452	0.012
16 walker lake CF11	Middle (3)	3.345	0.007	-2.831	0.013
WLE-petal 11	Middle (3)	3.178	0.012	-3.098	0.023
14. Walker Lake CF13 A12	Middle (3)	3.227	0.004	-2.956	0.014
16. Walker Lake CF 14.5	Most Distal (4)	3.095	0.006	-3.000	0.014
19 Walker lake CF14.5	Most Distal (4)	3.167	0.004	-2.844	0.015
20. Walker Lake CF15	Most Distal (4)	3.220	0.008	-2.827	0.019
17. Walker Lake CF 15 B4	Most Distal (4)	3.222	0.005	-2.850	0.010

Table 3 revised

	[Ca ²⁺] mM	TCO2 (mM)	δ ¹⁸ O VSMOW
Lake Water	0.35	78.83	2.1 ±0.1
Ground Water	1.78 (8 cm depth)	21.27 (8 cm depth)	-15.6 ±0.3 (well water)

Table 4 revised

Sample ID	water $\delta^{18}\text{O}$ (VSMOW)	predicted α calcite	$\delta^{18}\text{O}$ mineral (VPDB)	$\delta^{18}\text{O}$ mineral (VSMOW)
Hypothetical 0 degress	0.0	1.034169249	3.2	34.2
Hypothetical 5 degrees	0.0	1.03295818	2.1	33.0
Hypothetical 10 degrees	0.0	1.031726279	0.8	31.7
Hypothetical 15 degrees	0.0	1.030598771	-0.3	30.6
Hypothetical 20 degrees	0.0	1.02949527	-1.4	29.5
Hypothetical 25 degrees	0.0	1.028460383	-2.4	28.4
Hypothetical 30 degrees	0.0	1.02740906	-3.4	27.4

Table 5 revised

Zone	$\delta^{18}\text{O}$ water (reconstructed)	Ave Δ_{47} T (K)- Tang et al., 2014	Ave Δ_{47} T (C)- Tang et al., 2014	Lake depth (m)	Lake vol (km ³)
1	-3.24	282.7	9.5	30.1	116.6
2	-1.91	288.4	15.2	23.3	101.0
3	0.55	300.3	27.1	14.5	77.4
4	-2.79	284.5	11.3	23.2	100.9

Zone	$\delta^{18}\text{O}$ water (reconstructed)	Ave Δ_{47} T (K)- Ghosh et al., 2006	Ave Δ_{47} T (C)- Ghosh et al., 2006	Lake depth (m)	Lake vol (km ³)
1	-2.02	289.5	16.4	22.9	100.1
2	-1.17	293.3	20.1	19.5	91.3
3	0.36	300.9	27.8	14.5	77.4
4	-1.67	290.7	17.6	19.1	90.3

Zone	$\delta^{18}\text{O}$ water (reconstruct ed)	Ave Δ_{47} T (K)-ARF (Dennis et al., 2011)	Ave Δ_{47} T (C)- ARF (Dennis et al., 2011)	Lake depth (m)	Lake vol (km ³)
1	-1.3	292.75	19.6	22.6	99.2
2	-0.5	296.55	23.4	19.4	91.0
3	1.0	304.25	31.1	14.5	77.4
4	-1.0	293.95	20.8	18.8	89.5