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1 **Title: Depositional evolution of an extinct sinter mound from source to outflow, El**

2 **Tatio, Chile**

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19

20 **Abstract:**

21 Siliceous sinter deposits from El Tatio, Chile, preserve a wide variety of depositional
22 environments and biosignatures, from high-temperature vent-proximal facies to distal
23 deposits dominated by silicified microbial mats. Four cores were drilled into an El Tatio
24 sinter mound and associated distal apron to investigate changes in hydrothermal
25 environments over geologic timescales. Sedimentary and geochemical analysis of multiple
26 sinter cores records the initiation and accretion of diverse depositional features still observed
27 today in El Tatio. y. cal. BPFacies adjacent to hydrothermal vents are dominated by
28 laminated sinter crusts on the steep margins of a high-temperature pool, with sparse microbial
29 preservation. Outer margins of the same pool contain extensive sinter columns up to ten
30 centimeters in length, precipitated during repeated cycles of pool overflow and subsequent
31 evaporation. Low-relief hydrothermal pools also form minor deposits within distal debris
32 aprons, and analogous pools are still active close to sampling locations. Debris aprons are
33 dominated by palisade, tufted, and arborescent microbial fabrics, with distinct mat textures
34 revealing well preserved microfossils. Surficial deposits in all cores feature detrital-rich and
35 microbially-influenced sinters overlying higher-temperature facies, indicating a relative
36 decrease in hydrothermal activity over time. Geochemical proxies for hydrothermal fluids
37 and detrital input match depositional interpretations based on sedimentary structures. ¹⁴C ages
38 from core deposits extend the mound's history by 11,000 years, recording at least three
39 thousand years of sinter deposition on top of glacial sandstones (13,337 – 10,232 y. cal. BP).
40 Importantly, this work provides a detailed depositional model unavailable through surficial
41 sedimentology alone.

42 Keywords:

43 Siliceous sinter, hot springs, El Tatio, microbialites, microfossils

44 **1. Introduction**

45 Siliceous sinter deposits form when amorphous silica (opal-A) precipitates in
46 terrestrial hydrothermal environments, such as geysers and hot springs. Water temperatures in
47 hot springs decrease from near-boiling vents to distal outflow channels ($< 30\text{ }^{\circ}\text{C}$), often
48 within several meters. Thermal gradients along geothermal springs produce lateral variations
49 in aqueous geochemistry, biological communities, and physical depositional processes. The
50 complex relationships between these parameters produce an extraordinarily diverse range of
51 micro-environments within localized areas. Silica precipitation within sinters preserves
52 microscopic features in exquisite detail, especially microbial cells and their extracellular
53 polymeric substances (EPS) (Walter, 1976a; Schultze-Lam et al., 1995; Cady & Farmer,
54 1996; Konhauser & Ferris, 1996; Jones et al., 1997, 1998, 2004, 2005; Konhauser et al.,
55 2001, 2003; Handley et al., 2005; Campbell et al., 2015; Gong et al., 2019). Hydrothermal
56 siliceous sinters therefore have the potential to record fine-scale transitions in sedimentology,
57 biology, and geochemistry over relatively short distances. Early sedimentary deposits (> 3
58 billion-years-old (Ga)) on Earth and Mars contain the oldest observed sinters (Ruff and
59 Farmer, 2016; Djokic et al., 2017; Ruff et al., 2019). On Earth, such deposits include the
60 earliest evidence for life in non-marine environments (Djokic et al., 2017), making work on
61 modern hot springs highly relevant for investigating ancient sedimentology and biosignatures
62 in paleo- and astrobiology.

63 Lateral gradients in sinter architecture are preserved in modern and ancient
64 hydrothermal systems (Cady and Farmer, 1996; Walter et al., 1996; Guidry and Chafetz,
65 2003a; Lowe and Braunstein, 2003; Hinman and Walter, 2005;).Despite variations in local
66 geochemistry and climate, consistent patterns of biogenic and abiogenic textures are present
67 across multiple localities, forming a well-established model for sinter depositional facies. The
68 stratigraphic evolution of hydrothermal sinters over geologic timeframes has also been

69 studied in deposits of various ages (Trewin, 1993; Walter et al., 1996; Campbell et al., 2001),
70 including drill cores through recently active localities (Jones et al., 1998; Guidry and Chafetz,
71 2003b; Lynne et al., 2008, Campbell et al., 2020). However, compared with elastic and
72 carbonate environments, several aspects of sinter stratigraphy remain poorly constrained. For
73 example, recent sinter cores only record cooler vent-distal environments such as outflow
74 channels and debris aprons, with primary water temperatures interpreted around 30 °C (Jones
75 et al., 1998; Guidry and Chafetz, 2003b; Lynne et al., 2008, Campbell et al., 2020). In
76 contrast, only one study in Devonian strata interprets long-term stratigraphy of higher-
77 temperature sinter facies (> 60 °C) closer to vent sources (Walter et al., 1996). The study of
78 vent-proximal sinters is especially relevant to the field of geobiology, as higher-temperature
79 zones are often inhabited by different microbial communities than cooler outflow deposits.
80 Temperatures extremely close to vents in certain environments can be too hot for life, and
81 distinguishing which sinters are actively mediated by microbes, passively preserving traces of
82 life, or are completely abiogenic, has been a continuing topic of research for decades (Walter,
83 1976a,b; Jones et al., 1997; Konhauser et al., 2001, 2003; Lowe & Braunstein, 2003; Handley
84 & Campbell, 2011; Campbell et al., 2015).

85 Here, we present sedimentary, geochemical, and geochronological data from four
86 cores drilled within an extinct sinter mound in El Tatio, Chile, including the first stratigraphic
87 profiles of cores from vent-proximal sinters. Facies changes and the preservation of
88 biological communities were observed on various scales using stratigraphic logs,
89 petrographic analyses, and scanning electron microscopy (SEM). Shifts in siliciclastic input
90 and primary silica precipitation were studied using geochemical signatures of sinters and
91 underlying sandstone. The detailed sedimentology and stratigraphy of multiple cores provides
92 the basis for a depositional model of the sinter mound, spanning thousands of years in an arid,
93 high elevation geothermal setting.

94 **2. Geological setting**

95 The El Tatio geothermal field is located in the Andes Mountains of northern Chile, 10
96 km west of the Chile-Bolivia border and 80 km north of San Pedro de Atacama (Fig. 1A). El
97 Tatio contains the world's highest geyser field (~4300 m a.s.l.), and the largest geyser field
98 (~30 km²) in the Southern Hemisphere (Zeil, 1959; Trujillo et al., 1969; Jones & Renaut,
99 1997; Glennon and Pfaff, 2003; Tassi et al., 2005; Munoz-Saez et al., 2018). The extreme
100 altitude of El Tatio produces a different environment compared with many sinter locations.
101 At 4300 m elevation, atmospheric pressure is 0.58 atm, oxygen partial pressure is 0.12 atm,
102 and the boiling temperature of water is 86 °C. Daily temperatures have been measured
103 between -5 and 25 °C during spring (Munoz-Saez et al., 2015), and drop to -30 °C in winter
104 (Fernandez-Turiel et al., 2005), resulting in frequent freeze-thaw cycles (Nicolau et al.,
105 2014). Wind speeds also follow daily cycles, increasing to 7 km/h several hours after sunrise
106 (Slagter et al., 2019). El Tatio is located in the arid Chilean Altiplano, receiving a mean
107 annual precipitation of 44 mm/yr, with a maximum of 160 mm/yr over the last 40 years
108 (DGA, 2017). Ultraviolet radiation levels are also elevated at the extreme altitudes of El
109 Tatio, with maximum UV-A and UV-B levels at 33 and 6 W/m², respectively (Phoenix et al.,
110 2006). Many environmental factors at El Tatio are analogous to conditions modelled on early
111 Earth and Mars, including high ultra-violet radiation and lower atmospheric pressure
112 (Cockell, 2000; Som et al., 2012, 2016; Ruff & Farmer, 2016).

113 El Tatio occurs above the Altiplano-Puna Magma Body (Ward et al., 2014), which
114 had an extensive explosive phase of eruption during the Miocene-Pleistocene (10-1 Ma) (De
115 Silva, 1989; Leidig and Zandt, 2003). Surficial hydrothermal activity in El Tatio primarily
116 occurs in the hanging wall of a northeast trending half graben, which is filled with ~1 km of
117 sub-horizontal Miocene and Pliocene ignimbrites, tuffs, and lavas, and covered by
118 Pleistocene/Holocene alluvial, glacial and sinter deposits (Healy, 1974; Lahsen and Trujillo,

119 1975). Hydrothermal waters are sourced from meteoric precipitation in the Bolivian Andes at
120 >5,000 m elevation, ~15 km east of El Tatio (e.g., Giggenbach, 1978; Cortecchi et al., 2005;
121 Munoz-Saez et al., 2018). Meteoric waters enter the subsurface and are heated by the
122 Altiplano-Puna Magma Body as they flow westward (Healy and Hochstein, 1973); heated
123 water interacts with the surrounding host rock to generate the fluids in the El Tatio reservoir,
124 at an equilibrium temperature >200°C (e.g., Tassi et al., 2010; Munoz-Saez et al., 2018). The
125 residence time of reservoir water is still unknown, but the lack of tritium sets a boundary >60
126 years (e.g., Cortecchi et al., 2005; Munoz-Saez et al., 2018). Surface geothermal activity
127 initiated within the peripheral margins of El Tatio in the late Pleistocene (~27 ka, Munoz-
128 Saez et al., 2020), with continuous activity during the Holocene (~11 ka, Slagter et al., 2019).

129 **3. Methods**

130 3.1 Sedimentology & petrography

131 Four cores (A, C, D, E) were drilled along a vent-proximal (A, C) to vent-distal
132 terrace (D, E) transect on an extinct sinter mound (22° 19' 40.41" S, 68° 0' 23.81" W, Fig.
133 1B-E, 2) in the Upper Basin of El Tatio, referred to in Slagter et al., 2019 as "Mound 502".
134 The cores were drilled vertically, with the exception of core D which was drilled at 27° from
135 vertical, using a modified portable Shawtool Drill machine (diamond bit, 41 mm diameter)
136 lubricated by cooled spring water from an adjacent spring. The cores were sawed in half
137 using a diamond-edged tungsten carbide rotary saw to measure sedimentary features and
138 stratigraphy; the working half of each core was used for petrographic and geochemical
139 sample collection, while the corresponding archival halves were used for non-destructive
140 analyses and stored at the National Museum of Natural History of Paris. Thin-sections were
141 made from 27 core samples (Institut Universitaire Européen de la Mer, Brest, France; Thin
142 Section Lab, Toul, France), with standard petrographic techniques used to interpret primary

143 micro-scale fabrics and secondary alteration. Detailed photomosaics of all thin-sections were
144 created using a Zeiss Axio Zoom V16 Microscope equipped with Zeiss Zen imaging software
145 (Institut de Biologie Paris-Seine).

146 3.2 Scanning electron microscopy

147 Scanning electron microscopy (SEM) was performed on rough, unpolished surfaces of 14
148 core samples fractured into 3-5 mm chunks. Samples were placed directly on aluminium
149 stages with carbon glue, air-dried, and sputter-coated with gold before imaging. SEM images
150 were obtained at the IPGP PARI analytical platform with a Zeiss EVO MA-10, at 15 kV and
151 12-mm working distance on SE mode.

3.3 ^{14}C dating

152 Five sinter samples (10 g each) were cut, cleaned, and ground down: two samples
153 from Core C and four samples from Core E (Table 1, Fig. 2). The fine-grained samples were
154 later treated with 1 N HCl for 24 h and rinsed with ultrapure water. All samples were
155 subsequently digested in 40% hydrofluoric acid (HF). The remaining carbonaceous (non-
156 soluble) material was rinsed with ultrapure water and centrifuged. All samples, along with
157 ^{14}C -free wood and coal blanks, were dried, combusted, and cryogenically purified before
158 being converted to graphite and measured by accelerator mass spectrometry at the UC Irvine
159 Keck-CCAMS facility, using OX-I as the primary standard. The ages are reported in Table 1
160 as years before present (y. BP), where present is AD 1950 (Stuiver and Polach, 1977).
161 Calibrated ages obtained on CALIB (<http://calib.qub.ac.uk>) using the SHCal13 dataset (Hogg
162 et al., 2013) are also reported as mean ages according to a 2σ confidence interval. Ages in the
163 text are reported as calibrated years before present (y. cal. BP). $\delta^{13}\text{C}$ values for samples were
164 measured to a precision of $<0.1\%$ relative to PDB, using a Thermo Finnigan Delta Plus
165 IRMS with Gas Bench input at UC Irvine (Table 1).

166 3.4 Major and trace element analysis

167 Powders from 76 core samples (~150 mg each) were analyzed at the Institut
168 Universitaire Européen de la Mer (IUEM, Brest). All weighing and preparation of sample
169 powders for solution mode ICP-MS and ICP-OES was carried out in a class 1000 clean lab at
170 IUEM. PFA vials were acid washed internally and externally overnight in near-boiling,
171 concentrated nitric acid (HNO₃), rinsed three times with 18.2 MΩ-cm ultrapure water, then
172 dried at room temperature and stored under clean conditions.

173 For major element analyses, 150-250 mg of powder was weighed into 30ml PFA vials
174 to which 3 ml of analytical-grade concentrated HF, 3 ml of concentrated HNO₃, and 1 ml of
175 concentrated HCl were added. Samples were digested completely overnight at 90°C, with
176 kerogen as the only residue, and upon cooling in closed beakers, were diluted rapidly with 97
177 ml of concentrated H₃BO₃ solution to neutralize HF and prevent loss of Si as SiF₄, following
178 the method of Cotten et al. (1995). Diluted samples were analyzed for major elements using a
179 Yvon Joriba Ultima 2 ICP-OES calibrated against 8 similarly prepared international
180 geostandards (CB, BELC, JB, BEN, ACE, WSE, ME, and GSN), with pure quartz as a high-
181 SiO₂ standard, at the Pôle Spectrométrie Océan (PSO) at IUEM in Brest, France.

182 For trace element analyses, powders were completely digested with a three-step,
183 heated HF-HNO₃-HCl attack, with kerogen as the only residue, using ultra-pure acids
184 distilled in sub-boiling stills under clean lab conditions. First, approximately 100-150 mg of
185 powder was placed in an acid-cleaned PFA vial. 1.5 mL of 16 M HNO₃ and 1.5 mL of 22 M
186 HF were added. Samples were then capped and left to react overnight at 80°C. Samples were
187 then uncapped and left to evaporate to dryness for 24 hours. 3 mL of freshly mixed aqua regia
188 was added to each residue, samples were capped and allowed to react for 6 hours at room
189 temperature to avoid excessive gas buildup. After 6 hours, caps were removed, and the
190 sample was evaporated overnight at 80°C to dryness. Finally, 3 mL of 6 M HCl was added to
191 each residue and samples were heated for ~24h at 80°C. Prior to analysis, 100 mL of each

192 sample was diluted to 5 mL with 2% HNO₃, and 1.5 mL of each sample was
193 archived. International geostandards IF-G, BCR-2, and BHVO-2 were prepared in duplicate
194 in the same batch and treated as unknowns to monitor precision.

195 Diluted samples were measured in solution mode on a Thermo Element2 HR-ICP-MS
196 at the PSO. Samples were calibrated by gravimetrically-prepared multi-element standards at
197 concentrations of 0.5, 5, and 50 ng/g that were measured repeatedly throughout the session.
198 Additionally, a 5 ng/g In (Indium) standard was added to each sample, standard, and rinse,
199 and was used to monitor signal stability and to correct for instrumental drift across the
200 session. Each sample and standard were bracketed by a rinse for which data was also
201 acquired to determine the method detection limit.

202 Rare earth element (REE) concentrations were normalized against background clastic
203 deposition at Mound 502, as well as traditional standards such as chondritic meteorites and
204 post-Archean Australian shale (PAAS). REE concentrations in Mound 502 clastic deposits
205 were measured from a sinter-free massive sandstone sample from the lowest half-meter of
206 Core E (Fig. 2). REE data shown and discussed below are clastic-normalized, for easier
207 interpretation of clastic vs. hydrothermal influence in sinter deposits. Chondrite- and PAAS-
208 normalized REE values are present in Supplementary Table 1.

209 Eu anomalies were measured to compare the relative influence of hydrothermal fluids
210 between samples (Michard, 1989; Lewis et al., 1997; Feng et al., 2014). Eu anomalies
211 represent enrichments or depletions of Eu compared with Sm and Gd within samples, and are
212 typically calculated in a given sample using the equation:

213
$$\text{Eu anomaly} = 2 * \text{Eu} / (\text{Sm} + \text{Gd})$$

214 Anomaly values greater than 1 represent Eu enrichment, and are typically interpreted as
215 evidence for higher hydrothermal influence (Michard, 1989; Lewis et al., 1997; Feng et al.,
216 2014).

217 **4. Depositional facies**

218 Ten sedimentary facies were identified during core analysis, each representing a
219 different depositional environment (Fig. 2, Table 2). All facies described have modern
220 equivalents within active El Tatio hydrothermal systems. Figure 2 illustrates the progression
221 of sedimentary facies and interpreted environments over time, and Table 2 contains
222 diagnostic features for each facies.

223 Mound 502 rises 2-3 meters above surrounding topography, and is 20 meters in
224 diameter, with a roughly circular outline (Fig. 1C). Western margins are fairly steep, while
225 eastern margins gradually slope into a debris apron. The mound's interior is dominated by a
226 debris-filled pit 1.5 meters deep and five meters wide which represents the inactive former
227 vent, surrounded by a one-meter wide flat-topped sinter rim. As with exterior mound slopes,
228 inner pit margins are steeper on the western face, surrounding the pit's deepest surface in the
229 south-western corner (Fig. 1C, D). Mound 502 is bordered to the north by a low-lying marsh
230 with geothermal and meteoric water sources, bordered by vegetated hills further north and
231 west (Fig. 1B). Relatively flat, barren plains border the mound to the south and east (Fig. 1B).
232 While the mound itself contains minimal visible hydrothermal activity, adjacent plains within
233 ten meters to the south and east contain several active pools and widespread fumaroles.
234 Boiling pool margins contain small columnar and spicular sinters (see Section 4.4), but do not
235 form mounds above the surrounding flat topography.

236 Core A (3.38 m long) was drilled within the eastern sinter rim, adjacent to the inner
237 pit margin (Fig. 1C, D). Cores C (1.94 m long) and D (0.98 m long) were drilled ~1.5 meters

238 below the top of Core A along the outer eastern and southern margins, respectively. Core D
239 was drilled 27° from vertical (Fig. 1E). Core E (5.57 m long) was drilled in a flat debris apron
240 15 meters away from the sinter margin.

241 4.1 Inclined laminated sinter crust

242 Laminated sinter crusts are defined by dense, surface-normal fabrics of amorphous
243 silica, forming mm-scale laminae of consistent thickness (Fig. 3). Similar deposits in
244 Yellowstone siliceous sinters have been labeled “vent-proximal facies” and “laminated
245 crusts” (Guidry & Chafetz, 2003a, c), as well as “dendritic sinter facies” (Lowe &
246 Braunstein, 2003). The term “dendritic” is avoided in this study to prevent confusion with
247 branching arborescent sinter elsewhere in El Tatio cores (see Section 4.5).

248 Laminated crusts are divided into two separate facies in Mound 502 based on bed
249 orientation and depositional interpretation: inclined crusts and planar crusts (Section 4.2).
250 Inclined crusts are only observed in Core A, between 34 and 217 cm depth, with a shift in
251 textural preservation at 164 cm depth (see below). The inclination of laminae and bedding
252 varies between 45° and nearly vertical, with most areas around 70° or higher inclination (Fig.
253 3). Well-preserved zones are composed of silica laminae 1-2 mm thick, alternating between
254 pale surface-normal fabrics and thinner (< 1 mm), darker layers of more massive silica (Fig.
255 3A-C). No microbial textures or cells are observed, and detrital grains are rare or absent.

256 Below 164 cm depth in Core A, inclined crusts are gray or light green and have little
257 to no porosity. Detailed fabrics are only visible in thin sections, where individual laminae
258 contain surface-normal clusters of amorphous silica, which flare and diverge away from
259 bedding planes (Fig. 3D). Preservation varies between individual laminae, with more altered
260 layers only exhibiting faint layering (Fig. 3D). Samples also contain dark, linear cross-cutting

261 features occurring 1-3 cm apart (Fig. 3B). Cross-cutting textures disrupt and propagate
262 through several centimeters of inclined crusts, independent of textural preservation.

263 Above 164 cm, inclined crusts are more well-preserved, and surface-normal fabrics
264 become increasingly visible in hand samples (Fig. 3A-C). In the uppermost facies (34-51 cm
265 depth), inclined crusts are white, extremely friable, and contain very few detrital grains (Fig.
266 3C). Inclination decreases upward from 70 to 45°, with rare cross-cutting textures present. In
267 SEM images of the uppermost inclined facies, surface-normal fabrics are composed of
268 elongated tubes of silica ~50 µm wide and up to 1 mm in length (Fig. 3E, F). Closer
269 examination of individual tubes reveals irregularly porous cores with scalloped circular
270 textures, surrounded by µm-scale laminae around the outermost tube margins (Fig. 3F). Rare
271 silicified filaments are also present within individual laminae, comprising less than 5% of
272 inclined textures (Fig. 3G, H). Filaments are smaller than surrounding surface-normal tubes
273 (10-20 µm wide, <200 µm long), and are not consistently oriented (Fig. 3G, H). Instead,
274 filaments more closely resemble preserved cells in microbial facies (Figs. 8, 10, 11, see
275 Sections 4.5-4.7).

276 Laminated sinter crusts are observed in subaqueous hydrothermal environments
277 where silica precipitation typically outpaces microbial mat formation, such as vent-proximal
278 pools (Guidry & Chafetz, 2003a, Lowe & Braunstein, 2003). SEM images most closely
279 resemble “shrub columns” with abundant porous silica structures described by Jones &
280 Renaut (2003) from vent-proximal deposits in New Zealand. Internal porosity within
281 branches of shrub columns are attributed to secondary dissolution, which can replace
282 microbial cells or abiogenic mineral dendrites (Jones & Renaut, 2003). While porous surface-
283 normal textures are ambiguously biogenic, the occasional presence of filamentous
284 microfossils indicates that microbial communities were present.

285 The high angle of lamination over nearly 2 meters suggests extensive precipitation on
286 steep surfaces, most likely the inner subaqueous surfaces of a large high-temperature pool.
287 The interpretation fits with Mound 502's current near-vent architecture, which consists of a
288 5-meter-wide steep-sided pit. The darker surface-normal lineations which disrupt laminae are
289 potentially the propagations of corrugated or "rilled" bedforms commonly seen on vent walls
290 (Braunstein & Lowe, 2001; Guidry & Chafetz, 2003a; Lowe & Braunstein, 2003). Therefore,
291 the inclined fibrous facies in Core A are interpreted as the margins of a steep-walled pool
292 immediately above a hydrothermal vent, most likely above 70 °C.

293 4.2 Planar laminated sinter crusts

294 Planar laminated crusts contain thin layers of surface-normal silica fabrics (Fig. 4),
295 similar to inclined sinter crusts (see Section 4.1), but exhibit consistently horizontal bedding.
296 Planar crusts are only present in Core A from 11 to 34 cm depth, immediately above inclined
297 crust facies. Planar and inclined crusts are separated from each other by a sharp
298 unconformity. As with inclined crusts, laminae in planar facies are 1-2 mm thick, containing
299 pale surface-normal fabrics separated by thin, darker massive laminae (Fig. 4A). Boundaries
300 between light and dark layers are more distinct in planar than inclined fibrous facies. Detrital
301 sinter pieces and siliciclastic material are rare, but more abundant in darker laminae (Fig.
302 4A). Porosity in both light and dark layers is low on mm- to cm-scales.

303 SEM imaging of planar sinter crusts reveals two distinct styles of preservation (Fig.
304 4B,C). Light laminae are composed of adjacent vertical chains of silica spherules ~100 μm in
305 diameter (Fig. 4B). Vertical textures in light laminae superficially resemble surface-normal
306 textures within inclined sinters, but notably lack porous, scalloped cores (Figs. 3E, F, 4B). By
307 contrast, dark laminae contain linear highly-porous zones 20 μm wide, surrounded by rinds of

308 amorphous silica (Fig. 4C). Pores in planar crusts lack consistent orientation, and are
309 typically smaller and less interconnected than in inclined facies.

310 As with inclined sinter crusts, planar facies are observed within vent-proximal high-
311 temperature deposits (Guidry & Chafetz, 2003a, Lowe & Braunstein, 2003). In Mound 502,
312 planar laminated crusts are only observed within the flat-topped inner rim which lines the
313 inner pit (Fig. 1D). While inclined crusts were most likely deposited on the steep subaqueous
314 margins of the main inner pit, planar crusts potentially represent adjacent, shallower flat-
315 bottomed pools.

316 Lowe & Braunstein, 2003 described scattered microfossil filaments in Yellowstone
317 sinter crusts (“dendritic sinter facies”), though many more samples contained no trace of
318 microbial presence. Similarly, planar crusts in El Tatio cores contain no definitive
319 microfossils, instead sharing similarities with ambiguously biogenic “shrub columns”
320 described by Jones & Renaut, 2003. The lack of biosignatures does not preclude a microbial
321 presence in planar sinter crusts, but supports a silica precipitation patterns less conducive to
322 microfossil preservation (Walter, 1976b; Lowe & Braunstein, 2003; Hinman & Walter,
323 2005).

324 4.3 Particulate sinter

325 Particulate sinters are defined as beds of “botryoidal grains and chains” over 5 μm in
326 diameter, frequently associated with detrital grains (Braunstein & Lowe, 2001; Lowe &
327 Braunstein, 2003; Watts-Henwood et al., 2017). Particulate sinter facies occur at the top of
328 Core A, and in two zones within Core E (103-125 cm, 165-195 cm depth). Particulate beds
329 are typically gray with brown detrital material (Fig. 5A), and are closely associated with
330 laminated sinter crusts, arborescent sinters (see Section 4.5), and massive sinters (see Section
331 4.8).

332 Particulate sinters are defined by diffuse planar to wavy laminae between 1 and 5 mm
333 thick, comprised of rounded 50-100 μm -wide agglomerates of opal spherules surrounded by
334 pore space (Fig. 5B-E). Well-preserved hand samples of particulate facies show similar
335 textures to associated thin sections (Fig. 5A), though many others are difficult to discern from
336 massive sinter facies prior to microscopic analysis (see section 4.8). Detrital material is
337 abundant in specific layers as individual sand grains (Fig. 5C). No microfossils or preserved
338 microbial mat textures are visible using petrographic or scanning electron microscopy.

339 Particulate facies are interpreted as high-temperature deposits along the bases of non-
340 boiling hydrothermal pools, which occur both proximal and distal to vent sources. Spherical
341 silica particles precipitate onto various substrates such as microbial cells and wind-blown
342 detrital sediments, forming a sticky, laminar ooze comprised of amorphous, wetted silica and
343 thin biofilms (Rimstidt & Cole, 1983; Lowe & Brownstein, 2001; Braunstein & Lowe, 2003;
344 Watts-Henwood et al., 2017). Microbial cells can form a major component of particulate
345 sinters, preserving silicified cells or cell fragments of microbes from various temperature
346 regimes (Jones & Renaut, 1996; Braunstein & Lowe, 2003). Sediments are agitated by wind
347 and hydrothermal currents, forming sinuous, ripple-like bedforms on pool floors (Lowe &
348 Brownstein, 2001; Watts-Henwood et al., 2017). Similar textures are observed in extinct
349 sinter mounds elsewhere in El Tatio, bearing irregular surfaces of low-relief ridges (Sanchez-
350 Garcia et al., 2019). Modern high-temperature pools are present within debris aprons several
351 meters downhill from Core E, providing a useful analogue for distal pool environments
352 indicated by particulate facies.

353 4.4 Columnar and spicular sinter

354 Cores C and E contain several distinct facies dominated by clusters of elongate sinter
355 protrusions with convex laminae, which can be classified as columns and spicules (Figs. 6,

356 7). *Columns* are between 0.1 and 1 cm wide, and can exceed 5 cm in length (Walter et al.,
357 1976b). *Spicules* are less than 1 mm in diameter, reaching up to 1 cm long (Walter et al.,
358 1976b). The internal structure of spicules and columns are nearly identical, with definitions
359 for each based on size (Walter, 1976b), though terminology varies slightly between studies
360 (Jones & Renaut, 1997, 2003, 2006; Braunstein & Lowe 2001).

361 Columnar sinter is most prevalent in Core C, from 70 to 167.5 cm. Individual
362 columns are 0.5-1 cm wide, and 1-5.5 cm long (Fig. 6C-E). Core C often truncates column
363 lengths, but surface exposures of columnar facies verify the observed size ranges (Fig. 6A,
364 B). In transverse cross-section, columns are typically circular to oval in shape (Fig. 6E).
365 Columns frequently exhibit branching morphologies, with individual branches reaching no
366 more than 30° of inclination from parent columns (Fig. 6C, D). The only change in column
367 morphology occurs at 123.5 cm in Core C, where column orientation switches from
368 horizontal to 60° inclination. The change in angle occurs above an unconformity with planar
369 laminated sinter (Fig. 6E).

370 Sinter columns in Core C are internally laminated, with laminations visible in
371 longitudinal and transverse cross-sections (Figs. 6, 7). Laminae are convex in the direction of
372 growth, either upwards in angled columns, or laterally in horizontal columns (Fig. 6C, D, 7A,
373 C, D). The synoptic relief of individual layers within columns is relatively low, between 0.5
374 and 2.0 mm (Fig. 7A, C, D). Layers have low porosity, with sub-mm pores scattered
375 throughout columns. SEM imaging also reveal column layers (Fig. 7F), with the rare
376 presence of hollow silica tubes on the outer surfaces of columns (Fig. 7G). Interstitial gaps
377 between columns are porous, with coarse or fine sediments on the bottom of pores serving as
378 geopetal, or way-up, indicators (Fig. 7D, E). Fine geopetal sediments are often cross-cut with
379 secondary silica veins (Fig. 7E). Other mineral phases include rims of yellow and brown
380 oxidized crusts pervasively surrounding columnar edges (Fig. 7A, C, D). The presence of

381 crusts on the ends of broken columns indicates secondary precipitation after initial column
382 growth (Fig. 7C).

383 Columnar sinter is also present at 364 cm depth in Core E (Figs. 6F, 7B), within a 3
384 cm thick bed. Core E columns are vertical features ~0.5 cm in width and 1 cm in height,
385 occurring in isolated clusters of one to three structures. Internal morphology closely
386 resembles Core C columns, bearing convex-upwards laminae with 0.5 mm relief (Figs. 6F,
387 7B). Laminae in Core E columns are thicker than Core C (0.25 - 0.5 mm), and contain higher
388 amounts of detrital material. Lower columnar margins have minor yellow mineral crusts,
389 lacking dendritic textures observed in Core C. Columns are immediately overlain by tufted
390 microbial sinters, but no evidence of significant microbial communities is present at the time
391 of column deposition.

392 Spicular sinter occurs in a 5 cm thick bed immediately below columnar sinters in Core
393 C, at 175 cm depth. Spicules are 1 mm wide and up to 1 cm long, with individual structures
394 curving upwards from 30 to 70° inclination (Fig. 6G). As with Core C columnar sinters,
395 spicules are tightly packed and bordered by yellow and red mineral crusts. Laminae are not
396 observed in spicule cross-sections. The spicule layer is immediately and unconformably
397 overlain by horizontal columnar sinters.

398 Columnar and spicular sinters are commonly observed around boiling pools and
399 geysers, in zones which exhibit frequent wetting and drying (Walter, 1976b; Jones et al.,
400 1997, 1998; Jones & Renaut, 1997, 2003, 2006; Braunstein & Lowe, 2001; Guidry &
401 Chafetz, 2003a, Lowe & Braunstein, 2003; Boudreau & Lynne, 2012). Splashes and surges of
402 hot, silica -rich water evaporate on nearby surfaces, precipitating opal-A. Over time, silica
403 precipitation produces micro-scale topography that progresses into columnar morphologies,
404 with capillary motion of water within columns providing silica over time (Walter, 1972,

405 1976b; Braunstein & Lowe, 2001; Jones & Renaut 1997, 2003, 2006; Mountain et al., 2003).
406 Splash zones close to pools and vents form spicular textures, while larger columns occur
407 further away in surge and overflow environments (Walter, 1972, 1976b; Braunstein & Lowe
408 2001). The upward increasing angle of columnar growth in Core C is likely due to local
409 topography, with horizontal columns growing on nearly-vertical mound faces, and increasing
410 angles of growth on progressively flatter surfaces closer to the tops of sinter rims. Incipient
411 growth of vertical columns in El Tatio occurs around modern boiling pools several meters
412 south of Core E (Fig. 6B). The modern pool-margin textures are nearly identical in size and
413 orientation to Core E columns. Therefore, columnar sinters from Cores C and E are
414 interpreted as silica precipitates that formed in the former splash and surge zones of boiling
415 pools.

416 Columnar sinters frequently preserve microbial communities, both within columnar
417 laminae and interstitial spaces between columns (Cady & Farmer, 1996; Jones et al., 1997,
418 1998; Guidry & Chafetz, 2003; Jones & Renaut, 2003, 2006; Handley et al., 2005, 2008). In
419 other columnar sinter locations, biosignatures are less apparent (Walter, 1976b; Braunstein &
420 Lowe, 2001; Lowe & Braunstein, 2003). Most columnar facies are interpreted as forming via
421 biogenic and abiogenic silicification, with capillary forces, surge events, and precipitation
422 within microbial communities dominating at different points in time (Jones et al., 1997; Jones
423 & Renaut, 2003, 2006). In El Tatio cores, microbial textures in columnar sinter are obscured
424 by diagenesis, with only isolated, putative microfossils within inter-columnar spaces (Fig.
425 7G). The columnar sinters examined are conservatively interpreted as forming through
426 capillary forces after periodic surge events, with microbial communities likely present in
427 cooler, moist areas between columns.

428 4.5 Arborescent sinter

429 Arborescent sinter is defined in this study as horizontal beds composed of surface-
430 normal, extensively branching siliceous textures up to 1 cm in height (Fig. 8). The same
431 textures are nearly identical to “siliceous shrubs” described in detail by Guidry & Chafetz,
432 2003c. However, the term “shrub” has been used to describe various different sinter textures
433 such as laminated sinter crusts (“dendritic sinter facies” in Lowe & Braunstein, 2003),
434 microbial palisade fabrics (Cady & Farmer, 1998; Walter et al., 1998; Campbell et al., 2001;
435 see Section 4.6), and geyser-associated facies (Jones & Renaut, 2003). Similarly, the term
436 “dendritic” is commonly applied to branching amorphous or mineral phases, and has been
437 used in various sinter fabrics, such as laminated crusts, spicules, and microbial mats (Jones &
438 Renaut, 2003; Lowe & Braunstein, 2003; Hinman & Walter; 2005), as well as phases other
439 than amorphous silica, including quartz, calcite, and manganese (Bargar & Beeson 1981;
440 Jones et al., 1998; Walter et al., 1998; Guidry & Chafetz, 2003b). By contrast, the term
441 “arborescent” is only applied to siliceous shrubs as described in Guidry & Chafetz, 2003a &
442 c, the previously published descriptions which most closely resemble the El Tatio sinter
443 facies presented here.

444 Various horizons in Cores C and E (Fig. 2) contain arborescent sinter. In Core C (0-3,
445 30-34 cm depth) and deeper in Core E (310-320 cm depth), arborescent sinters form minor
446 units of 1 cm-thick beds interspersed between sandstones, microbial, and massive sinter
447 facies (Fig. 8A-B). Arborescent layers higher in Core E form persistent beds between 142-
448 187 cm (Fig. 8C) and are closely associated with particulate sinters (see Section 4.3). In
449 Cores C and E, arborescent sinter contains individual branching structures that reach up to 1
450 cm high, with individual branches less than 1 mm in diameter (Fig. 8A-C). Arborescent zones
451 form highly porous areas compared with over- and underlying facies, and are often separated
452 by mm-scale interstitial layers of less porous sinter (Fig. 8B, C). Interstitial sinter layers also

453 contain higher abundances of detrital grains, often appearing darker in color than white or
454 light gray arborescent layers (Fig. 8B, C).

455 Petrographic analysis of arborescent structures reveals branching chains of
456 amalgamated siliceous spheres up to ~100 μm in diameter (Fig. 8D). Individual arborescent
457 features and branches are often separated by pore space, with more extensive silica
458 precipitation resulting in smaller interstitial voids. SEM images show similar branching
459 textures as observed in thin sections (Fig. 8E), as well as the presence of silicified
460 microfossils and higher quantities of detrital material at the top of individual arborescent beds
461 (Fig. 8E-G). Microbial filaments are 10-20 μm wide and over 200 μm long, resembling cells
462 preserved in palisade mat and tufted mat facies (see Sections 4.6 & 4.7). Microfossils are
463 found in clusters within and around larger branching features at lower densities than other
464 microbial facies in El Tatio cores (Figs. 10, 11).

465 In active hydrothermal systems, arborescent sinters form within discharge channels as
466 hydrothermal waters flow away from vent sources (Guidry & Chafetz, 2003a,c). Within such
467 depositional environments, arborescent facies appear to be temperature-independent, forming
468 in subaqueous zones between 15 and 80 $^{\circ}\text{C}$ (Guidry & Chafetz, 2003c). Microbial
469 communities preserved in El Tatio arborescent facies corroborate findings from similar
470 Yellowstone sinters (Guidry & Chafetz, 2003c). Guidry & Chafetz, 2003c noted the presence
471 of branching “siliceous shrubs” in discharge channels with higher outflow volume, while
472 laminated microbial mats were abundant in lower outflow zones. Based on these
473 observations, arborescent sinters in El Tatio cores are interpreted as silicified microbial
474 communities in high-volume discharge channels, both in vent-proximal zones and distal
475 aprons.

476 4.6 Fenestral palisade sinter

477 Palisade sinters are defined by silicified filamentous microfossils alternating between
478 layers of surface-normal and surface-parallel orientation (Cady & Farmer, 1996; Walter et al.,
479 1996; Campbell et al., 2001; Gong et al., 2019). Similar textures are observed, but not
480 specifically labeled as palisades, in hydrothermal stromatolites (Konhauser et al., 2004; Jones
481 et al., 1997, 1998, 2005; Berelson et al., 2011), with a variety of terms applied to microfossil
482 orientation (erect vs. prostrate, U- vs. P-laminae, and upright vs. flat-lying). Palisades occur
483 in the top 50 centimeters of Cores C, D, and E, and between 215 and 244 cm depth in Core E.
484 Most El Tatio core palisades contain high abundances of fenestrae (Figs. 9, 10), defined as
485 primary pore spaces preserved within microbial textures (Tebbutt et al., 1965; Mata et al.,
486 2012).

487 Fenestral porosity alternates between individual laminae, resulting in light-dark
488 layering within hand samples and thin sections. Lighter, more porous laminae are 2-5 mm
489 thick, while darker laminae are less than 2 mm thick and contain fewer fenestrae. Detrital
490 grains are present but not abundant throughout light and dark palisade laminae (Fig. 9E, G).
491 Both styles of lamination contain filamentous microfabrics visible in thin-section and SEM
492 imaging (Figs. 9D-G, 10A-E). Filaments in less-porous laminae are oriented parallel to
493 bedding planes. In contrast, filaments in porous laminae contort around curved fenestral
494 margins in various orientations.

495 Palisades in Cores C, D, and E differ in fenestral size and morphology (Fig. 9A-C).
496 Fenestrae in Cores C and E are sub-mm in scale, with sub-circular margins in cross-section
497 (Fig. 9A, C). By contrast, Core D fenestrae have elongated oval or lobate morphologies
498 between 0.5 and 5 mm in diameter (Fig. 9B, D). Elongate fenestrae are especially prevalent
499 in thicker porous layers of Core D. Long axes of such fenestrae are oriented normal to
500 bedding planes, separated by thin vertical columns of silicified microbial filaments (Fig. 9D).

501 Darker layers of Core D contain smaller, more circular fenestrae comparable to morphologies
502 observed in Cores C and E.

503 Filamentous microfossils are abundant in all fenestral palisades (Fig. 10A-C). In SEM
504 images, Cores C and D exhibit round voids lined with 5- μ m wide filaments up to 100 μ m in
505 length (Fig. 10A, B). Filaments retain equal diameter throughout their length, terminating
506 with rounded, untapered ends. Individual microfossils frequently form bridges across
507 fenestral pore spaces. While most sinters between fenestrae only preserve microfossils on
508 exterior surfaces, exquisitely preserved regions reveal columnar bundles of filaments
509 stretching across pore spaces (Fig. 10D, E). Secondary overprinting is especially prevalent in
510 Core E, where microfossils are still abundant, but rounded fenestral textures are less distinct
511 (Fig. 10C).

512 Palisades and similar microbial fabrics are globally present in hydrothermal sinter
513 deposits of various ages (Walter, 1976a; Walter et al., 1996, 1998; Jones et al., 1997, 1998,
514 2005; Campbell et al., 2001, 2015; Konhauser et al., 2004; Fernandez-Turiel et al., 2005;
515 Berelson et al., 2011; Gong et al., 2019). Variations in filament orientation are interpreted
516 differently in various locations, including seasonal growth patterns (Walter, 1976a; Jones et
517 al., 1998, 2005), pulses in hydrothermal sheet-flow (Campbell et al., 2015), and quasi-daily
518 shifts in water temperatures (Berelson et al., 2011). On steeper surfaces such as the edges of
519 sinter mounds, palisade formation is driven by cycles of wetting and drying (Gong et al.,
520 2019).

521 Palisade sinters closely resemble cyanobacterial communities living at or below 40-45
522 °C, typically interpreted as *Calothrix* (Walter, 1976a; Cady & Farmer, 1996; Campbell et al.,
523 2001, 2015), though similar textures have been noted in communities of *Fischerella*,
524 *Phormidium*, and *Leptolyngbya* (Jones et al., 2005; Pepe-Rannek et al., 2012; Gong et al.,

525 2019). In active El Tatio geyser terraces, morphological and molecular evidence show a
526 predominance of *Leptolyngbya*, and an absence of *Calothrix* or its more halotolerant sister
527 genus, *Rivularia* (Gong et al., 2019). Due to the robust structures of Calothrix-type
528 organisms, their absence is not likely due to a preservation bias, but rather interpreted as
529 absence from the characterized terraces.

530 In many sinter localities, erect palisade filaments are often tightly packed together,
531 leaving no space for mm-scale fenestrae. However, several locations contain more porous
532 fenestral palisades in distal sinter aprons and terraces, including other zones in El Tatio
533 (Jones et al., 1998; Munoz-Saez et al., 2016). One environmental exception occurs in
534 Obsidian Pool Prime in Yellowstone National Park (Berelson et al., 2011; Mata et al., 2012),
535 where fenestral textures occur in stromatolites around higher-temperature pool margins (>45
536 °C). However, Obsidian Pool Prime fenestrae have distinctly different morphologies from El
537 Tatio cores, forming sub-spherical voids less than 1 mm in diameter. Instead, fenestrae in
538 core samples more closely resemble elongate pores observed in cooler, distal settings
539 elsewhere in El Tatio and New Zealand (Jones et al., 1998; Munoz-Saez et al., 2016).

540 4.7 Tufted mat sinter

541 Tufted mat sinters are defined as filamentous microbial mat textures with sharp peaks
542 and cones irregularly separated by 1-2 mm concave depressions (Cady & Farmer, 1996;
543 Walter et al., 1996, 1998; Campbell et al., 2001; Jones et al., 2005), most notable in thin-
544 section (Fig. 11A, B). Tufted facies occur within two zones of Core E (0-103 cm, 289-367 cm
545 depth), interspersed as individual beds up to 15 cm thick. The top meter of Core E contains
546 tufted beds interspersed between palisade fabrics, while lower units briefly occur between
547 thicker units of arborescent sinters and laminated sandstone deposits. Individual tufted layers
548 are ~100 µm thick, and are difficult to discern in hand samples. Laminae laterally increase in

549 thickness to 200 μm at tufted peaks, and diminish to 50 μm in interstitial depressions.
550 Textures immediately below tufts vary between layers. Most laminae do not preserve
551 additional microbial textures below tufts (Fig. 11B), while some laminae contain relict
552 microbial filaments (Fig. 11A). Occasionally, both highly circular and elongate fenestrae are
553 preserved, bounded by thin pillars of filamentous sinter (Fig. 11C).

554 SEM imaging of tufted sinters shows exquisite preservation of microbial mat textures.
555 Three-dimensional tufts are composed entirely of microbial filaments between 2-5 μm in
556 diameter (Fig. 11D-F). Filaments form diffuse meshwork textures at the base of tufts, while
557 sharp peaks exhibit denser accumulations, reflecting differences in thickness and density
558 observed in thin-sections (Fig. 11A, B). Large rounded fenestrae are also present, bounded
559 and separated from each other by silicified filaments (Fig. 10F, G). Fenestrae are less
560 frequent in tufted facies than palisades, but are larger and more spherical.

561 Tufted facies closely resemble modern mats dominated by the cyanobacterial taxa
562 *Phormidium* and *Leptolyngbya* (Walter, 1976a; Campbell et al., 2001; Fernandez-Turiel et
563 al., 2005; Lau et al., 2005; Bosak et al., 2012; Reyes et al., 2013, Bradley et al., 2017). Many
564 tufted mats grow between 45 and 60 $^{\circ}\text{C}$ (Walter, 1976a; Campbell et al., 2001; Lau et al.,
565 2005; Bradley et al., 2017), though tufted *Leptolyngbya* mats are also reported between 17
566 and 45 $^{\circ}\text{C}$ (Bosak et al., 2012; Reyes et al., 2013). Large rounded fenestrae below tufts are
567 also distinct from fenestral palisade textures, more closely resembling spherical gas bubbles
568 preserved within tufted hydrothermal mats (Lynne & Campbell, 2003; Bosak et al., 2008,
569 2010; Mata et al., 2012). Tufted facies are interspersed with palisade fabrics in the top meter
570 of Core E, with changes in mat morphology potentially facilitated by changes in water
571 temperatures or hydrochemistry.

572 4.8 Massive sinter

573 Massive sinters exhibit either obscured or no primary features such as laminae,
574 bedding, detrital grains, or microfossils (Fernandez-Turiel et al., 2005; Walter et al., 1996).
575 Massive sinter facies occur at various depths in Cores A, D, and E. The lowest meter of Core
576 A consists of green and gray low-porosity sinter with pale yellow and orange minerals along
577 secondary fissures (Fig. 12A). While massive Core A sinter contains vertically oriented
578 lineations in hand sample similar to inclined sinter crusts, all traces of vertical fabric are
579 obscured in thin section (Fig. 12B). Sinters at the base of Core D (72 – 97 cm) have similarly
580 low porosity, and lack bedding, vertical lineation, or clear laminae in hand sample or thin
581 section (Fig. 12C). Massive sinters occur in 5-cm thick beds at various depths in Core E (130-
582 135 cm, 320-325 cm, 330-335 cm). Core E facies are lighter in color than in Cores A and D,
583 and have gradational contacts with particulate and arborescent sinters. Microscopy of massive
584 sinters and adjacent facies also shows extensive growth of secondary silica phases within
585 pore spaces (Fig. 12D, E).

586 Massive sinter fabrics are present in modern and ancient sinter deposits (Walter,
587 1976a; Walter et al., 1996; Guidry & Chafetz 2003a). The lack of primary fabrics in hand
588 samples and thin-sections implies secondary alteration, such as recrystallization and
589 overgrowth of original sinter deposits, most likely by high temperature silica-rich subsurface
590 fluids. Adjacent primary facies in Cores A and E support a high-temperature alteration
591 hypothesis. Massive sinter in Core A is overlain by more than 2 meters of inclined and planar
592 sinter crusts, representing the sides and floors of vent-proximal pools respectively. In Core E,
593 massive sinters are associated with particulate sinters, also interpreted as subaqueous deposits
594 near secondary hydrothermal vents distal to Mound 502. In contrast, facies at the base of
595 Core D are not associated with high-temperature pool or vent facies. Based on high-
596 temperature associations elsewhere, massive sinters in Core D are interpreted as former

597 microbial mat sinters which were secondarily altered by subsurface fluid flow and
598 recrystallization.

599 4.9 Laminated sandstone and sinter

600 Laminated sandstone and sinter (LSS) deposits are defined as interbedded siliciclastic
601 and sinter layers between 0.1 and 5 cm thick. The LSS facies is only observed in the lower
602 2.5 meters of Core E. Sandstones are medium- to coarse-grained, and are composed of sub-
603 rounded to sub-angular volcanoclastic grains and occasionally detrital sinter pieces (Fig. 13A,
604 C, E). No evidence of cross-bedding is observed in sandstone beds, while sinters contain
605 millimeter- scale planar laminae lacking fenestrae or filaments. Laminated sinters often have
606 irregular, unconformable contacts with overlying sandstones.

607 The ratio of sinter to sandstone in LSS facies varies in different areas of Core E.
608 Below 4 m depth, sandstone layers are thicker, containing isolated, centimeter-scale sinter
609 deposits. Unconformities separate lower LSS facies from intervening massive sandstones
610 lacking sinter in Core E (described below). LSS facies above 4 meters in Core E are
611 dominated by planar sinter, punctuated with sub-centimeter sandy layers. Upper LSS facies in
612 Core E are typically less than 5 cm thick and are often under- or overlain by tufted microbial
613 textures, distinguished by filamentous fabrics.

614 LSS facies in Core E are interpreted as distal sheet-flow deposits, with sinter layers
615 representing relative lulls in siliciclastic input. Beds below 4 m depth indicate higher-energy
616 environments where clastic sedimentation prevented persistent sinter deposition. In contrast,
617 LSS facies above 4 m depth exhibit minor pulses of siliciclastic deposition in a calmer, sinter-
618 dominated regime. In both zones, sinter deposits were consistently eroded by subsequent
619 sandstone deposition, forming irregular upper contacts.

620 4.10 Massive sandstone

621 Massive sandstones contain no bedding textures or interstitial sinter deposits, and are
622 only found in the lowest meter of Core E. Similar to LSS facies, sands are medium to coarse-
623 grained, and are composed of sub-rounded and sub-angular volcanoclastic grains (Fig. 13B,
624 D, F). The majority of siliciclastic material is composed of volcanic lithic fragments,
625 feldspar, mica and minor portions of redeposited sinter material. The sandstone is notably
626 free of detrital quartz, but largely cemented by siliceous material such as opal and chalcedony
627 (Fig. 13F).

628 Massive sandstone units have unconformable boundaries with adjacent sinter and
629 sandstone beds, and occasionally contain unconformities within otherwise homogenous units.
630 One notable unconformity occurs at 538 cm depth in Core E. Sandstones containing 1-2 cm
631 wide matrix-supported rounded green clasts are conformably overlain by planar beds of sands
632 fining upward from coarse to medium-grained. Both units are eroded by an unconformably
633 overlying homogenous, coarse-grained sandstone. Vitreous brown cements follow the
634 unconformity, cross-cutting into over- and underlying sediments (Fig. 13B). No other
635 sedimentary structures are apparent in massive sandstones.

636 Massive sandstones are interpreted as debris flows with little to no hydrothermal
637 influence. The only hydrothermal silica precipitation present (538 cm) occurs as secondary
638 precipitation following unconformities and cross-cutting primary lithologies in a narrow
639 zone.

640 **5. Core geochemistry**

641 Sinters in Cores A, C, D, and the upper 4 meters of Core E are largely composed of
642 siliceous material with more than 94 wt.% of SiO₂ (Fig. 14). Cores closer to the vent source
643 (A, C, D) are especially enriched in SiO₂, with all measured values above 98.5 wt.%. SiO₂
644 concentrations in Cores A and C decrease towards surface deposits, with higher values in

645 sinter crusts and columnar sinters, and lower values in mat and particulate sinter facies. Core
646 D shows little variability between massive sinters and palisade mat facies, while sandstones
647 at the base of Core E are relatively SiO₂-depleted (70-85 wt.%), with a sharp increase to >90
648 wt.% at 361 cm depth. SiO₂ concentrations in Core E increase to ~99 wt.% in arborescent,
649 particulate and massive sinters between 1-2 m depth before slightly decreasing within the
650 uppermost meter of mat facies, in a similar fashion to Cores A and C (Fig. 14).

651 Other major elements are anti-correlated to SiO₂ concentrations in all cores (Fig. 14,
652 Supplementary Information). For example, Al₂O₃, Fe₂O₃, TiO₂, etc. increase upwards in
653 Cores A and C as SiO₂ decreases. In Core E, sandstones below 4 m depth have different
654 geochemical signatures than overlying sinters. Many elements are several orders of
655 magnitude more abundant in Core E sandstone deposits (Fig. 14), including Al₂O₃
656 concentrations above 10 wt.%. In Core E sinter facies above 4 m depth, elemental
657 abundances reach minimum values at ~2 m (e.g., Al₂O₃ concentrations at 0.1 wt.%) before
658 slowly increasing within the top meter of mat sinter facies. While certain elements are
659 elevated in individual layers (e.g., P₂O₅ in Core E, 103 cm), larger shifts are consistent
660 between elements in all cores.

661 Most trace elements exhibit similar stratigraphic trends to major elements in all four
662 cores. For example Li, Cs, Rb, Se, U, Cr, and many other elements are enriched in Core E
663 sandstones below 4 m depth, and depleted in overlying sinters, with slightly higher
664 concentrations of Li and Cs (Supplementary Information). In rare cases, certain elements are
665 enriched in sinters compared with clastic deposits. For example, Sb concentrations increase in
666 Core E sinters above 4 m, and represent the most abundant trace element in all sinter samples
667 (Supplementary Information). As is slightly more abundant in sinter than sandstone facies,
668 with elevated concentrations in the top 0.5 m of all cores (Supplementary Information).

669 Rare earth element (REE) core profiles are nearly identical to most major and trace
670 elements, with strong anti-correlation to SiO₂ concentration (Fig. 14, 15). When REE
671 concentrations of individual samples are normalized to clastic deposits of Mound 502, neither
672 light nor heavy rare earth elements (LREE and HREE) are enriched (Fig. 15, Supplementary
673 Information). Slight HREE enrichment occurs when core samples are normalized to
674 chondritic meteorites and post-Archean Australian shale (PAAS) (Supplementary
675 Information).

676 Total REE abundance increases towards the tops of Cores A and C, with a slight
677 decrease around 1.2 m depth in Core A (Fig. 15). Core D has consistent REE concentrations
678 over a limited sampling distribution. As with major elements, REE concentration in Core E
679 varies greatly based on depth and facies. Sandstones below 4 m depth have concentrations
680 one to two orders of magnitude higher than overlying sinters, forming a sample population
681 distinct from facies in any other core (Fig. 15). REE abundance drops drastically by
682 approximately one order of magnitude at 4 m depth, when sinter becomes more abundant
683 relative to sandstone. Between ca. 3.6 and 3.0 m depth, REE abundance decreases steadily by
684 one order of magnitude where sandstone bearing beds become less abundant (Fig. 15).
685 Arborescent and particulate sinters between 1 and 2 meters contain the lowest REE
686 concentrations in Core E, with values gradually increasing in overlying mat facies.

687 Different core facies also vary in REE patterns between individual samples (Fig. 15).
688 When normalized to values of clastic deposits within Mound 502, REE patterns are
689 essentially flat in facies with elevated concentrations, such as LSS and various microbial mat
690 facies. In samples with depleted REE, certain elements are relatively enriched, including La,
691 Eu, Y, and Lu (Fig. 15). Enrichment of such elements typically increases as total REE
692 concentration decreases, and is most prevalent in sinter crust, particulate, and columnar sinter
693 facies. For example, positive Eu anomalies (enrichments in Eu relative to standards) are

694 highest in sinter crust and columnar facies in Cores A and C, respectively (Fig. 15, 16). The
695 highest positive Eu anomalies in Core E occur in particulate sinters between 1.5 and 2.0 m
696 depth, with all other facies exhibiting slightly negative or no Eu anomalies (Fig. 16).

697 **6. Discussion: Core depositional histories**

698 Depositional interpretations based on core sedimentology are well-supported by
699 geochemical data, especially regarding proxies for siliciclastic material relative to mineral
700 precipitation in hydrothermal fluids. For example, Al-concentration in chemical sediments is
701 frequently used as a proxy for siliciclastic components, such as clay minerals and feldspar
702 (Calvert, 1976; Tribovillard et al., 2006). The high concentration of Al₂O₃ and other elements
703 in sandstone beds compared with sinters confirms such interpretations for El Tatio systems
704 (Fig. 14). Therefore, subtle differences in Al₂O₃ and associated elements within sinters can
705 determine relative siliciclastic input and corroborate environmental interpretations based on
706 sedimentary structures. Similarly, positive Eu anomalies are often applied as a proxy to
707 estimate the relative influence of hydrothermal fluids on lithologies (Michard, 1989; Lewis et
708 al., 1997; Feng et al., 2014). On broad lateral scales, positive Eu anomalies in El Tatio cores
709 are larger with proximity to the vent source (Fig. 16). Such trends confirm the utility of Eu
710 anomalies within cores to determine relative fluctuations in hydrothermal activity.

711 Al and Eu proxies appear to be broadly connected in El Tatio cores, with samples
712 forming a spectrum between two geochemical end-member proxies. Facies with higher Al₂O₃
713 concentrations such as sandstones and microbial mat sinters have little to no Eu anomaly,
714 while samples with high positive Eu anomalies (sinter crust, particulate, columnar sinters)
715 have relatively low Al₂O₃ abundance (Figs. 14-16). This dichotomy is most likely due to
716 differences in REE abundance between hydrothermal sinters and detrital sandstones.
717 Sandstones have the highest \sum REE in any core facies, with values orders of magnitude above

718 associated sinters (Fig. 15). The abundance of REE in sandstone implies that even a relatively
719 minor siliciclastic component in sinter can overpower the REE signature of depleted sinters.
720 Separating Σ REE into LREE or HREE reveals no enrichment of either group in clastic-
721 normalized samples (Fig. 15), while normalizations to other standards reveal only minor
722 enrichments in HREE (Supplementary Information).

723 Therefore, in El Tatio cores, positive Eu anomalies are conservatively interpreted as
724 the relative influence of hydrothermal precipitation compared to siliciclastic influx. The
725 dichotomy between Al₂O₃ concentrations and Eu anomalies persists across all four cores, and
726 matches interpretations based on sedimentary and geochemical evidence. Distal environments
727 have higher abundances of siliciclastic material relative to hydrothermal silica precipitation
728 than vent-proximal zones. Microbial mats in particular are noted for trapping detrital material
729 with filamentous textures and adhesive extracellular polymers (Gebelein, 1969; Riding, 2000;
730 Frantz et al., 2015).

731 Trace element proxies support previous observations of El Tatio water and sinter
732 chemistry. Two primary types of hydrothermal fluids exist in the El Tatio geyser field: alkali-
733 chloride and acid-sulfate (Giggenbach, 1978; Cortecchi et al., 2005). Alkali-chloride waters in
734 El Tatio are enriched in As, Cs, and Li, while acid-sulfate waters are only enriched in B
735 compared to meteoric sources (Cortecchi et al., 2005). While B was not examined in this
736 study, cored sinters are slightly enriched in Li, Cs, and As compared with most other trace
737 elements (Supplementary Information), indicating the presence of alkali-chloride waters
738 similar to modern Upper Basin environments (Giggenbach, 1978; Cortecchi et al., 2005). In
739 addition, elevated concentrations of Sb in Mound 502 sinters are consistent with previous
740 sinter studies at El Tatio and other Chilean hydrothermal zones (Landrum et al., 2009;
741 Sanchez-Yanez et al., 2017). Combined sedimentary and geochemical data therefore provide
742 depositional histories for each core and form a local model for sinter mound growth (Fig. 17).

743 6.1 Core A

744 Core A was drilled on top of Mound 502, along the inner rim of the steep-walled
745 inner pit. Core A is the first sedimentary core described from vent-proximal sinter deposits,
746 and it provides several insights into high-temperature facies which corroborate observations
747 of active modern hydrothermal systems.

748 The lowermost meter of Core A is massive sinter, showing no primary texture in hand
749 sample or thin section (Fig. 12A, B). Massive facies contain low Al_2O_3 concentrations and a
750 significant positive Eu anomaly (Fig. 14-16), indicating extensive silica precipitation in
751 hydrothermal waters with low siliciclastic input. Such precipitation can occur during primary
752 vent-proximal deposition or secondary subsurface alteration. While specific depositional
753 features are absent, Al_2O_3 concentrations in massive sinters as low as 0.16 wt.% indicating
754 very little to no siliciclastic input. Therefore, we interpret massive Core A sinters as vent-
755 proximal hydrothermal deposits secondarily altered by subsurface flow, consistent with
756 previous studies of massive sinter (Walter, 1976b; Walter et al., 1996; Guidry & Chafetz
757 2003a).

758 Inclined sinter crust is only observed in Core A, forming a nearly continuous unit ~2
759 meters thick. Depositional angles and bed thicknesses correspond to the surficial topography
760 of Mound 502's inner pit, which descends ~1.5 meters below rim margins (Fig. 1D). Sint-
761 ers from 121 cm depth contain the lowest measured concentrations of Al_2O_3 and $\sum\text{REE}$ in all
762 core samples, as well as the highest percentages of SiO_2 and a strongly pronounced positive
763 Eu anomaly (Fig. 14-16). Geochemistry and sinter topography support the presence of a high-
764 temperature pool sourced by subsurface hydrothermal fluids (Fig. 17), currently represented
765 by the inner pit of the extinct mound. Microscopic surface-normal silica fabrics support a
766 hydrothermal pool interpretation for sinter crusts (Figs. 3, 4), closely resembling fabrics

767 observed in subaqueous conditions over 73 °C in Yellowstone pools (Lowe & Braunstein,
768 2003). While microbial textures are occasionally observed in SEM, the prevailing surface-
769 normal fabrics more closely resemble ambiguously biogenic “shrub columns” (Jones &
770 Renaut, 2003) or abiogenic “dendritic sinter fabrics” (Lowe & Braunstein, 2003), forming
771 when silica precipitation outpaces microbial growth.

772 Positive Eu anomalies and SiO₂ percentages decrease further up section while Al₂O₃
773 concentrations increase, indicating an increase in siliciclastic input relative to sinter
774 precipitation. An angular unconformity separates inclined sinter crusts from overlying planar
775 facies. Planar sinter crusts match the flat-topped topography of Mound 502’s inner rims,
776 potentially signifying a shift from silica accretion within steep-sided pools to vertical build-
777 up of pool-side margins (Fig. 17). No textural features in planar crusts indicate the presence
778 of microbial mats or indicate biological mediation of silica precipitation. The top ten
779 centimeters of Core A are particulate sinter with abundant siliciclastic grains visible in hand
780 sample, signifying a decrease in relative hydrothermal activity compared to clastic
781 sedimentation (Fig. 17).

782 Vent-proximal sinter facies are frequently studied in active hydrothermal systems,
783 exhibiting a diverse range of environments including geysers and pools with varying degrees
784 of boiling, surging, and overflow (Braunstein & Lowe, 2001; Guidry & Chafetz, 2003; Lowe
785 & Braunstein, 2003; Hinman & Walter, 2005). However, the stratigraphic context of high-
786 temperature vents is less well-understood. The most detailed vent-proximal stratigraphy
787 comes from Devonian sinters in the Drummond Basin of Australia (Walter et al., 1996,
788 1998). Extensive stratigraphic columns of Drummond exposures revealed several high-
789 temperature facies: massive and brecciated sinters interpreted as sub-surface or altered vent
790 zones, as well as thin flat-bedded and “diffusely layered” sinters interpreted as pool-bottom
791 facies (Walter et al., 1996). Vent-proximal sinters are spread out in areas over > 100 m² in the

792 Drummond Basin, and are frequently intercalated with lower-temperature mat facies,
793 reaching no more than 50 cm thickness. In contrast, no sinter crusts or particulate facies are
794 described. The differences between Core A and Drummond strata highlight the diversity in
795 vent-proximal facies, and the need for further stratigraphic investigation.

796 6.2 Core C

797 Core C was drilled five meters to the east and 1.5 meters below the top of Core A, on
798 the outer margin of Mound 502. The lower meter of Core C is a nearly continuous unit of
799 distinctive columnar sinter (Figs. 6, 7). Despite the major morphological differences between
800 columnar sinters in Core C and sinter crusts in Core A, both have similar chemical signatures.
801 Al_2O_3 , ΣREE , and many other elements increase in roughly equivalent concentrations up-
802 section, while SiO_2 shows a steady decrease by ~1 wt.% (Figs. 14, 15). Positive Eu anomalies
803 near the base of Core C are the highest in any core, including sinter crust facies in Core A
804 (Fig. 16). The geochemical signatures indicate a high-temperature depositional environment
805 with little to no siliciclastic input (Fig. 17). Such interpretations are strongly supported by
806 current observations of columnar and spicular sinter formation in splash and spray zones
807 close to hydrothermal vents and pools (Walter, 1976b; Jones et al., 1997, 1998; Braunstein &
808 Lowe, 2001; Lowe & Braunstein, 2003; Jones & Renaut 2003; Guidry & Chafetz, 2003a).

809 Despite the general consistency of columnar morphologies in Core C, several hiatuses
810 and depositional changes are present. The transition from mm-scale spicules to cm-scale
811 columns at the base of Core C is separated by a brief section of missing core. The sequence
812 most likely represents a depositional hiatus between splash zones immediately adjacent to a
813 hydrothermal vent, and surge-flow zones forming larger columns further away from
814 hydrothermal activity. At 1.5 m depth, laminated arborescent sinter immediately overlies
815 horizontal columns, with no evidence of intervening erosional activity (Fig. 6E). Overlying

816 columnar sinters maintain similar widths and branching morphologies, but form at increasing
817 depositional angles over the next 70 centimeters (Fig. 6C, D). The brief interval of
818 arborescent sinter formation indicates an outflow-dominated period with active microbial
819 communities colonizing the sinter (Fig. 6E).

820 Microbial cells and their growth textures are either extremely altered or absent in
821 columnar sinter facies (Fig. 7F, G), as opposed to abundant and well-preserved textures in
822 palisade mat fabrics overlying columnar sinters (Figs. 9, 10). Columnar facies in various
823 locations appear to form with or without the presence of microbial communities (Walter,
824 1976b; Jones et al., 1997, 1998; Braunstein & Lowe, 2001; Jones & Renaut 2003; Lowe &
825 Braunstein, 2003). As with sinter crusts, the paucity of microbial textures in well-preserved
826 columns does not preclude the earlier presence of microbes, but supports the dominance of
827 abiogenic processes of sinter growth.

828 Palisade mats in the top meter of Core C contain chemical signatures with no Eu
829 anomalies and elevated siliciclastic signatures (Fig. 14-16). Palisade mats are typically
830 interpreted as forming at water temperatures at or below 40-45 °C (Walter, 1976a; Campbell
831 et al., 2015; Gong et al., 2019), though the lack of an Eu-anomaly combined with higher
832 concentrations of Al₂O₃ in Core C palisades is more indicative of increased detritus than
833 relative temperature. Arborescent sinter appears as thin interstitial layers between palisade-
834 mat sinters, and separate palisade mat facies from underlying columnar facies at 0.5 m depth.
835 No chemical data was collected from arborescent sinters, but visual observations of detrital
836 grains, and comparisons with “siliceous shrubs” from Yellowstone imply outflow channels
837 with potentially higher discharge than palisade mat facies. The top half meter of Core C is
838 interpreted as a shift from a hot splashing zone to more clastic-dominated outflow
839 environments with occasional pulses of hydrothermal activity (Fig. 17).

840 As with vent-proximal facies in Core A, columnar and spicular sinters of Core C are
841 well-described from surface deposits in active and recently inactive hydrothermal settings
842 (Walter, 1976b; Jones et al., 1997, 1998; Braunstein & Lowe, 2001; Lowe & Braunstein,
843 2003; Jones & Renaut 2003; Guidry & Chafetz, 2003a), but this study is the first to observe
844 columnar sinters in drill core. Similar laminated columns (≤ 20 cm long) were observed in
845 stratigraphic studies of Devonian sinters from the Drummond Basin in Australia (Walter et
846 al., 1996, 1998), but poor exposure prevented correlation to adjacent facies. In contrast, the
847 transition from columns to palisades in Core C from records changing sinter deposition over
848 time, from splash and surge-dominated forces to extensive microbial preservation.

849 6.3 Core D

850 Core D was drilled 6 m southeast of Core A, starting at a roughly equivalent elevation
851 to Core C (Fig. 1E). Unlike the vertical cores of A, C, and E, D was drilled at a 27° angle
852 from vertical alignment along the outer edge of Mound 502, with the drilling axis pointed
853 towards the NW. Core D is the shortest core, and contains the lowest diversity in sedimentary
854 facies and geochemical signatures.

855 Massive sinters at the base of Core D are not visibly associated with high-temperature
856 facies, such as sinter crusts, particulate, or columnar sinters, although such deposits could
857 underlie the core base (Fig. 2). However, the single chemistry data point within massive
858 sinters is nearly identical to overlying palisade mat facies, with a minor positive Eu anomaly
859 and a moderate siliciclastic signature (Fig. 14-16). Massive facies in Core D are therefore
860 interpreted as relatively distal sinters secondarily altered by subsurface fluid flow (Fig. 17).

861 Fenestral palisade fabrics at the top of Core D are nearly identical to facies observed
862 in Cores C and E (Fig. 9B), preserving microbial communities visible in thin-section and
863 SEM imaging. Chemical analyses reveal slight differences between palisade sinters of

864 different cores, with Core D palisades bearing slightly higher positive Eu anomalies and
865 lower siliciclastic signatures such as Al_2O_3 than Cores C and E (Fig. 14-16). While it is
866 possible that Core D mats were deposited under warmer conditions than other palisade
867 textures, the relative lack of data keeps temperature estimates conservatively around 40-45 °C
868 as described in previous surficial and stratigraphic studies (Walter, 1976a, 1996, 1998; Jones
869 et al., 1998; Lynne et al., 2008; Campbell et al., 2001, 2015, 2020; Gong et al., 2019). In
870 either case, the top of Core D matches broad trends observed in surficial facies of all cores,
871 featuring strata deposited in cooler water temperatures with higher amounts of siliciclastic
872 material before termination of hydrothermal activity (Fig. 17).

873 6.4 Core E

874 Core E was drilled into a sinter debris apron 30 meters southeast of Core A, and 15
875 meters southeast of Mound 502 itself. The core begins approximately four meters below the
876 top of Core A, and is the deepest of the four cores, reaching 560 cm depth. Core E records the
877 highest diversity of sedimentary facies and geochemical signatures, including higher-
878 temperature columnar sinters, lower-temperature mat facies, and clastic sandstone beds.

879 Core E is the only core examined to record non-hydrothermal sediments deposited
880 prior to extensive sinter formation, appearing as sandstones below 4 m depth (Fig. 13). Most
881 sandstones present are massive and are likely the result of debris flows from surrounding
882 hillsides into valleys to the south (Fig. 1B). However, minor layers of laminated sinter occur
883 throughout sandstone beds, indicating an influence from hydrothermal activity. The high
884 abundance of SiO_2 throughout the sandstone succession in combination with the lack of
885 detrital quartz (Fig. 13F) indicates that this hydrothermal influence persisted throughout the
886 sandstone deposits. Minor variations in the SiO_2 concentration are most likely due to
887 differences in the amount of silica cement.

888 Layers between 2.9 and 3.7 m depth in Core E form the most diverse assemblage of
889 facies in any core measured in this study. Six different facies occur within 80 centimeters,
890 with no single unit exceeding ten centimeters in thickness (Fig. 2). Such disparate lithologies
891 imply relatively rapid environmental shifts on Mound 502's outer distal apron, including
892 hydrothermal vent sources separate from the mound's central pool. Two sedimentary cycles
893 are noted, each shifting from subaqueous or pool-side facies to outflow environments. The
894 lower cycle (3.4-3.7 m depth) features a thin layer of small columnar sinter (Figs. 6F, 7B)
895 overlain by tufted mat sinters and laminated sandstones. The upper cycle (2.9-3.4 m depth)
896 contains massive and arborescent sinters overlain by mixed tufted mats and sandstones. In
897 other cores, columnar and massive sinters typically contain elevated positive Eu anomalies
898 indicative of extensive hydrothermal silica precipitation. However, between 2.9 and 3.7 m
899 depth in Core E the relative concentration of Eu is similar to that of the underlying sandstone,
900 but increases only slightly up section (Fig. 15). By contrast, concentrations of Al₂O₃ and
901 other elements decrease by orders of magnitude (Fig. 14). Together, petrographic-, Eu-, and
902 Al₂O₃-data depict a location experiencing locally rapid depositional shifts, with an underlying
903 trend of increasing hydrothermal sinter precipitation and decreasing siliciclastic input (Fig.
904 16).

905 The only section of Core E recovered between 2-3 meters is a 30-centimeter section
906 (2.2-2.5 m depth) of palisade mat fabrics underlain by thin beds of massive sinter. The
907 abrupt transition from massive to palisade sinters is similar to Core D, with more detailed
908 geochemical data sampling revealing distinct shifts in temperature and siliciclastic influence.
909 Positive Eu anomalies in massive sinters are equivalent to core facies at 3 m depth (Fig. 16).
910 In contrast, immediately overlying palisade mat sinters contain no Eu anomaly, but quickly
911 return over the subsequent twenty centimeters. Al₂O₃ and other siliciclastic constituents are
912 anti-correlated to relative Eu concentration, increasing sharply in lower palisade sinters and

913 decreasing up-section (Fig. 14, 15). Geochemical trends indicate hydrothermal environments
914 with a low siliciclastic component, succeeded by palisade mats potentially growing at ~40 °C
915 (Walter, 1976a; Campbell et al., 2015; Gong et al., 2019), though potentially persisting in
916 higher temperatures. No change in mat architecture is associated with changes in relative Eu
917 concentrations.

918 Between 1.0 and 2.0 m depth, Core E is entirely composed of particulate, massive,
919 and arborescent sinters. Particulate and massive facies are frequently interpreted as
920 subaqueous deposits in high-temperature pools (Walter, 1976b; Braunstein & Lowe, 2001;
921 Lowe & Braunstein, 2003; Hinman & Walter, 2005), and Core E geochemistry supports
922 similar conclusions. The highest positive Eu anomalies in Core E occur between 1.5 to 2.0 m
923 depth, similar to values observed in sinter crusts and columnar sinters in Cores A and C (Fig.
924 16). As in more vent-proximal cores, positive Eu anomalies gradually decrease up-section,
925 eventually approaching similar relative Eu ratios observed in lower Core E sinters.
926 Concentrations of Al₂O₃, Σ REE and other siliciclastic proxies reach minimum values in Core
927 E between 1.0 and 2.0 meters, and remain consistently low throughout dendritic and
928 associated sinters (Fig. 14, 16). Unlike inclined sinter crusts in Core A and columnar sinter in
929 Core C, high-temperature facies in Core E appear to have precipitated on flat-bottomed pools
930 on distal aprons, separate from the elevated sinter mound (Fig. 17). Such pools are still active
931 several meters away from Core E, supporting their potential presence during Core E
932 deposition. The presence of arborescent sinter between particulate facies most likely
933 represents a shift from pool deposits to outflow channels with microbial communities (Guidry
934 & Chafetz, 2003c).

935 The top meter of Core E contains the thickest continuous package of microbial facies
936 in any core examined. Palisades comprise the majority of preserved mats, with nearly
937 identical fenestral morphologies to the upper beds of Cores C and D (Fig. 9C). The lower

938 boundary between particulate sinters and overlying palisades is associated with an immediate
939 loss of the positive Eu anomaly, followed by fluctuations around the average for clastic
940 deposition around Mound 502 (Fig. 16). Unlike Cores C and D, palisades in Core E contain
941 distinct interstitial layers of tufted microbial sinters. Tufted sinters are typically assigned to
942 higher-temperature communities (45-60 °C) than palisade mats (< 45 °C), based on
943 morphologies of modern hot spring communities (Walter, 1976a; Campbell et al., 2001;
944 Lynne & Campbell, 2003; Fernandez-Turiel et al., 2005). No clear difference in REE
945 signatures exists between tufted and palisade mat sinters in El Tatio cores, most likely due to
946 the entrapment of siliciclastic material on sticky microbial surfaces.

947 The diversity of facies observed in Core E resembles previous stratigraphic studies
948 from distal sinter deposits around the world (Trewin, 1993; Walter et al., 1996, 1998; Jones et
949 al., 1998; Campbell et al., 2001, 2020; Guidry & Chafetz, 2003b; Lynne et al., 2008).
950 Palisade and tufted microbial mats are ubiquitous facies in cooler environments (Walter et al.,
951 1996, 1998; Jones et al., 1998; Campbell et al., 2001, 2020), while arborescent textures are
952 described less frequently (Walter et al., 1996, 1998; Campbell et al., 2001). The frequent
953 interbedding between microbial facies and more massive deposits in Core E (1-2, 3-3.5 m
954 depth) is also a common feature of distal sinter stratigraphy (Walter et al., 1996; Campbell et
955 al., 2001; Lynne et al., 2008). In contrast, Core E lacks extensive preservation of metazoan
956 communities such as plants or diatoms, which are often observed in distal sinters (Trewin,
957 1993; Walter et al., 1996, 1998; Campbell et al., 2001, 2020; Guidry & Chafetz, 2003b).
958 While the climate at El Tatio is more extreme than many other sinter localities, scattered
959 areas with plant life are present. However, areas selected for coring in Mound 502 lack
960 extensive vegetation, and the absence of plant remains suggests similar environmental
961 conditions during sinter deposition. Core stratigraphy from Mound 502 therefore presents a
962 closer environmental analogue to ancient sinters on Earth or Mars than other distal deposits.

963 6.5 Sinter ages

964 Five ^{14}C ages were measured from Cores C and E (Table 1, Fig. 2), ranging between
965 6,907 and 13,337 y. cal. BP. Core E contains a consistent progression from 13,337 to 10,028
966 y. cal. BP, underlain by an age of 6,907 y. cal. BP at 347 cm depth (Table 1, Fig. 2). Similar
967 irregularities have been noted in previous sinter cores (Lynne et al., 2008; Lowenstern et al.,
968 2016), and have several potential explanations. The presence of younger dates below a series
969 of consistently older data points can be explained by transport of younger material into
970 deeper sediments by subsurface infiltration through fractures and pores (Lynne et al., 2008;
971 Lynne, 2012). In support of this hypothesis, various core depths contain fracture zones
972 associated with secondary mineral phases, including massive sandstones deeper in Core E
973 (Fig. 13B). While no specific fractures are apparent at 347 cm, the sinter bed is fairly porous,
974 and is ten centimeters below a massive, altered sinter (Fig. 2). We propose that younger
975 material permeated through fissures and pores in older sinters via secondary fluid flow,
976 producing an anomalous radiocarbon date thousands of years younger than three succeeding
977 deposits.

978 The four remaining ^{14}C ages (one in Core C, three in Core E) record more than three
979 thousand years of sinter deposition in Mound 502 (Figs. 2, 17, Table 1). A transition from
980 clastic sandstones to siliceous sinter occurred more than 13,337 y. cal. BP in distal debris
981 aprons (Core E, 310.5 cm depth). Between 13,337 and 10,931 y. cal. BP (Core E, 33 cm
982 depth), Core E experienced shifts between various sinter facies, including microbial mats and
983 detrital-poor vent proximal zones (Figs. 2, 17). After 10,931 y. cal. BP, palisade mat sinters
984 dominate deposition in Cores C and E (and presumably D) until at least 10,028 y. cal. BP
985 (Core E, 6 cm depth). The ages from Mound 502 supports previous timelines of hydrothermal
986 activity in El Tatio, which initiated at least 27 ka, and persisted until the present day without
987 major hiatuses (Slagter et al., 2019, Munoz-Saez et al., 2020).

988 The ^{14}C ages from Mound 502 cores are thousands of years older than surficial
989 deposits from the same location (Slagter et al., 2019). Four data points collected on a transect
990 between Cores A and C (Fig. 17B, C) record a period of deposition between 2,241 and 197 y.
991 cal. BP (Slagter et al., 2019). The age difference between the two datasets is interpreted as a
992 depositional hiatus separating two phases of sinter deposition. The first phase formed the
993 majority of deposits in Cores C and E between ~14,000 and 10,000 y. cal. BP. After ~7,000
994 years, surficial hydrothermal activity resumed, forming surficial sinters between ~2,500 and
995 200 y. cal. BP (Slagter et al., 2019). Depositional hiatuses of similar length have previously
996 been reported from hydrothermal zones in Yellowstone and New Zealand (Foley, 2006;
997 Drake et al., 2014; Hurwitz & Lowenstern, 2014), and are most likely due to changes in
998 subsurface hydrothermal systems.

999 The age of Core A remains uncertain relative to other cores. One unconformity exists
1000 in Core A at 35 cm depth, indicating a prolonged period of hydrothermal activity, followed
1001 by a hiatus and minor subsequent sinter deposition. The current set of ^{14}C ages from surficial
1002 and core samples presents two potential timelines (Fig. 17A). In one scenario, most of Core A
1003 was deposited during the initial period of hydrothermal activity, forming an elevated sinter
1004 mound more than 10,000 y. cal. BP followed by a resurgence of activity after 2,000 y. cal.
1005 BP. Alternatively, the inner mound and Core A could have formed after 2,000 y. cal. BP,
1006 with Cores C, D, and E representing relatively low-relief mound and outflow deposits (Fig.
1007 17A). In either case, depositional interpretations for all cores remain the same, with timing
1008 more well-constrained between more distal depositional facies.

1009 **7. Conclusions**

1010 The four cores examined provide the first complete stratigraphic transect of a recent
1011 sinter mound from vent to outflow. Lateral trends of hydrothermal deposits record detailed

1012 transitions in chemistry, sedimentology and biology over broad areas, while vertical profiles
1013 provide similar records for local environments over geologic time. Initial depositional
1014 interpretations from sedimentology and petrography match geochemical data from the same
1015 facies. Combining all approaches within a single location provides a detailed environmental
1016 history that includes the initiation, accretion, and eventual cessation of sinter deposition
1017 within a single mound.

1018 Hydrothermal activity started in the western margins of El Tatio's upper basin prior to
1019 13,337 y. cal. BP. Sinter deposits formed on underlying glacial sandstones, intermittently at
1020 first, then forming a wide diversity of biogenic and abiogenic textures. Deposits closest to the
1021 original hydrothermal vent are dominated by subaqueous precipitation on the steep margins
1022 of a high-temperature pool. Several meters away, extensive columnar sinter formed on outer
1023 pool margins experiencing alternate periods of pool overflow and subaerial evaporation. Less
1024 extensive pools with subaqueous and columnar facies also formed in distal aprons.

1025 Microfossils are rare to absent in high-temperature facies (>70 °C), with silica precipitation
1026 most likely driven by physico-chemical factors such as silica concentrations, cooling, and
1027 evaporation. On the other hand, extensive microbial deposits are present in cooler, distal
1028 environments with more siliciclastic input. Palisade, tufted, and arborescent sinters indicate
1029 the presence of multiple distinct microbial communities, potentially fluctuating with water
1030 temperature and flow dynamics. Mat deposits and detrital-rich facies eventually replace high-
1031 temperature sinters in the upper sections of all cores, signifying cooling and eventual
1032 termination of local deposition no more than 10,232 y. cal. BP. In other areas of the same
1033 mound, hydrothermal activity persisted for several thousand years (Slagter et al., 2019).

1034 As seen within the El Tatio cores, sinter stratigraphy is complex, and varies both
1035 laterally and vertically on relatively fine scales. The deposits examined represent only one
1036 type of hydrothermal outflow, a topographically-elevated boiling, surging pool. Similar

1037 coring studies performed on extinct structures such as fountain geysers and low-relief, non-
1038 surging pools are likely to reveal different facies patterns. Further sinter stratigraphy,
1039 especially in the harsh environmental conditions of El Tatio, also provides a comparison for
1040 Archean and Martian sinters and their biosignature potential. Microfossils are present in all
1041 four cores, but are far more prevalent and well-preserved in outflow and distal deposits.
1042 Many vent-proximal sinters are ambiguously biogenic, but the sparse remains of microbial
1043 filaments indicate a biological presence, and such deposits should not be overlooked in future
1044 stratigraphic studies.

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1052 Competing Interests:

1053 The authors have no competing interests to declare.

1054 References

- 1055 Bargar, K.E., Beeson, M.H., 1981. Hydrothermal alteration in research drill hole Y-2, Lower
1056 Geysers basin, Yellowstone National Park, Wyoming. *Am. Mineral.* 66, 473–490.
- 1057 Berelson, W.M., Corsetti, F.A., Pepe-Ranne, C., Hammond, D.E., Beaumont, W., Spear,
1058 J.R., 2011. Hot spring siliceous stromatolites from Yellowstone National Park:

1059 Assessing growth rate and laminae formation. *Geobiology* 9, 411–424.
1060 <https://doi.org/10.1111/j.1472-4669.2011.00288.x>

1061 Bosak, T., Liang, B., Sim, M.S., Petroff, A.P., 2009. Morphological record of oxygenic
1062 photosynthesis in conical stromatolites. *Proc. Natl. Acad. Sci.* 106, 10939–10943.
1063 <https://doi.org/10.1073/pnas.0900885106>

1064 Bosak, T., Bush, J.W.M., Flynn, M.R., Liang, B., Ono, S., Petroff, A.P., Sim, M.S., 2010.
1065 Formation and stability of oxygen-rich bubbles that shape photosynthetic mats.
1066 *Geobiology* 8, 45–55. <https://doi.org/10.1111/j.1472-4669.2009.00227.x>

1067 Bosak, T., Liang, B., Wu, T.D., Templer, S.P., Evans, A., Vali, H., Guerquin-Kern, J.L.,
1068 Klepac-Ceraj, V., Sim, M.S., Mui, J., 2012. Cyanobacterial diversity and activity in
1069 modern conical microbialites. *Geobiology*. [https://doi.org/10.1111/j.1472-](https://doi.org/10.1111/j.1472-4669.2012.00334.x)
1070 [4669.2012.00334.x](https://doi.org/10.1111/j.1472-4669.2012.00334.x)

1071 Boudreau, A.E., Lynne, B.Y., 2012. The growth of siliceous sinter deposits around high-
1072 temperature eruptive hot springs. *J. Volcanol. Geotherm. Res.* 247–248, 1–8.
1073 <https://doi.org/10.1016/j.jvolgeores.2012.07.008>

1074 Bradley, J.A., Daille, L.K., Trivedi, C.B., Bojanowski, C.L., Stamps, B.W., Stevenson, B.S.,
1075 Nunn, H.S., Johnson, H.A., Loyd, S.J., Berelson, W.M., Corsetti, F.A., Spear, J.R.,
1076 2017. Carbonate-rich dendrolitic cones: Insights into a modern analog for incipient
1077 microbialite formation, Little Hot Creek, Long Valley Caldera, California. *npj Biofilms*
1078 *Microbiomes* 3. <https://doi.org/10.1038/s41522-017-0041-2>

1079 Braunstein, D., Lowe, D.R., 2001. Relationship between spring and geyser activity and the
1080 deposition and morphology of high temperature (> 73 C) siliceous sinter, Yellowstone
1081 National Park, Wyoming, USA. *Journal of Sedimentary Research*, 71, 747-763.

- 1082 Cady, S.L., Farmer, J.D., 1996. Fossilization processes in siliceous thermal springs: Trends in
1083 preservation along thermal gradients. *CIBA Found. Symp.* 150–173.
1084 <https://doi.org/10.1002/9780470514986.ch9>
- 1085 Calvert, S.E., 1976. The Mineralogy and Geochemistry of Near-shore Sediments, in:
1086 *Chemical Oceanography*. pp. 187–280. [https://doi.org/10.1016/b978-0-12-588606-](https://doi.org/10.1016/b978-0-12-588606-2.50014-x)
1087 [2.50014-x](https://doi.org/10.1016/b978-0-12-588606-2.50014-x)
- 1088 Campbell, K.A., Sannazzaro, K., Rodgers, K.A., Herdianita, N.R., Browne, P.R.L., 2001.
1089 *Sedimentary Facies and Mineralogy of the Late Pleistocene Umukuri Silica Sinter,*
1090 *Taupo Volcanic Zone, New Zealand. J. Sediment. Res.* 71, 727–746.
1091 <https://doi.org/10.1306/2dc40964-0e47-11d7-8643000102c1865d>
- 1092 Campbell, K.A., Lynne, B.Y., Handley, K.M., Jordan, S., Farmer, J.D., Guido, D.M.,
1093 Foucher, F., Turner, S., Perry, R.S., 2015. Tracing Biosignature Preservation of
1094 Geothermally Silicified Microbial Textures into the Geological Record. *Astrobiology*.
1095 <https://doi.org/10.1089/ast.2015.1307>
- 1096 Campbell, K.A., Nicholson, K., Lynne, B.Y., Browne, P.R.L., 2020. 3D Anatomy of a 60-
1097 year-old siliceous hot spring deposit at Hipaua-Waihi-Tokaanu geothermal field, Taupo
1098 Volcanic Zone, New Zealand. *Sediment. Geol.*
1099 <https://doi.org/10.1016/j.sedgeo.2020.105652>
- 1100 Cockell, C.S., 2000. Ultraviolet radiation and the photobiology of earth's early oceans. *Orig.*
1101 *Life Evol. Biosph.* 30, 467–500. <https://doi.org/10.1023/A:1006765405786>
- 1102 Cortecchi, G., T. Boschetti, M. Mussi, C. H. Lameli, C. Mucchino, and M. Barbieri, 2005.
1103 New chemical and original isotopic data on waters from El Tatio geothermal field,
1104 northern Chile, *Geochem. J.*, 39(6), 547–571

1105 Cotten, J., Le Dez, A., Bau, M., Caroff, M., Maury, R., Dulski, P., Fourcade, S., Bohn, M.,
1106 Brousse, R., 1995. Origin of rare-earth element and yttrium enrichments in subaerial
1107 exposed basalts: evidence from French Polynesia. *Chem. Geol.* 119, 115-138.
1108 [https://doi.org/10.1016/0009-2541\(94\)00102-E](https://doi.org/10.1016/0009-2541(94)00102-E)

1109 De Silva, S.L., 1989. Altiplano-Puna volcanic complex of the central Andes. *Geology* 17,
1110 1102–1106. [https://doi.org/10.1130/0091-7613\(1989\)017<1102:APVCOT>2.3.CO;2](https://doi.org/10.1130/0091-7613(1989)017<1102:APVCOT>2.3.CO;2)

1111 De Silva, S.L., Self, S., Francis, P.W., Drake, R.E., Ramirez R., C., 1994. Effusive silicic
1112 volcanism in the Central Andes: the Chao dacite and other young lavas of the
1113 Altiplano-Puna volcanic complex. *J. Geophys. Res.* 99.
1114 <https://doi.org/10.1029/94jb00652>

1115 DGA Direccion general de aguas de chile. Información Oficial Hidrometeorológica y de
1116 Calidad de Aguas en Línea. <https://dga.mop.gob.cl/>. Accessed Mar 2017

1117 Djokic, T., Van Kranendonk, M.J., Campbel, K.A., Walter, M.R., Ward, C.R., 2017. Earliest
1118 signs of life on land preserved in ca. 3.5 Ga hot spring deposits. *Nat. Commun.* 8.
1119 <https://doi.org/10.1038/ncomms15263>

1120 Feng, J.L., Zhao, Z.H., Chen, F., Hu, H.P., 2014. Rare earth elements in sinters from the
1121 geothermal waters (hot springs) on the Tibetan Plateau, China. *J. Volcanol. Geotherm.*
1122 *Res.* 287, 1–11. <https://doi.org/10.1016/j.jvolgeores.2014.09.009>

1123 Fernandez-Turiel, J.L., Garcia-Valles, M., Gimeno-Torrente, D., Saavedra-Alonso, J.,
1124 Martinez-Manent, S., 2005. The hot spring and geyser sinters of El Tatio, northern
1125 Chile. *Sediment. Geol.* 180, 125–147. <https://doi.org/10.1016/j.sedgeo.2005.07.005>

- 1126 Frantz, C.M., Petryshyn, V.A., Corsetti, F.A., 2015. Grain trapping by filamentous
1127 cyanobacterial and algal mats: Implications for stromatolite microfabrics through time.
1128 *Geobiology* 13, 409–423. <https://doi.org/10.1111/gbi.12145>
- 1129 Gebelein, C.D., 1969. Distribution, morphology, and accretion rate of recent subtidal algal
1130 stromatolites, Bermuda. *J. Sediment. Petrol.* [https://doi.org/10.1306/74D71BE0-2B21-](https://doi.org/10.1306/74D71BE0-2B21-11D7-8648000102C1865D)
1131 [11D7-8648000102C1865D](https://doi.org/10.1306/74D71BE0-2B21-11D7-8648000102C1865D)
- 1132 Giggenbach, W.F., 1978. The isotopic composition of waters from the El Tatio geothermal
1133 field, Northern Chile. *Geochim. Cosmochim. Acta* 42, 979–988.
1134 [https://doi.org/10.1016/0016-7037\(78\)90287-9](https://doi.org/10.1016/0016-7037(78)90287-9)
- 1135 Glennon, J.A. and Pfaff, R.M., 2003. The extraordinary thermal activity of El Tatio geyser
1136 field, Antofagasta Region, Chile. *GOSA Trans* 8, 31-78.
- 1137 Gong, J., Myers, K.D., Munoz-Saez, C., Homann, M., Rouillard, J., Wirth, R., Schreiber, A.,
1138 van Zuilen, M.A., 2019. Formation and Preservation of Microbial Palisade Fabric in
1139 Silica Deposits from El Tatio, Chile. *Astrobiology* 20, 1–25.
1140 <https://doi.org/10.1089/ast.2019.2025>
- 1141 Guidry, S.A., Chafetz, H.S., 2003a. Anatomy of siliceous hot springs: Examples from
1142 Yellowstone National Park, Wyoming, USA. *Sediment. Geol.* 157, 71–106.
1143 [https://doi.org/10.1016/S0037-0738\(02\)00195-1](https://doi.org/10.1016/S0037-0738(02)00195-1)
- 1144 Guidry, S.A., Chafetz, H.S., 2003b. Depositional Facies and Diagenetic Alteration in a Relict
1145 Siliceous Hot-Spring Accumulation: Examples from Yellowstone National Park,
1146 U.S.A. *J. Sediment. Res.* 73, 806–823. <https://doi.org/10.1306/022803730806>

- 1147 Guidry, S.A., Chafetz, H.S., 2003c. Siliceous shrubs in hot springs from Yellowstone
1148 National Park, Wyoming, U.S.A. *Can. J. Earth Sci.* 40, 1571–1583.
1149 <https://doi.org/10.1139/e03-069>
- 1150 Handley, K.M., Campbell, K.A., 2011. Character, Analysis, and Preservation of Biogenicity
1151 in Terrestrial Siliceous Stromatolites from Geothermal Settings.
1152 https://doi.org/10.1007/978-94-007-0397-1_16
- 1153 Handley, K.M., Campbell, K.A., Mountain, B.W., Browne, P.R.L., 2005. Abiotic-biotic
1154 controls on the origin and development of spicular sinter: In situ growth experiments,
1155 Champagne Pool, Waiotapu, New Zealand. *Geobiology* 3, 93–114.
1156 <https://doi.org/10.1111/j.1472-4669.2005.00046.x>
- 1157 Handley, K.M., Turner, S.J., Campbell, K.A., Mountain, B.W., 2008. Silicifying biofilm
1158 exopolymers on a hot-spring microstromatolite: templating nanometer-thick laminae.
1159 *Astrobiology*. <https://doi.org/10.1089/ast.2007.0172>
- 1160 Healy, J. and Hochstein, M.P., 1973. Horizontal flow in hydrothermal systems. *Journal of*
1161 *Hydrology (New Zealand)*, pp.71-82.
- 1162 Hinman, N.W., Walter, M.R., 2005. Textural Preservation in Siliceous Hot Spring Deposits
1163 During Early Diagenesis: Examples from Yellowstone National Park and Nevada,
1164 U.S.A. *J. Sediment. Res.* 75, 200–215. <https://doi.org/10.2110/jsr.2005.016>
- 1165 Hogg, A.G., Hua, Q., Blackwell, P.G., Niu, M., Buck, C.E., Guilderson, T.P., Heaton, T.J.,
1166 Palmer, J.G., Reimer, P.J., Reimer, R.W., Turney, C.S.M., Zimmerman, S.R.H., 2013.
1167 SHCal13 Southern Hemisphere Calibration, 0–50,000 Years cal BP. *Radiocarbon* 55,
1168 1889–1903. https://doi.org/10.2458/azu_js_rc.55.16783

- 1169 Jones, B., Renaut, R.W., 1996. Influence of thermophilic bacteria on calcite and silica
1170 precipitation in hot springs with water temperatures above 90°C: Evidence from Kenya
1171 and New Zealand. *Can. J. Earth Sci.* <https://doi.org/10.1139/e96-008>
- 1172 Jones, B., Renaut, R.W., 1997. Formation of silica oncoids around geysers and hot springs at
1173 El Tatio, northern Chile. *Sedimentology* 44, 287–304. <https://doi.org/10.1111/j.1365-3091.1997.tb01525.x>
- 1175 Jones, B., Renaut, R.W., 2003. Petrography and genesis of spicular and columnar geyselite
1176 from the Whakarewarewa and Orakeikorako geothermal areas, North Island, New
1177 Zealand. *Can. J. Earth Sci.* 40, 1585–1610. <https://doi.org/10.1139/e03-062>
- 1178 Jones, B., Renaut, R.W., 2006. Growth of Siliceous Spicules in Acidic Hot Springs,
1179 Waiotapu Geothermal Area, North Island, New Zealand. *Palaios* 21, 406–423.
1180 <https://doi.org/10.2110/palo.2006.p06-026>
- 1181 Jones, B., Renaut, R.W., Rosen, M.R., 1997. Biogenicity of Silica Precipitation Around
1182 Geysers and Hot-Spring Vents, North Island, New Zealand. *SEPM J. Sediment. Res.*
1183 Vol. 67, 88–104. <https://doi.org/10.1306/d42684ff-2b26-11d7-8648000102c1865d>
- 1184 Jones, B., Renaut, R.W., Rosen, M.R., 1998. Microbial biofacies in hot-spring sinters; a
1185 model based on Ohaaki Pool, North Island, New Zealand. *J. Sediment. Res.* 68, 413–
1186 434. <https://doi.org/10.2110/jsr.68.413>
- 1187 Jones, B., Konhauser, K.O., Renaut, R., and Wheeler, R., 2004. Microbe silicification in
1188 Iodine Pool, Waimangu geothermal area, North Island, New Zealand: Implications for
1189 recognition and identification of ancient silicified microbes. *Journal of the Geological*
1190 *Society of London*, 161:983-993.

- 1191 Jones, B., Renaut, R.W., Konhauser, K.O., 2005. Genesis of large siliceous stromatolites at
1192 Frying Pan Lake, Waimangu geothermal field, North Island, New Zealand.
1193 *Sedimentology* 52, 1229–1252. <https://doi.org/10.1111/j.1365-3091.2005.00739.x>
- 1194 Jones, B., Renaut, R.W., Owen, R.B., 2011. Life cycle of a geyser discharge apron: Evidence
1195 from Waikite Geyser, Whakarewarewa geothermal area, North Island, New Zealand.
1196 *Sediment. Geol.* 236, 77–94. <https://doi.org/10.1016/j.sedgeo.2010.12.008>
- 1197 Konhauser, K.O., Phoenix, V.R., Bottrell, S.H., Adams, D.G., Head, I.M., 2001. Microbial-
1198 silica interactions in Icelandic hot spring sinter: Possible analogues for some
1199 Precambrian siliceous stromatolites. *Sedimentology* 48, 415–433.
1200 <https://doi.org/10.1046/j.1365-3091.2001.00372.x>
- 1201 Konhauser, K.O. and Ferris, F.G., 1996. Diversity of iron and silica precipitation by
1202 microbial biofilms in hydrothermal waters, Iceland: Implications for Precambrian iron
1203 formations. *Geology*, 24:323-326.
- 1204 Konhauser, K.O., Jones, B., Reysenbach, A.L., Renaut, R.W., 2003. Hot spring sinters: Keys
1205 to understanding Earth's earliest life forms. *Can. J. Earth Sci.* 40, 1713–1724.
1206 <https://doi.org/10.1139/e03-059>
- 1207 Lahsen, A. and Trujillo, P., 1976. The geothermal field of El Tatio, Chile. In *Proceeding,*
1208 *Second United Nations Symposium on the Development and Use of Geothermal*
1209 *Resources, San Francisco* (Vol. 1, pp. 170-177).
- 1210 Landrum, J.T., Bennett, P.C., Engel, A.S., Alsina, M.A., Pastén, P.A., Milliken, K., 2009.
1211 Partitioning geochemistry of arsenic and antimony, El Tatio Geyser Field, Chile. *Appl.*
1212 *Geochemistry.* <https://doi.org/10.1016/j.apgeochem.2008.12.024>

- 1213 Lau, E., Nash, C.Z., Vogler, D.R., Cullings, K.W., 2005. Molecular diversity of
1214 cyanobacteria inhabiting coniform structures and surrounding mat in a Yellowstone hot
1215 spring. *Astrobiology* 5, 83–92. <https://doi.org/10.1089/ast.2005.5.83>
- 1216 Leidig, M., Zandt, G., 2003. Modeling of highly anisotropic crust and application to the
1217 Altiplano-Puna volcanic complex of the central Andes. *J. Geophys. Res. Solid Earth*
1218 108, ESE 5-1-ESE 5-15. <https://doi.org/10.1029/2001jb000649>
- 1219 Lewis, A.J., Palmer, M.R., Sturchio, N.C., Kemp, A.J., 1997. The rare earth element
1220 geochemistry of acid-sulphate and acid-sulphate-chloride geothermal systems from
1221 Yellowstone National Park, Wyoming, USA. *Geochim. Cosmochim. Acta* 61, 695–706.
1222 [https://doi.org/10.1016/S0016-7037\(96\)00384-5](https://doi.org/10.1016/S0016-7037(96)00384-5)
- 1223 Lloyd, E.F., 1972. Geology and hot springs of Orakei Korako. *New Zealand Geological*
1224 *Survey, Bulletin* 85, 1–164.
- 1225 Lowe, D.R., Braunstein, D., 2003. Microstructure of high-temperature (>73 °C) siliceous
1226 sinter deposited around hot springs and geysers, Yellowstone National Park: The role of
1227 biological and abiological processes in sedimentation. *Can. J. Earth Sci.* 40, 1611–
1228 1642. <https://doi.org/10.1139/e03-066>
- 1229 Lowenstern, J.B., Hurwitz, S., McGeehin, J.P., 2016. Radiocarbon dating of silica sinter
1230 deposits in shallow drill cores from the Upper Geyser Basin, Yellowstone National
1231 Park. *J. Volcanol. Geotherm. Res.* 310, 132–136.
1232 <https://doi.org/10.1016/j.jvolgeores.2015.12.005>
- 1233 Lynne, B.Y., 2012. Mapping vent to distal-apron hot spring paleo-flow pathways using
1234 siliceous sinter architecture. *Geothermics* 43, 3–24.
1235 <https://doi.org/10.1016/j.geothermics.2012.01.004>

- 1236 Lynne, B.Y., Campbell, K.A., 2003. Diagenetic transformations (opal-A to quartz) of low-
1237 and mid-temperature microbial textures in siliceous hot-spring deposits, Taupo
1238 Volcanic Zone, New Zealand. *Can. J. Earth Sci.* 40, 1679–1696.
1239 <https://doi.org/10.1139/e03-064>
- 1240 Lynne, B.Y., Campbell, K.A., Moore, J., Browne, P.R.L., 2008. Origin and evolution of the
1241 Steamboat Springs siliceous sinter deposit, Nevada, U.S.A. *Sediment. Geol.* 210, 111–
1242 131. <https://doi.org/10.1016/j.sedgeo.2008.07.006>
- 1243 Mata, S.A., Harwood, C.L., Corsetti, F.A., Stork, N.J., Eilers, K., Berelson, W.M., Spear,
1244 J.R., 2012. Influence of Gas Production and Filament Orientation on Stromatolite
1245 Microfabric. *Palaios* 27, 206–219. <https://doi.org/10.2110/palo.2011.p11-088r>
- 1246 Michard, A., 1989. Rare earth element systematics in hydrothermal fluids. *Geochim.*
1247 *Cosmochim. Acta* 53, 745–750. [https://doi.org/10.1016/0016-7037\(89\)90017-3](https://doi.org/10.1016/0016-7037(89)90017-3)
- 1248 Mountain, B.W., Benning, L.G., Boerema, J.A., 2003. Experimental studies on New Zealand
1249 hot spring sinters: Rates of growth and textural development. *Can. J. Earth Sci.* 40,
1250 1643–1667. <https://doi.org/10.1139/e03-068>
- 1251 Munoz-Saez, C., Saltiel, S., Manga, M., Nguyen, C., Gonnermann, H., 2016. Physical and
1252 hydraulic properties of modern sinter deposits: El Tatio, Atacama. *J. Volcanol.*
1253 *Geotherm. Res.* <https://doi.org/10.1016/j.jvolgeores.2016.06.026>
- 1254 Munoz-Saez, C., Manga, M., Hurwitz, S., Rudolph, M.L., Namiki, A., Wang, C.Y., 2015.
1255 Dynamics within geyser conduits, and sensitivity to environmental perturbations:
1256 Insights from a periodic geyser in the El Tatio geyser field, Atacama Desert, Chile. *J.*
1257 *Volcanol. Geotherm. Res.* 292, 41–55. <https://doi.org/10.1016/j.jvolgeores.2015.01.002>

1258 Munoz-Saez, C., Manga, M., Hurwitz, S., 2018. Hydrothermal discharge from the El Tatio
1259 basin, Atacama, Chile. *J. Volcanol. Geotherm. Res.* 361, 25–35.
1260 <https://doi.org/10.1016/j.jvolgeores.2018.07.007>

1261 Munoz- Saez, C., Manga, M., Hurwitz, S., Slagter, S., Churchill, D.M., Reich, M., Damby,
1262 D. and Morata, D., 2020. Radiocarbon dating of silica sinter and postglacial
1263 hydrothermal activity in the El Tatio geyser field. *Geophysical Research Letters* 47,
1264 e2020GL087908.

1265 Nicolau, C., Reich, M., Lynne, B., 2014. Physico-chemical and environmental controls on
1266 siliceous sinter formation at the high-altitude El Tatio geothermal field, Chile. *J.*
1267 *Volcanol. Geotherm. Res.* 282, 60–76. <https://doi.org/10.1016/j.jvolgeores.2014.06.012>

1268 Pepe-Ranney, C., Berelson, W.M., Corsetti, F.A., Treants, M., Spear, J.R., 2012.
1269 Cyanobacterial construction of hot spring siliceous stromatolites in Yellowstone
1270 National Park. *Environ. Microbiol.* 14, 1182–1197. [https://doi.org/10.1111/j.1462-](https://doi.org/10.1111/j.1462-2920.2012.02698.x)
1271 [2920.2012.02698.x](https://doi.org/10.1111/j.1462-2920.2012.02698.x)

1272 Phoenix, V.R., Bennett, P.C., Engel, A.S., Tyler, S.W., Ferris, F.G., 2006. Chilean high-
1273 altitude hot-spring sinters: A model system for UV screening mechanisms by early
1274 Precambrian cyanobacteria. *Geobiology* 4, 15–28. [https://doi.org/10.1111/j.1472-](https://doi.org/10.1111/j.1472-4669.2006.00063.x)
1275 [4669.2006.00063.x](https://doi.org/10.1111/j.1472-4669.2006.00063.x)

1276 Reyes, K., Gonzalez, N.I., Stewart, J., Ospino, F., Nguyen, D., Cho, D.T., Ghahremani, N.,
1277 Spear, J.R., Johnson, H.A., 2013. Surface orientation affects the direction of cone
1278 growth by *leptolyngbya* sp. Strain C1, a likely architect of coniform structures octopus
1279 spring (Yellowstone National Park). *Appl. Environ. Microbiol.* 79, 1302–1308.
1280 <https://doi.org/10.1128/AEM.03008-12>

1281 Riding, R., 2000. Microbial carbonates: The geological record of calcified bacterial-algal
1282 mats and biofilms. *Sedimentology* 47, 179–214. [https://doi.org/10.1046/j.1365-](https://doi.org/10.1046/j.1365-3091.2000.00003.x)
1283 [3091.2000.00003.x](https://doi.org/10.1046/j.1365-3091.2000.00003.x)

1284 Rimstidt, J.D., Cole, D.R., 1983. Geothermal mineralization I: The mechanism of formation
1285 of the Beowawe, Nevada, siliceous sinter deposit. *Am. J. Sci.*
1286 <https://doi.org/10.2475/ajs.283.8.861>

1287 Ruff, S.W., Farmer, J.D., 2016. Silica deposits on Mars with features resembling hot spring
1288 biosignatures at El Tatio in Chile. *Nat. Commun.* 7.
1289 <https://doi.org/10.1038/ncomms13554>

1290 Ruff, S.W., Campbell, K.A., Van Kranendonk, M.J., Rice, M.S., Farmer, J.D., 2020. The
1291 Case for Ancient Hot Springs in Gusev Crater, Mars. *Astrobiology* 20, 475–499.
1292 <https://doi.org/10.1089/ast.2019.2044>

1293 Sanchez-Garcia, L., Fernandez-Martinez, M.A., García-Villadangos, M., Blanco, Y., Cady,
1294 S.L., Hinman, N., Bowden, M.E., Pointing, S.B., Lee, K.C., Warren-Rhodes, K.,
1295 Lacap-Bugler, Cabrol, N.A., Parro, V., Carrizo, D., 2019. Microbial Biomarker
1296 Transition in High-Altitude Sinter Mounds From El Tatio (Chile) Through Different
1297 Stages of Hydrothermal Activity. *Front. Microbiol.*,
1298 <https://doi.org/10.3389/fmicb.2018.03350>

1299 Sanchez-Yanez, C., Reich, M., Leisen, M., Morata, D., Barra, F., 2017. Geochemistry of
1300 metals and metalloids in siliceous sinter deposits: Implications for elemental
1301 partitioning into silica phases. *Appl. Geochemistry*.
1302 <https://doi.org/10.1016/j.apgeochem.2017.03.008>

1303 Schultze-Lam, S., Ferris, F.G., Konhauser, K.O., Wiese, R.G., 1995. In situ silicification of
1304 an Icelandic hot spring microbial mat: implications for microfossil formation. *Can. J.*
1305 *Earth Sci.* 32, 2021–2026. <https://doi.org/10.1139/e95-155>

1306 Slagter, S., Reich M., Munoz-Saez, C., Southon J., Morata, D., Barra F., Jian Gong, J., Skok,
1307 J.R., 2019. Environmental controls on silica sinter formation revealed by radiocarbon
1308 dating: *Geology*, 47, 1–5, <https://doi.org/10.1130/G45859.1>

1309 Som, S.M., Buick, R., Hagadorn, J.W., Blake, T.S., Perreault, J.M., Harnmeijer, J.P., Catling,
1310 D.C., 2016. Earth’s air pressure 2.7 billion years ago constrained to less than half of
1311 modern levels. *Nat. Geosci.* 9, 448–451. <https://doi.org/10.1038/ngeo2713>

1312 Som, S.M., Catling, D.C., Harnmeijer, J.P., Polivka, P.M., Buick, R., 2012. Air density 2.7
1313 billion years ago limited to less than twice modern levels by fossil raindrop imprints.
1314 *Nature* 484, 359–362. <https://doi.org/10.1038/nature10890>

1315 Tassi, F., Martinez, C., Vaselli, O., Capaccioni, B., Viramonte, J., 2005. Light hydrocarbons
1316 as redox and temperature indicators in the geothermal field of El Tatio (northern Chile).
1317 *Appl. Geochemistry* 20, 2049–2062. <https://doi.org/10.1016/j.apgeochem.2005.07.013>

1318 Tassi, F., Aguilera, F., Darrah, T., Vaselli, O., Capaccioni, B., Poreda, R.J., Delgado Huertas,
1319 A., 2010. Fluid geochemistry of hydrothermal systems in the Arica-Parinacota,
1320 Tarapacá and Antofagasta regions (northern Chile). *J. Volcanol. Geotherm. Res.* 192,
1321 1–15.

1322 Trewin, N.H., 1993. Depositional environment and preservation of biota in the Lower
1323 Devonian hot-springs of Rhynie, Aberdeenshire, Scotland. *Trans. R. Soc. Edinb. Earth*
1324 *Sci.* 84, 433–442. <https://doi.org/10.1017/S0263593300006234>

- 1325 Tribovillard, N., Algeo, T.J., Lyons, T., Riboulleau, A., 2006. Trace metals as paleoredox and
1326 paleoproductivity proxies: An update. *Chem. Geol.* 232, 12–32.
1327 <https://doi.org/10.1016/j.chemgeo.2006.02.012>
- 1328 Trujillo, P., 1969. Estudio para el desarrollo geotermico en el norte de Chile–Manifestaciones
1329 termales de El Tatio. *Provincia de Antofagasta: CORFO Project Report.*
- 1330 Walter, M.R., 1976a. Chapter 8.8 Hot-Spring Sediments in Yellowstone National Park. *Dev.*
1331 *Sedimentol.* 20, 489–498. [https://doi.org/10.1016/S0070-4571\(08\)71153-1](https://doi.org/10.1016/S0070-4571(08)71153-1)
- 1332 Walter, M.R., 1976b. Chapter 3.3 Geyserites of Yellowstone National Park: An Example of
1333 Abiogenic “Stromatolites.” *Dev. Sedimentol.* 20, 87–112.
- 1334 Walter, M.R., 1972. A hot spring analog for the depositional environment of Precambrian
1335 iron formations of the Lake Superior Region. *Econ. Geol.*
- 1336 Walter, M.R., Des Marais, D., Farmer, J.D., Hinman, N.W., 1996. Lithofacies and Biofacies
1337 of Mid-Paleozoic Thermal Spring Deposits in the Drummond Basin, Queensland,
1338 Australia. *Palaios* 11, 497. <https://doi.org/10.2307/3515187>
- 1339 Walter, M.R., McLoughlin, S., Drinnan, A.N., Farmer, J.D., 1998. Palaeontology of
1340 Devonian thermal spring deposits, Drummond Basin, Australia. *Alcheringa* 22, 285–
1341 314. <https://doi.org/10.1080/03115519808619328>
- 1342 Ward, K. M., Zandt, G., Beck, S. L., Christensen, D. H., & McFarlin, H. (2014). Seismic
1343 imaging of the magmatic underpinnings beneath the Altiplano-Puna volcanic complex
1344 from the joint inversion of surface wave dispersion and receiver functions. *Earth and*
1345 *Planetary Science Letters*, 404, 43-53.
- 1346 Watts-Henwood, N., Campbell, K.A., Lynne, B.Y., Guido, D.M., Rowland, J. V., Browne,
1347 P.R.L., 2017. Snapshot of hot-spring sinter at Geyser Valley, Wairakei, New Zealand,

- 1348 following anthropogenic drawdown of the geothermal reservoir. *Geothermics*.
- 1349 <https://doi.org/10.1016/j.geothermics.2017.03.002>
- 1350 Zeil, W., 1959. Junger Vulkanismus in der Hochkordillere der Provinz Antofagasta (Chile).
- 1351 *Geol. Rundschau* 48, 218–232. <https://doi.org/10.1007/BF01801827>
- 1352

1353 Figure Captions:

1354 Figure 1

1355 Sampling location. A: Map showing general location of El Tatio in Chile. B: Aerial view of
1356 El Tatio's North Basin. Yellow dots represent significant hydrothermal outflow sources.
1357 Dashed outline represents the area shown in C and D. C: Aerial view of extinct sinter mound
1358 and drill locations. D: Oblique view of extinct sinter mound, highlighting topography of the
1359 sinter mound, interior pit, and drill locations. E: Profile of extinct sinter mound along transect
1360 in Fig. 1C.

1361 Figure 2

1362 Generalized sedimentary facies in El Tatio cores. Scale represents core depth below surface,
1363 see Fig. 1 for relative core positions within sinter mound. Asterisks by age dates represent
1364 secondary influence on ^{14}C data. Table 1 lists specific depths for ^{14}C samples, Supplementary
1365 Table 1 lists specific sample depths for all other chemistry samples.

1366 Figure 3

1367 Inclined sinter crust. A, B and C are hand specimen photographs. A, B: Nearly vertical sinter
1368 crusts in Core A at 1.4 m (A) and 1.6 m depth (B). Arrows in B highlight darker surface-
1369 normal lineations which transect through sinter crusts. C: Core A, 0.4 m depth. Lower-angle
1370 laminations at the top of inclined beds. Note paler, more porous textures compared with
1371 deeper fabrics. D: Photomicrograph showing the boundary between well-preserved sinter
1372 crusts in the lower right and more altered textures in the upper left, marked by dashed line.
1373 Arrows highlight layers with surface normal fabrics flaring upward and to the left. Core A, 2
1374 m depth. E-H: SEM images of Core A, 0.4 m depth. E: Cross-section of surface-normal
1375 fabric, composed of silica tubes growing from right to left, highlighted by dashed lines. F:
1376 Detail of surface-normal silica tube, highlighting scalloped porous core surrounded by a

1377 thinly-laminated cortex of silica. G, H: Filamentous microfossils within inclined sinter. Note
1378 the difference in size and preservation between microfossils and surface-normal fabrics in E
1379 and F.

1380 Figure 4

1381 Planar sinter crust. A: Hand specimen photograph, Core A, 0.2 m depth. Note the presence of
1382 fine detrital grains in darker layers. B, C: SEM images of Core A, 0.25 m depth. B: Light
1383 layer, comprised of vertical chains of silica spheres separated by interstitial pore space. C:
1384 Dark layer, highlighting porous zones surrounded by massive silica rinds.

1385 Figure 5

1386 Particulate sinter. A, B: Hand specimen photographs of particulate facies in Core A, 0.1 m
1387 depth (A), and Core E, 1.1 m depth, highlighting diffuse, wavy layering. Orange areas are
1388 secondary mineral phases. C: Photomicrograph from Core E, 1.2 m depth. Particulate
1389 microfacies. Lighter, rounded textures are opal-A spherules, accompanied by larger detrital
1390 grains. D, E: SEM images from Core E, 1.2 m depth, showing a cross-section of particulate
1391 sinter layers. Note the absence of microbial filaments or well-defined laminae.

1392 Figure 6

1393 Macroscale features of columnar and spicular sinter. A: Surficial exposure of columnar sinter
1394 growing upwards. Exposure location is on the outer margin of the cored sinter, 5 m northwest
1395 of Core A (Fig. 1C). B: Recent columnar sinter fragments adjacent to active boiling pools in
1396 El Tatio, ~40 m southeast of Core E. C: Horizontal columns with various stages of branching,
1397 Core C, 1.6 m depth. D: Angled branching columns, Core C, 1.1 m depth. E: Cross-section
1398 through horizontal columns, Core C, 1.5 m depth. Planar sinter laminae immediately
1399 overlying columns mark the boundary between lower horizontal columns and upper angled
1400 columns. F: Vertical column showing convex laminae, overlain by tufted microbial sinter.

1401 Core E, 3.65 m depth. G: Curved spicules. Note the size difference between spicules and
1402 columns. Core C, 1.7 m depth.

1403 Figure 7

1404 Petrography of columnar sinter. A: Rounded termination of laminated sinter column, Core C,
1405 1 m depth. B: Two adjacent columns showing coarser, more porous laminae than Core C,
1406 from Core E, 3.65 m depth. C-E: Core C, 1 m depth. C: Broken termination of sinter column.
1407 Note the faint convex laminations within the column, as well as thin yellow minerals present
1408 around all margins, including broken surfaces. D: Boundary between finely-laminated
1409 column and surrounding porous, detrital-rich interstitial sinter. A detrital-rich geopetal feature
1410 is highlighted by the white arrow. E: Fine geopetal silts located between columns, extensively
1411 cross-cut and separated by secondary silica veins. F, G: Core C, 1.8 m depth. F: Cross-section
1412 from the top down of an individual column, showing multiple generations of silica
1413 precipitation. G: Hollow silicified tube extending from a columnar surface, potentially
1414 representing a rare preserved microfossil.

1415 Figure 8

1416 Arborescent sinter. A-C: hand specimen photographs. A: Arborescent facies at the top of
1417 Core C, with textures branching upward into open air. B: Pronounced cm-scale arborescent
1418 facies, over- and underlain by layers of dense, less porous sinter. Core E, 3.15 m depth. Note
1419 the branching textures in individual “shrubs”. C: Upper branches of arborescent structures
1420 extend into detrital-rich darker overlying sinter, noted by a white arrow. Core E, 1.5 m depth.
1421 D: Photomicrograph of arborescent facies, showing branching chains of silica spherules
1422 separated by pore space. Core E, 1.4 m depth. E-G: SEM images from Core E, 3.15 m depth.
1423 E: Individual branching structure, highlighting the presence of microbial filaments. F:

1424 Magnified image of highlighted area in E, showing microbial filaments with associated
1425 angular detrital material. G: Microbial filaments without detrital material present.

1426 Figure 9

1427 Palisade mat sinter. A-C : Hand specimen photographs. A: Core C, 0.1 m depth. B: Core D,
1428 0.2 m depth. Note that fabrics at the top of the sample maintain fenestral qualities, but lose
1429 distinct lamination. C: Core E, 2.2 m depth. D-G: Photomicrographs of palisade sinters. D:
1430 Palisade fabric alternating between darker, less porous laminae and lighter, more porous
1431 zones with extensive fenestrae. Core D, 0.1 m depth. E: Palisade sinter with alternating low-
1432 and high-porosity laminae, with fewer fenestrae compared to Cores C and D. Low-porosity
1433 laminae are lighter in thin-sections, but darker in hand samples (see Fig. 9C). Core E, 2.35 m
1434 depth. F: Fenestrae within a light porous lamina, Core D, 0.2 m depth. G: Detrital sinter
1435 fragment (most likely columnar or spicular) embedded in palisade fabric. Core D, 0.1 m
1436 depth.

1437 Figure 10

1438 SEM imagery of different microbial sinters. A-C: Palisade fabrics from (A) Core C, 0.1 m
1439 depth, (B) Core D, 0.2 m depth, (C) Core E, 2.2 m depth. Fenestrae are less abundant in C
1440 than A or B. D: Filamentous microfossils separating palisade fenestrae. Core D, 0.2 m depth.
1441 E: Detailed view of filaments in D. F: Rounded fenestrae in tufted mat sinters, Core E, 3.6 m
1442 depth. G: Detailed view of microbial sinter separating fenestrae in F, showing well-preserved
1443 microbial filaments on top of more silicified zones.

1444 Figure 11

1445 Tufted mat sinter. A-C: Photomicrographs of microbial tufts from (A) Core E, 0.3 m depth
1446 (B) Core E, 0.5 m depth. C: Core E, 3.6 m depth. Rounded fenestrae separated by filamentous
1447 hourglass structures (see Fig. 10 for SEM of similar features). D-F: SEM images from Core

1448 E, 3.6 m depth. Single microbial tuft, highlighting (D) the entire tuft, (E) the basal meshwork
1449 of filamentous microfossils, and (F) individual filaments near the tufted tip.

1450 Figure 12

1451 Massive sinter. A: Hand specimen photograph from Core A, 2.7 m depth. B-D:
1452 Photomicrographs of massive sinter in Core A, 2.4 m depth (B); Core D, 0.8 m depth (C);
1453 Core E, 1.5 m depth, cross-polarized light (D). Sample comes from an intercalation of
1454 massive sinter within dendritic sinter, highlighting extensive white rims of secondary silica
1455 precipitation. E: SEM image from Core E, 1.9 m depth.

1456 Figure 13

1457 Sandstone facies. A-B: Hand specimen photographs. A: Laminated sandstone and sinter,
1458 Core E, 3.5 m depth. B: Massive sandstone containing matrix-supported pebbles and
1459 secondary silica cements propagating along fracture zones, Core E, 5.4 m depth. C-E:
1460 Photomicrographs under plane polarized light. C: Laminated sandstone and sinter, Core E,
1461 3.5 m depth. D: Massive sandstone, Core E, 5.2 m depth. E: Laminated sandstone and sinter
1462 containing a fragment of columnar or spicular sinter, Core E, 3.5 m depth. F: Massive
1463 sandstone, photomicrograph in cross-polarized light, Core E, 5.2 m depth.

1464 Figure 14

1465 Major element patterns in all cores. Due to the disparity in elemental concentrations between
1466 sandstones and overlying sinters in Core E, elements in the upper four meters are shown
1467 separately on different axes. Supplementary Table 1 contains detailed concentration and
1468 depth measurements for all major and trace elements.

1469 Figure 15

1470 Rare earth element patterns in all cores, normalized to clastic sediments within Mound 502.
1471 Depth profiles show total REE concentrations in each core, as well as profiles for Ce and Yb
1472 representing light and heavy REE signatures. Note the nearly identical values for Ce and Yb
1473 in all cores. Darker shades in REE patterns represent deeper samples in core, as shown in
1474 profiles. Core E is further split into four sub-sections dominated by different facies. Specific
1475 values are presented in the Supplementary Information.

1476 Figure 16

1477 Europium anomalies in all cores. Values are unit-less, and represent the relative enrichment
1478 or depletion of Eu compared to Sm and Gd (See Section 3.4 for calculations), in contrast with
1479 Eu concentrations (mg/kg) shown Figure 15 and the Supplementary Information. Values
1480 above 1 are enriched in Eu, and are interpreted here as a proxy for the relative influence of
1481 high-temperature hydrothermal fluid compared to clastic sedimentation.

1482 Figure 17

1483 Depositional history of the cored El Tatio sinter mound. ¹⁴C dates interpreted as detrital
1484 signatures or secondary alteration are not included. A: Schematic reconstruction of mound
1485 formation through time. B: Field photograph representing the sampling sites (yellow stars)
1486 from Slagter et al., 2019 compared to the position of drill cores presented in this study
1487 labelled with white arrows, modified from Slagter et al., 2019. C: Drone image showing the
1488 mound from above, representing the position of the features described in B.