

1 **Holocene atmospheric circulation in the central North Pacific: a new terrestrial**
2 **diatom and $\delta^{18}\text{O}$ dataset from the Aleutian Islands**

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22 *Highlights*:

- 23 ▪ New Holocene oxygen isotope record from the Aleutian Islands
- 24 ▪ Diatom $\delta^{18}\text{O}$ reflects shifts in synoptic-scale atmospheric circulation
- 25 ▪ Warmer/wetter early-mid Holocene, cooler/drier after 4.5 ka
- 26 ▪ Enhanced winter circulation corresponds to Holocene glacier advances
- 27 ▪ Current environmental changes unprecedented within past 9.6 ka

28 **Abstract**

29 The North Pacific is a zone of cyclogenesis that modulates synoptic-scale atmospheric
30 circulation, yet there is a paucity of instrumental and paleoclimate data to fully constrain its
31 long-term state and variability. We present the first Holocene oxygen isotope record
32 ($\delta^{18}\text{O}_{\text{diatom}}$) from the Aleutian Islands, using siliceous diatoms preserved in Heart Lake on
33 Adak Island (51.85° N, 176.69° W). This study builds on previous work demonstrating that
34 Heart Lake sedimentary $\delta^{18}\text{O}_{\text{diatom}}$ values record the $\delta^{18}\text{O}$ signal of precipitation, and correlate
35 significantly with atmospheric circulation indices over the past century. We apply this
36 empirical relationship to interpret a new 9.6 ka $\delta^{18}\text{O}_{\text{diatom}}$ record from the same lake,
37 supported by diatom assemblage analysis. Our results demonstrate distinct shifts in the
38 prevailing trajectory of storm systems that drove spatially heterogeneous patterns of moisture
39 delivery and climate across the region. During the early-mid Holocene, a warmer/wetter
40 climate prevailed due to a predominantly westerly Aleutian Low that enhanced advection of
41 warm ^{18}O -enriched Pacific moisture to Adak, and culminated in a $\delta^{18}\text{O}_{\text{diatom}}$ maxima (33.3 ‰)
42 at 7.6 ka during the Holocene Thermal Maximum. After 4.5 ka, relatively lower $\delta^{18}\text{O}_{\text{diatom}}$
43 indicates cooler/drier conditions associated with enhanced northerly circulation that persisted
44 into the 21st century. Our analysis is consistent with surface climate conditions inferred from
45 a suite of terrestrial and marine climate-proxy records. This new Holocene dataset bridges the
46 gap in an expanding regional network of paleoisotope studies, and provides a fresh
47 assessment of the complex spatial patterns of Holocene climate across Beringia and the
48 atmospheric forces driving them.

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54 **1. Introduction**

55 Numerous paleoenvironmental studies now contribute to a global synthesis and
56 understanding of Holocene climate change over the past 11.7 ka [*Mayewski et al.* 2004;
57 *Marcott et al.* 2013; *Rehfeld et al.* 2018]. By comparing common trends between individual
58 proxy records, these studies provide a means to infer the timing, scale, and spatial extent of
59 major Holocene climatic features. These include stepwise climate transitions, intervals
60 exceeding twentieth century warmth, and the low-frequency behaviour and modes of natural
61 climate variability. At broad (i.e. global) spatial and temporal scales these trends are
62 relatively coherent and unambiguous, yet at finer spatial scales, climate variability is more
63 pronounced due to local and regional factors. Such variability is highlighted in two recent
64 paleoclimate syntheses focused on west and eastern Beringia – the region extending from
65 northeast Siberia to northwest Canada (Fig. 1a) [*Brooks et al.* 2015; *Kaufman et al.* 2016].
66 While general circulation models (GCM) typically emphasise insolation as the key driver of
67 millennial-scale Holocene climate change [*Renssen et al.* 2009], these compilations indicate a
68 more complex and spatially heterogeneous climate evolution than implied by linear insolation
69 forcing alone. For example, major climatic features previously considered ubiquitous, such as
70 a prominent Holocene thermal maximum (HTM) [*Kaufman et al.* 2004], are now recognised
71 to be spatially asynchronous across this vast region [*Kaufman et al.* 2016]. Moreover,
72 existing terrestrial water isotope records are also shown to be ambiguous and contradictory
73 during the Holocene [*Kaufman et al.* 2016], and the most recent suite of model-data
74 comparisons reveal significant mismatches between simulated and reconstructed Holocene
75 temperatures in Alaska [*Zhang et al.* 2017].

76 At the synoptic scale, Beringia is located within the main centre of influence of the
77 Aleutian Low; one of the most dominant ocean-atmospheric systems in the Northern
78 Hemisphere and of global climate significance [*Rodionov et al.* 2007]. However, virtually all
79 available terrestrial paleoclimate data are restricted to mainland Alaska and eastern Russia

80 [Sundqvist *et al.* 2014; Brooks *et al.* 2015; Kaufman *et al.* 2016], and compared to lower
81 latitude regions, paleoisotope reconstructions are sparse [Kaufman *et al.* 2016]. This partly
82 reflects a lack of base-line water isotope measurements for constraining the regional water
83 isotope cycle [e.g. Welker, 2000; Anderson *et al.* 2016], as well as a paucity of lake core
84 studies with continuous sequences of carbonate-rich sediments – or suitable alternatives – for
85 isotopic analysis. Hence, to elucidate past and future climate in this region, there is an
86 outstanding requirement for greater spatial coverage of highly resolved and accurately dated
87 paleoclimate datasets, as well as an empirical-based understanding of the atmospheric and
88 environmental controls driving them.

89 To address this, we present the first Holocene oxygen isotope record from the
90 Aleutian Islands in south west Alaska. Our isotope measurements derive from siliceous
91 diatoms ($\delta^{18}\text{O}_{\text{diatom}}$) preserved in the sediments of Heart Lake, on Adak Island (Fig. 1b), and
92 are supported by diatom assemblage analysis of the same sedimentary sequence. We build on
93 earlier work by Bailey *et al.* [2015] who demonstrate that Heart Lake $\delta^{18}\text{O}_{\text{diatom}}$ values
94 correlate significantly with North Pacific climate indices over the past hundred years ($r =$
95 0.43 ; $p < 0.02$, $n = 28$). Here, we apply this empirically-derived understanding to interpret
96 new $\delta^{18}\text{O}_{\text{diatom}}$ data from a longer Heart Lake sediment core which extends back to 9.6 ka.
97 The primary aims are to: (1) investigate the forcing and response of this remote region to a
98 warming climate system as it transitioned from the last glacial period; (2) develop a Holocene
99 reconstruction of North Pacific atmospheric circulation; and (3) bridge the gap in the regional
100 network of proxy records to synthesise and assess complex spatio-temporal patterns of
101 natural climate variability across Beringia.

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103 **2. Regional Setting**

104 Heart Lake is a small ($\sim 0.25 \text{ km}^2$), freshwater through-flow system on Adak Island in the
105 central North Pacific (51.85° N , 176.69° W) (Fig. 1c). The island is volcanic and forms part

106 of the 1900-km-long Aleutian archipelago extending from mainland Alaska to the Russian-
107 Kamchatka Peninsula. The lake watershed area is $\sim 8 \text{ km}^2$ and is situated in low-relief hills
108 surrounded by mountainous terrain (Fig. 1c). There is a single lake basin with a maximum
109 depth of 8 m. One stream inflows from two larger lakes and a small outflow channel drains to
110 the Bering Sea $\sim 2 \text{ km}$ to the west. Lake volume is $\sim 8 \times 10^5 \text{ m}^3$ and water retention is an
111 estimated two weeks, based on the available stream gauge inflow data [TDX, 2013].
112 Inspection of available satellite imagery reveals that Heart Lake freezes over in winter and
113 this ice surface remains into spring [USGS, 2017].

114 Adak Island has a mild maritime climate compared to mainland Alaska and is
115 strongly affected by persistent fog and light rain in the summer, and frequent storms and
116 strong winds during winter [Rodionov *et al.* 2007]. Mean annual air temperature is $4.3 \text{ }^\circ\text{C}$,
117 and mean winter (December–February) and summer (June–August) values are $1.0 \text{ }^\circ\text{C}$ and 9.0
118 $^\circ\text{C}$, respectively (1949–2016) [NOAA, 2017]. Mean December and July precipitation is 163
119 mm and 71 mm, respectively (Fig. 1d) [NOAA, 2017]. Of the total 1.3 m annual precipitation,
120 $\sim 75 \%$ (1.0 m) falls from September to February.

121 The regional climate reflects the configuration of large scale atmospheric–ocean
122 systems, namely the Aleutian Low: a synoptic-scale feature of mean low sea level pressure
123 (SLP) and the leading driver of North Pacific climate [Mock *et al.* 1998]. When the Aleutian
124 Low is ‘weak’, storms tend to track north over the central Aleutian Islands (Fig. 2a); when
125 the pressure system is ‘strong’, storms track south of the Aleutians and into the Gulf of
126 Alaska (Fig. 2b) [Mock *et al.* 1998; Rodionov *et al.* 2007]. These circulation patterns vary on
127 interannual to decadal timescales and induce characteristic climate responses that are well
128 expressed in coupled modes of the North Pacific Index (NPI) and the Pacific Decadal
129 Oscillation (PDO) [Trenberth and Hurrell, 1994; Mantua *et al.* 1997]. Typically, a strong
130 Aleutian Low ($-NPI/+PDO$) will induce positive sea surface temperatures (SST), surface air
131 temperatures (SAT), and precipitation anomalies in the Gulf of Alaska and negative

132 anomalies in the central North Pacific, with contrary conditions during a weak Aleutian Low
133 (+NPI/−PDO) (see Supplementary Fig. 1).

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135 **3. Materials and Methods**

136 **3.1. Sediment and water recovery**

137 Sediment cores and bottom lake water samples were recovered from Heart Lake during the
138 summers of 2009 and 2010. A Garmin GPS sonar was used to survey its bathymetry and
139 revealed a single basin with a maximum depth of 8 m, surrounded by a shallow platform < 2
140 m deep (see Supplementary Figure 2). Coring sites were selected adjacent to the deepest part
141 of the basin at a depth of 7.6 m. Seven sediment cores were extracted using percussion and
142 hand-held gravity coring devices operated from a floating platform. Bottom lake water
143 samples were collected *in situ* at the sediment-water interface during gravity coring.

144 Following core extraction the water was directly siphoned from the corer and sealed in 50 ml
145 vials, ensuring no head space. Sediment cores were then split lengthways, packaged, and
146 shipped with water samples to Northern Arizona University where they were stored at 4 °C
147 until they were sub-sampled and analyzed. Our study focuses on the longest percussion core
148 (10-AS-1D; 5.9 m) and two accompanying surface gravity cores (09-AS-1A, 0.81 m; and 09-
149 AS-1B, 0.44 m). For a detailed description of the sediment core's lithostratigraphy, see
150 *Krawiec et al.* [2013].

151 **3.2. Chronology**

152 The composite age model for 10-AS-1D and 09-AS-1A is presented in a separate paper
153 devoted to the tephrostratigraphy and radiometric dating of the Heart Lake sedimentary
154 sequence [*Krawiec et al.* 2013] (Supplementary Fig. 3a). In summary, a Monte Carlo
155 approach was employed to model the age-depth relation of 16 macrofossil AMS radiocarbon
156 (¹⁴C) dates, together with a peak in recent ²³⁹+²⁴⁰Pu activity and the age of the sediment-water

157 interface (2009 AD) [*Krawiec et al.* 2013]. Tephrostratigraphy was used to independently
158 cross check the accuracy of the chronology, whereby the ages of down core tephra horizons
159 from Heart Lake were compared with tephra ages from nearby Andrew Lake and previously
160 published outcrop studies [*Krawiec et al.* 2013]. The chronology for surface core 09-AS-1B
161 derives from radiometric dating of ^{210}Pb , ^{226}Ra , ^{137}Cs and ^{241}Am by direct gamma assay on 14
162 dried sediment samples from the upper core section [*Bailey et al.* 2015] (Supplementary Fig.
163 3b). The cores were cross-correlated using a prominent tephra horizon found in all three
164 sedimentary sequences [*Krawiec et al.* 2013; *Bailey et al.* 2015]. All ages herein are
165 expressed as thousands of calendar years (ka) prior to 1950 AD, where 1 ka = 1000 cal yr BP.

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167 **3.3. Stable isotope analyses**

168 A total of 147 sediment samples were processed for $\delta^{18}\text{O}_{\text{diatom}}$ analysis. These samples range
169 in age from 9.6 ka (587 cm depth) to 2009 AD, and are sub-/decadally resolved for the most
170 recent 1500 years and at centennial resolution thereafter. From the 5.9 m-long core 10-AS-
171 1D, 1 cm³ of sediment was extracted at 7 cm intervals from the base (587 cm) to the top of
172 the core. This was the optimal sampling resolution to avoid tephra layers which could
173 potentially cause contamination issues [*Lamb et al.* 2007]. The surface cores 09-AS-1A and
174 09-AS-1B were both sampled in contiguous 0.5 cm increments. This detail was used to
175 capture sub-decadal changes in $\delta^{18}\text{O}_{\text{diatom}}$ over the past century for direct comparison with
176 instrumental datasets [see *Bailey et al.* 2015]

177 Sediment samples were prepared using a hybrid process of chemical digestion,
178 sieving, and heavy liquid separation adapted from *Morley et al.* [2004]. To remove organic
179 and carbonate material, samples were treated with 30% H_2O_2 at 90°C until reactions ceased,
180 before using 5 % HCl at ambient temperature. Samples were then centrifuged in sodium
181 polytungstate ($3\text{Na}_2\text{WO}_4 \cdot 9\text{WO}_3 \cdot \text{H}_2\text{O}$) (SPT) heavy liquid at 2500 rpm for 20 minutes,
182 resulting in the separation and suspension of diatoms from the heavier detritus. This

183 procedure was repeated three times for each sample using specific gravities of 2.50, 2.30 and
184 2.25 g ml⁻¹. After the final SPT separation, samples were washed five times in ultrapure water
185 (UPW) at 1500 rpm for 5 minutes and vacuum filtered through a 3 µm cellulose nitrate
186 membrane to remove potential clay minerals and/or broken diatom fragments. The < 3 µm
187 fraction was discarded as it was too small (< 1 mg) to be analyzed and, upon further
188 inspection under a light microscope, contained only small broken diatom fragments and
189 detritus. The remaining samples were treated with a final stage of 30 % H₂O₂ at 60 °C for one
190 week to ensure no traces of organic matter remained.

191 Purified diatom samples were analyzed for δ¹⁸O_{diatom} using the stepwise fluorination
192 method [Leng and Sloane, 2008] at the NERC Isotope Geosciences Laboratory in Keyworth,
193 UK. The outer hydrous layer of the diatom was removed in a pre-fluorination stage using a
194 BrF₅ reagent at low temperature [Leclerc and Labeyrie, 1987]. This was followed by a full
195 reaction at high temperature to liberate oxygen that was converted to CO₂ [Clayton and
196 Mayeda, 1963] and measured for δ¹⁸O_{diatom} using a MAT 253 dual-inlet mass spectrometer.
197 Replicate analyses indicate an analytical reproducibility of ±0.19 ‰ (1σ) for the samples, and
198 ±0.30 ‰ (1σ) for the diatom standard BFC_{mod}. All δ¹⁸O values were converted to the Vienna
199 Standard Mean Ocean Water (VSMOW) scale using the BFC_{mod} standard for calibration.

200 Two Heart Lake water samples were measured for their oxygen and hydrogen (δD)
201 isotope composition using a Thermo-Finnigan Deltaplus XL gas mass spectrometer at the
202 Colorado Plateau Stable Isotope Laboratory, Northern Arizona University, USA. Analytical
203 precision on internal working standards was ±0.1 ‰ for δ¹⁸O and ±1 ‰ for δD. All values are
204 reported here in per mil (‰) relative to VSMOW.

205

206 **3.3.1. Contamination assessment**

207 All purified diatom samples (n = 147) were visually inspected for contamination using an
208 OLYMPUS BX40 light microscope. Thirty samples were selected down-core and further

209 inspected using a Hitachi S-4700 field emission scanning electron microscope (SEM). In
210 addition, fourier transform infrared spectroscopy (FTIR) was applied to assess the chemical
211 composition and sample purity of 16 diatom samples from core 10-AS-1D [Swann and
212 Patwardham, 2011]. These samples, together with the BFC_{mod} diatom standard, were
213 analyzed using FTIR at the British Geological Survey in Keyworth, UK [Bailey et al. 2014].
214 FTIR analyses of all purified diatom isotope samples measured (n=16) indicate peaks
215 corresponding to the BFC_{mod} standard, known to represent clean, fossilised diatomite
216 (Supplementary Fig. 4). Spectral deviation from the standard would indicate additional
217 compounds and contamination by non-diatom components [Swann and Patwardhan, 2011];
218 peaks centred at ~450 cm⁻¹, ~800 cm⁻¹ and ~1100 cm⁻¹ confirm pure silica and the integrity of
219 our diatom isotope samples [Bailey et al. 2014].

220

221 **3.4. Diatom assemblage analysis**

222 Fifty-seven sub-samples of the purified diatom material used for $\delta^{18}\text{O}_{\text{diatom}}$ analysis were
223 retained for diatom species analysis. These include 33 samples selected at *c.* 13 cm intervals
224 from AS-10-1D, and 24 samples at a contiguous 0.5 cm resolution from AS-09-1B. Diatom
225 slides were prepared on a hot plate using Naphrax[®] mounting medium. A minimum of 300
226 diatom frustules per sample were counted along transects at x1000 magnification, under an
227 OLYMPUS BX40 light microscope. Taxonomic identification was based on classifications in
228 *Camburn and Charles* [2000] and *Krammer and Lange-Bertalot* [1986–1991].

229 Following diatom identification, species counts were converted to percentage
230 abundance and evaluated using the software package Tilia (v.2.0.41) [Grimm, 2015]. For
231 diatom zone demarcation, a constrained incremental sum-of-squares cluster analysis
232 (CONISS) [Grimm, 1987] was applied to all dominant taxa with a relative abundance >5 %
233 in at least one sample. To quantitatively assess the down core trends in diatom assemblages, a
234 principal components analysis (PCA) [ter Braak and Prentice, 1988] was applied to a

235 correlation matrix based on the dominant (>5 %) diatom species in all 57 samples. The
236 analysis was performed on untransformed percentage data using the program C2 (v.1.7.6)
237 [Juggins, 2014].

238

239 **4. Results**

240 **4.1. Diatom flora**

241 Diatom frustules are well preserved in all samples and show no sign of valve dissolution. The
242 flora is diverse and a total of 155 different freshwater diatom species were identified. Of
243 these, 11 species account for > 90 % of all diatoms present in all samples. These include
244 species belonging to the genera *Aulacoseira*, *Cyclotella*, *Rossithidium*, and small fragilarioid
245 taxa (consisting of the genera *Fragilaria*, *Pseudostaurosira*, *Staurosira*, *Stauroforma*, and
246 *Staurosirella*). Species with an abundance ≥ 5 % in at least one stratigraphic level are
247 presented (Fig. 3), and the record is divided into four zones based on the CONISS
248 dendrogram: *Zone 1* (9.6–8.6 ka; 587–452 cm), *Zone 2* (8.6–4.4 ka; 452–352 cm), *Zone 3*
249 (4.4 ka–1860 AD; 352–13.25 cm), and *Zone 4* (1860–2009 AD; 13.25–0 cm). Species are
250 grouped into one of three habitat types (planktonic, benthic, or facultatively planktonic) based
251 on classifications by Spaulding *et al.* [2017] (Fig. 3).

252 Diatom *Zone 1* (587–452 cm; ca. 9.6–8.6 ka) is dominated by *Staurosirella pinnata*
253 (33 %), *Cyclotella ocellata* (18 %), and other small fragilarioid taxa (60 %) (Fig. 3). By ca.
254 9.0 ka the abundance of *S. pinnata* decreases to 10 % and the planktonic species *Cyclotella*
255 *rossii* (10–30 %), *Aulacoseira subarctica* (4–25 %) and *Cyclotella ocellata* (5–14 %) are
256 more dominant. Some of the small benthic species all show slight increases in abundance at
257 this time, including *Psammothidium levanderi* (9 %) and *Achnantheidium minutissimum* (6%),
258 albeit at a low relative abundance.

259 In *Zone 2* (452–352 cm; ca. 8.6–4.4 ka) the planktonic species *C. ocellata*, *A.*
260 *subarctica*, and *C. rossii* begin to dominate the assemblage (Fig. 3). Collectively these

261 species reach a maximum abundance of 75 % between 8.5–7.6 ka; a time when small benthic
262 and facultatively planktonic taxa are at their overall lowest Holocene abundances (0–5 %).
263 Increases in abundances of *Rossithidium pussillum* and other small fragilarioid taxa occur *ca.*
264 7.6 and 6.8 ka, concurrent with a decrease in planktonic taxa (Fig. 3). After *ca.* 5.0 ka, the
265 abundance of planktonic species gradually decrease, paralleled by increasing abundance of
266 facultatively planktonic taxa.

267 At the onset of Zone 3 (352–13.25 cm; 4.4 ka–1860 AD) a large increase in the
268 facultatively planktonic taxa is paralleled by declines in planktonic taxa (Fig. 3). Collectively,
269 the small fragilarioid taxa make up ~80 % of the assemblages in this zone and several species
270 attain their maximum Holocene abundance, including *S. pinnata* at 4.2 ka (39 %) and
271 *Staurosira construens* at 3.8 ka (28 %). In contrast, planktonic species decline from a mean
272 abundance of 55 % in Zone 2, to 5 % in Zone 3. Only *Tabellaria flocculosa* shows relatively
273 little change in abundance from Zone 2, remaining at ~4%. Of the benthic taxa, *Stauroforma*
274 *exiguiformis* and *R. pusillum* are also present in high abundances throughout Zone 3, with the
275 former attaining a maximum Holocene abundance of 26 % at *ca.* 2.2 ka.

276 In Zone 4 (13.25–0 cm; *ca.* 1860–2009 AD) the small fragilarioid taxa continue to
277 dominate the assemblage, comprising ~75 % of the total assemblage *ca.* 1910 AD (Fig. 3).
278 After this time, the abundance of facultatively planktonic taxa steadily decreases as the
279 benthic and planktonic species increase. After *ca.* 1970 AD, the numbers of *A. subarctica*
280 decreases substantially, such that only a few individual frustules were counted per sample.

281 Stratigraphic changes in diatom flora are captured in the first two PCA components,
282 which collectively account for 71 % of the total assemblage variance (Fig. 4). Additional
283 eigenvectors defined by the PCA (3–5) were not considered given they explain progressively
284 lower proportions of the total variance ($\lambda_3=0.108$, $\lambda_4=0.059$, $\lambda_5=0.038$). PCA 1 represents 57
285 % of total variance and correlates to the planktonic species at the positive extreme, and the
286 facultatively planktonic species at the negative extreme. PCA 2 accounts for 14 % of total

287 variance, and correlates to the small fragilarioid taxa (Fig. 4). The Holocene succession of
288 diatom communities in Heart Lake is further illustrated by the time-series of the 54 sample
289 scores on PCA axis 1 (Fig. 3).

290

291 **4.2. Oxygen isotopes**

292 Holocene $\delta^{18}\text{O}_{\text{diatom}}$ values vary between 24.6 ‰ (1805 AD) and 33.3 ‰ (7.6 ka) ($\bar{x} = 29.7$
293 ‰, $n = 137$) (Fig. 5) with a range of ± 8.7 ‰ that is appreciably greater than the standard
294 deviation of all samples (± 0.19 ‰) and diatom standards (± 0.30 ‰) measured. The base of
295 the Heart Lake sediment core has a $\delta^{18}\text{O}_{\text{diatom}}$ value of 29.7 ‰ at 9.6 ka, and values steadily
296 increase to the maximum Holocene value of 33.3 ‰ at *ca.* 7.6 ka (Fig. 5). After 4.9 ka
297 $\delta^{18}\text{O}_{\text{diatom}}$ becomes progressively lower until *ca.* 3.5 ka (27.8 ‰) where values remain stable
298 at ~ 29 – 30 ‰ until *ca.* 1.0 ka. After *ca.* 1.0 ka, $\delta^{18}\text{O}_{\text{diatom}}$ exhibits high variability to lower
299 values *ca.* 1250–1340 AD and 1430–1525 AD, and after 1640 AD there is a shift to overall
300 lower $\delta^{18}\text{O}_{\text{diatom}}$ values, including the Holocene minimum $\delta^{18}\text{O}_{\text{diatom}}$ value of 24.6 ‰ at 1805
301 AD. The $\delta^{18}\text{O}_{\text{diatom}}$ values then slightly increase between 1805–1903 AD, before decreasing
302 to the present day (29.8 ‰) (Fig. 5). Using the sub-division age of 4.2 ka for the mid-late
303 Holocene boundary [Walker *et al.* 2012], late Holocene $\delta^{18}\text{O}_{\text{diatom}}$ is significantly ($p < 0.001$)
304 lower than in the early–mid Holocene.

305

306 **5. Discussion**

307 **5.1. Oxygen isotope paleohydrology and paleoclimatology**

308 Oxygen isotope ratios measured in precipitation ($\delta^{18}\text{O}_{\text{P}}$) at Adak airport (1962–67, 1972–73;
309 $n = 60$) indicate mean annual precipitation-weighted $\delta^{18}\text{O}_{\text{P}}$ is -8.8 ‰, with small seasonal
310 differences between January (-9.4 ‰) and July (-8.9 ‰) [IAEA/WMO, 2017]. The
311 correspondence between Heart Lake water $\delta^{18}\text{O}$ and the local and global meteoric water lines
312 confirms that (1) Heart Lake water $\delta^{18}\text{O}$ reflects local precipitation, and (2) evaporative

313 effects influencing precipitation and lake water $\delta^{18}\text{O}$ are minimal with no isotopic enrichment
314 (Fig. 6). Specifically, the two Heart Lake bottom water ($\delta^{18}\text{O}_{\text{water}}$) samples collected in
315 summer 2009 and 2010 ($\bar{x} = -9.5 \text{‰}$) are directly comparable, within error, to the long term
316 winter and spring $\delta^{18}\text{O}_{\text{P}}$ values from Adak airport. These data indicate the lake water budget
317 is dominated by winter and spring precipitation (i.e. snowfall and melt) similar to many lakes
318 and streams across Alaska [Clegg and Hu, 2010; Lachniet et al. 2016; Vachula et al. 2017].

319 There is no correlation between mean monthly $\delta^{18}\text{O}_{\text{P}}$ and SAT ($r = 0.15$, $n = 72$) or
320 precipitation amount ($r = 0.03$, $n = 72$) at Adak airport. Instead, Bailey et al. [2015] found
321 that Adak Island $\delta^{18}\text{O}_{\text{P}}$ values are primarily controlled by the moisture source and trajectory
322 of local precipitating storm systems. Specifically, winters with intensified Aleutian Low
323 circulation are characterized by precipitation with significantly ($p < 0.05$) lower than mean
324 $\delta^{18}\text{O}_{\text{P}}$ values. These variations are explained by systematic shifts in the central foci of the
325 Aleutian Low; when the SLP minimum is near Adak (strong Aleutian Low), polar air masses
326 are drawn south and advect water vapor and precipitation that is relatively depleted in ^{18}O ,
327 along with lower-than-average winter temperatures and increased snowfall (Fig. 2b)
328 [Rodionov et al. 2007; Bailey et al. 2015]. In contrast, a weakened and westerly displaced
329 Aleutian Low increases the southerly Pacific moisture flux to Adak via an enhanced south-
330 westerly storm track (Fig. 2a) [Rodionov et al. 2007]. These systems carry warm ^{18}O -
331 enriched moisture, and bring higher-than-average temperatures and increased precipitation to
332 Adak Island [Bailey et al. 2015].

333 $\delta^{18}\text{O}_{\text{diatom}}$ is controlled by several environmental parameters which depend on local
334 hydrology, climate, and the seasonality of diatom growth [Barker et al. 2001; Rioual et al.
335 2001; Jones et al. 2004; Rosqvist et al. 2004; Leng and Barker, 2006; Schiff et al. 2009;
336 Mackay et al. 2011; Meyer et al. 2015; Chaplignin et al. 2016]. Previous work by Bailey et al.
337 [2015] showed that the surface core $\delta^{18}\text{O}_{\text{diatom}}$ record from Heart Lake correlates significantly
338 with the winter NPI during the instrumental period (1900–2009 AD) ($r = 0.43$, $p < 0.02$, $n =$

339 28). This positive relationship confirms that Heart Lake diatoms precipitate their silica
340 frustule in isotopic equilibrium with the lake water in which they grow [*Labeyrie, 1974;*
341 *Leclerc and Labeyrie, 1987*], independent of size or species-related vital effects [*Bailey et al.*
342 2014]. During the spring thaw, it is evident that winter season precipitation ($\delta^{18}\text{O}_p$) enters
343 Heart Lake coincident with onset of the spring diatom bloom. A limited component of
344 residual summer growth might be expected, but bulk $\delta^{18}\text{O}_{\text{diatom}}$ analysis is weighted toward
345 the main period of diatom growth in spring [*Leng et al. 2001; Bailey et al. 2014*]. Under the
346 assumption that similar climatic controls on $\delta^{18}\text{O}_p$ prevailed before 1900 AD, we use this
347 extended $\delta^{18}\text{O}_{\text{diatom}}$ record as a proxy for atmospheric circulation throughout the Holocene.

348

349 **5.2. Holocene environmental history of Adak Island**

350 **5.2.1. Early-mid Holocene, 9.6 – 4.4 ka**

351 Adak Island, along with the Aleutian chain, was glaciated during the last glacial maximum,
352 though there are few chronological constraints on the onset and pattern of ice retreat [*Coats,*
353 *1956; Bradley, 1948; Fraser and Snyder, 1959; Black, 1976*]. At Heart Lake, percussion
354 coring ceased at a depth of 587 cm without penetrating bedrock or till, indicating the
355 catchment deglaciated prior to 9.6 ka.

356 From 9.6–9.0 ka, the dominance of fragilarioid and other small benthic taxa reflect a
357 temperate oligotrophic shallow lake with an extensive littoral zone. These pioneering taxa
358 dominate polar to subpolar and mountainous tundra lakes [*Lotter and Bigler, 2000; Rühland*
359 *et al. 2003; Hausmann and Pienitz, 2009; Devlin and Finkelstein, 2011*] and their presence
360 suggests a relatively short growth season with cool air temperatures [*Smol et al. 2005;*
361 *Rühland et al. 2008; Hausmann and Pienitz, 2009*]. Cool/dry conditions at this time are
362 further supported by low concentrations of biogenic silica (BSi) and organic matter (OM) in

363 nearby Andrew Lake [Krawiec and Kaufman, 2014] and the dominance of *Salix* and
364 *Empetrum* in northern Adak [Heusser, 1978].

365 Heart Lake was increasingly colonized by planktonic diatoms between 9.3–4.4 ka
366 (Fig. 3). Of these, *A. subarctica* is common across Arctic and subarctic zones, and typically
367 shows pronounced periodicity with the spring maximum in non-stratified lakes [Bradbury et
368 al. 2002; Baier et al. 2004; Rioual et al. 2007; Gibson et al. 2003; Solovieva et al. 2015]. It is
369 a heavily silicified species, forming colonies that require turbulence-induced suspension to
370 remain within the photic zone [Rühland et al. 2008; Lotter et al. 2010], and indicates
371 persistent strong seasonal winds, together with associated turbulent water mixing and nutrient
372 upwelling [Wang et al. 2008; Andrén et al. 2015; Solovieva et al. 2015]. In contrast,
373 *Cyclotella* species have a competitive advantage over the heavily silicified *A. subarctica*
374 during strong stratification [Andrén et al. 2015] and typically bloom after ice-out in subarctic
375 regions [Rühland et al. 2008; Hoff et al. 2015]. In Kamchatka, *Cyclotella* spp. prosper during
376 warmer years [Lepskaya et al. 2010], and are broadly considered warm water indicators due
377 to their recent expansion across Arctic lakes [Smol et al. 2005; Rühland et al. 2008].
378 Collectively, these early-mid Holocene diatom assemblages reveal a phase of overall high
379 lake mixing and turbidity, reduced lake ice cover, and relatively high Si/P ratios [Interlandi et
380 al. 1999; Rühland et al. 2003; Rioual et al. 2007]. These changes are further summarized by
381 the Holocene time series of PCA 1 sample scores (Fig. 3).

382 The isotope composition of Heart Lake water was significantly ($p < 0.001$) higher
383 during the early-mid Holocene compared to the late Holocene (Fig. 5), reflecting the
384 prevalence of southerly storms delivering abundant precipitation with higher $\delta^{18}\text{O}$ values
385 [Bailey et al. 2015]. Such warm, southerly winter storms would promote turbulent mixing
386 and limit the development of winter lake ice, thereby extending the open-water growing
387 season and allowing for a spring diatom assemblage dominated by planktonic species (Fig.
388 3). *Aulacoseira subarctica*, in particular, is abundant in modern lake systems during years

389 with short, warm winters [Gibson et al. 2003; Horn et al. 2011]. Elevated pollen percentages
390 of *Cyperaceae* and other wetland species in northern Adak also imply warm/wet conditions at
391 this time [Heusser, 1978] and correspond to higher local lake levels prior to 4.0 ka [Krawiec
392 and Kaufman, 2014]. Peak $\delta^{18}\text{O}_{\text{diatom}}$ (33.3 ‰) suggests maximum Holocene warmth at 7.6
393 ka, an inference supported by the simultaneous maximum Holocene abundance of the warm
394 water indicator *C. ocellata* [Rühland et al. 2008] (Fig. 3).

395 The $\delta^{18}\text{O}_{\text{diatom}}$ record correlates positively with the time series of PCA 1 scores ($r =$
396 0.48, $p < 0.001$) and demonstrates that diatom community structure is indirectly connected to
397 climate over millennial timescales. It also indicates that diatom species changes are a natural
398 ecological response to climatically-driven shifts in lake water $\delta^{18}\text{O}$, as reflected in the
399 $\delta^{18}\text{O}_{\text{diatom}}$ record, rather than the converse (i.e. changes in diatom species drive $\delta^{18}\text{O}_{\text{diatom}}$
400 variation).

401

402 **5.2.2. Mid-late Holocene, 4.4 ka – present**

403 At around 4.4 ka, a major shift in diatom composition occurred with marked changes from a
404 predominantly planktonic assemblage to the dominance of small fragilarioid and benthic taxa
405 (Fig. 3). During this transition the relatively warm, deep, and well-mixed open-water
406 conditions of the early-mid Holocene (9.6–4.4 ka) gave way to a less turbulent, potentially
407 shallower lake. This transition coincides with a shift to lower $\delta^{18}\text{O}_{\text{diatom}}$ values in the late
408 Holocene, reflecting an increase of isotopically depleted water (i.e. snow and/or ice melt)
409 during the spring thaw [Bailey et al., 2015; Streletskiy et al. 2015], and reduced warm, ^{18}O -
410 enriched southerly storms that characterized the early-mid Holocene.

411 An increase in northerly winds and lower temperatures during the late Holocene
412 would have enhanced formation of winter lake ice, which in turn was insulated and prolonged
413 by increased winter snowfall [Mock et al. 1998]. Persistence of lake ice into the spring
414 shortens the aquatic growth season and restricts light penetration into the water column

415 during the spring bloom, thereby precluding the growth and development of planktonic
416 communities requiring an ice-free lake for photosynthesis and a turbulent, well-mixed water
417 column. Instead, the mid-late Holocene flora at Heart Lake is dominated by fragilarioid
418 species known to colonise benthic and periphytic habitats under lake ice cover [Lotter and
419 Bigler, 2000; Douglas and Smol, 2010; Biskaborn et al. 2016]. These benthic communities
420 would have further benefitted from the absence of competition for nutrients from planktonic
421 diatoms, which do not thrive under ice [Lepskaya et al. 2010; Roberts et al. 2015]. A
422 reduction in turbulent wind-driven lake mixing at this time may have also been responsible
423 for increased benthic production and a simultaneous expansion of the littoral zone and
424 benthic habitat [Bradbury, 1988]. Increased winter precipitation and subsequent spring snow
425 melt would account for the sedimentation increase at 3.8 ka from 0.2 to 0.8 mm/yr [Krawiec
426 and Kaufman, 2014]. This turbidity would have further reduced light penetration into the
427 benthic zone, thereby promoting fragilarioid taxa which thrive under limited light and
428 generally turbid conditions [Lotter and Bigler, 2000; Douglas and Smol, 2010].

429 The simultaneous changes in diatom species assemblages and $\delta^{18}\text{O}_{\text{diatom}}$ values ca. 4.4
430 ka reflect numerous factors affecting vertical mixing patterns, availability of resources (e.g.
431 light, nutrients), and thereby the algal production and composition of Heart Lake. These
432 pronounced changes broadly coincided with other paleoenvironmental changes on Adak
433 Island centred ca. 4.4 ka. For example the BSi and inferred chlorophyll-*a* record from nearby
434 Andrew Lake also indicates increased aridity after 4.0 ka [Krawiec and Kaufman, 2014],
435 while reconstructed plant assemblages show a reduction in *Cyperaceae* after ca. 4.5 ka as
436 cooler/drier conditions prevailed over Adak Island [Heusser, 1978].

437 Between 950 AD and 1200 AD, higher $\delta^{18}\text{O}_{\text{diatom}}$ indicates a transition to overall
438 warmer and wetter conditions on Adak (Fig. 5). A decrease in *Empetrum* vegetation across
439 northern Adak also indicates increased moisture [Heusser, 1978], while Krawiec and
440 Kaufman [2014] interpret sustained low BSi and chlorophyll-*a* content from Andrew Lake as

441 the stormiest interval on record. Our $\delta^{18}\text{O}_{\text{diatom}}$ values exhibit high variability between 950
442 and 1900 AD, implying the local climate was also wetter and more variable since 950 AD.
443 These conditions would account for the continued dominance of fragilarioid taxa over this
444 period with unstable lake conditions [Smol *et al.* 2005; Rühland *et al.* 2008; Hausmann and
445 Pienitz, 2009]. Additionally, a peak in sedimentation *ca.* 1.0 ka, attributed to increased
446 storminess [Krawiec and Kaufman, 2014], rendered conditions unfavourable for planktonic
447 diatom species due to increased sediment suspension and reduced light penetration. Unlike
448 numerous diatom assemblage records across the subarctic and Arctic, in Heart Lake there is
449 no major shift toward those taxa favouring longer growing seasons under warming climatic
450 conditions (e.g. *Cyclotella*) [Smol *et al.* 2005]. Conversely, benthic assemblages show an
451 increase after *ca.* 1860 AD (Fig. 3), reflecting an overall strengthening of Aleutian Low
452 circulation since 1900 AD [Trenberth and Hurrell, 1994] and increasingly unstable
453 environmental conditions on Adak Island over the past century. These findings are consistent
454 with observations in North America and Greenland that suggest shifts in *Cyclotella*
455 abundances are more closely related to lake mixing, water clarity and resource availability,
456 rather than direct temperature effects [Saros and Anderson, 2015].

457

458 **5.3. Regional paleoenvironmental context**

459 Our $\delta^{18}\text{O}_{\text{diatom}}$ reconstruction reveals distinct shifts in the prevailing trajectory of storm
460 systems delivering moisture to Adak Island. The primary trends suggest a relatively weak and
461 westerly positioned Aleutian Low during the early-mid Holocene (9.7–4.5 ka), with a
462 strengthening eastward shift after *ca.* 4.5 ka (Fig. 5). Based on 21st century observations,
463 typical climatic responses to a weakened Aleutian Low are: (1) a weakening of Pacific mid-
464 latitude storm tracks; (2) increased meridional flow to the central-western Bering Sea; and (3)
465 reduced winter sea surface heat loss in the central-western Bering Sea and enhanced heat loss
466 from the Okhotsk Sea [Mock *et al.* 1998; Rodionov *et al.* 2007]. Under this synoptic regime

467 the following conditions would be anticipated in regional paleoclimate records: (1) a
468 reduction in winter storms and precipitation in the Gulf of Alaska region; (2) positive
469 precipitation and temperature anomalies in the central-western Aleutian Islands; and (3) SST
470 warming and reduced winter sea ice extent in the central-western Bering Sea and contrary
471 conditions in the Okhotsk Sea.

472 In support of this synoptic-scale picture, vegetation and lake-level reconstructions
473 provide independent evidence for considerably drier winter conditions in eastern Beringia
474 during the early-mid Holocene [*Anderson et al.* 2005; *RS Anderson et al.*, 2006; *Zander et al.*
475 2013]. For example, numerous lakes in southern Alaska and the Yukon record lower-than-
476 present water levels during the early Holocene until *ca.* 8 ka [*Kaufman et al.* 2016], reflecting
477 a combination of higher summer temperatures and lower winter precipitation. Furthermore,
478 an inferred decrease in frequency and intensity of winter storms steered into the Gulf of
479 Alaska accounts for marked episodes of glacial retreat at this time [*Solomina et al.* 2015],
480 driven by reduced winter snowfall/accumulation and negative net mass balance.

481 The SST patterns associated with a weakened wintertime Aleutian Low are also
482 evident during the early-mid Holocene. Relatively warm early Holocene SSTs are
483 documented from the western Bering Sea [*Max et al.* 2012], reflecting a persistently negative
484 phase of the PDO during the early-mid Holocene and an increase in Pacific storms tracking
485 into the region [*Rodionov et al.* 2007]. In the Okhotsk Sea, alkenone-derived SST estimates
486 correspond well with Heart Lake $\delta^{18}\text{O}_{\text{diatom}}$ between *ca.* 9.6–5.0 ka (Fig. 7), whereby higher
487 $\delta^{18}\text{O}_{\text{diatom}}$ and an inferred weak Aleutian Low corresponds to lower early-mid Holocene SSTs
488 [*Max et al.* 2012]. This relationship conforms to modern northerly geostrophic wind
489 anomalies during a weakened and westward displaced Aleutian Low that cool and enhance
490 polynya growth in the Okhotsk Sea [*Itaki and Ikahara*, 2004; *Harada et al.* 2014].
491 Specifically, warm (cold) winter SSTs in the Bering Sea (Okhotsk Sea) presently occur when
492 the Aleutian Low is shifted west and the Siberian High dominates over central western

493 Siberia [Rodionov *et al.* 2007]. These anti-correlated trends also manifest in sea-ice anomalies
494 on weekly to monthly time-scales during the 21st century [Cavaliere and Parkinson, 1987]
495 and are linked to the east–west migration of the Siberian High and Aleutian Low.

496 We find independent support for the Holocene migration of the Siberian High from
497 the Pechora Lake $\delta^{18}\text{O}$ record in northern Kamchatka [Hammarlund *et al.* 2015] (Fig. 7). A
498 north-eastward shift of the Siberian High, concurrent with a strong and eastward shifted
499 Aleutian Low, is linked to periods of increased winter snow contributions to Pechora Lake
500 and overall lower $\delta^{18}\text{O}$ values [Hammarlund *et al.* 2015]. The coherency of abrupt and
501 persistent change between the Heart and Pechora Lake $\delta^{18}\text{O}$ records between 9.6–3.5 ka
502 provides convincing evidence that the Aleutian Low–Siberian High system prevailed
503 throughout the early-mid Holocene (Fig. 7). Moreover, we propose that the synchronous
504 west-east migration of these systems may have been partially responsible for the non-linear
505 and heterogeneous climatic patterns reconstructed across east and west Beringia at this time
506 [Brooks *et al.* 2015; Kaufman *et al.* 2016].

507 Maximum values of $\delta^{18}\text{O}_{\text{diatom}}$ in Heart Lake at 7.6 ka broadly coincide with the
508 northern high-latitude (65 °N) summer insolation maxima *ca.* 8.0 ka (Fig. 7) [Berger and
509 Loutre 1991]. Significantly ($p < 0.001$) higher $\delta^{18}\text{O}_{\text{diatom}}$ in Heart Lake – relative to both the
510 modern (1900 AD–present) and long-term (9.6 ka–present) mean $\delta^{18}\text{O}_{\text{diatom}}$ – implies a HTM
511 in the central Aleutian Islands at 7.6 ka characterized by persistently weak Aleutian Low
512 circulation, and coincident with maximum abundances of warm water indicator species [Smol
513 *et al.* 2005] (Fig. 3). Similarly, a Holocene SST maximum is evident *ca.* 7.5 ka in both the
514 northwest Pacific [Minoshima *et al.* 2007] and the subarctic North Pacific [Harada *et al.*
515 2014], and from GCMs which indicate maximum SATs and SSTs in the Bering Sea and
516 Aleutian Islands *ca.* 7.0–8.0 ka [Renssen *et al.* 2012]. In southern Kamchatka, the majority of
517 paleoenvironmental records demonstrate a HTM *ca.* 7.0–5.3 ka [Brooks *et al.* 2015],
518 consistent with warm temperatures across eastern Beringia [Kaufman *et al.* 2016]. These

519 results contrast with previous paleoclimate studies from Alaska and the northwest Pacific that
520 identify an earlier HTM at *ca.* 11.3–9.1 ka [*Kaufman et al.* 2004; *Max et al.* 2012]. Such
521 uncertainty in these early Holocene warming patterns is highlighted by *Zhang et al.* [2017]
522 who found large discrepancies between modelled and reconstructed Holocene temperatures
523 across Alaska. Hence, it is difficult to fully constrain the timing of the HTM in the central
524 Aleutian Islands, particularly given that our record does not extend the full Holocene epoch
525 coupled with a paucity of local alternative studies.

526 Simultaneous shifts in diatom flora and $\delta^{18}\text{O}_{\text{diatom}}$ after the HTM at *ca.* 4.5 ka indicate
527 multiple and inter-related environmental changes that impacted Heart Lake. These
528 pronounced changes coincide with local proxy inferences demonstrating increased aridity
529 under a prevailing northerly circulation pattern [*Heusser, 1978; Corbett et al.* 2010; *Krawiec*
530 *and Kaufman, 2014*]. This mid-Holocene perturbation coincides with a return to cooler
531 conditions, increased winter precipitation and extensive glacial advance in Kamchatka
532 [*Nazarova et al.* 2013; *Barr and Solomina, 2014; Meyer et al.* 2015]. Widespread cooling is
533 also evident in eastern Beringia during the late Holocene [*Kaufman et al.* 2016], and
534 mountain glaciers across Alaska advanced between *ca.* 4.5 and 3.0 ka [*Solomina et al.* 2015],
535 in phase with those in Kamchatka and demarking onset of the Neoglacial across Beringia
536 [*Savoskul, 1999; Barr and Solomina, 2014*]. Though temperature is proposed as the principal
537 control on regional glacier mass balance through the Holocene [*Solomina et al.* 2015], the
538 observed glacial maxima in Alaska are asynchronous with the timing of pronounced cold
539 intervals [*Kaufman et al.* 2016]. Instead, our data suggest the transition to intensified Aleutian
540 Low circulation after 4.5 ka, coincident with declining summer insolation [*Berger and*
541 *Loutre, 1991*], drove widespread Neoglacial advance through the combined effect of
542 increased winter snowfall under a generally cooler regime, yielding a marked regional
543 positive mass balance perturbation. In particular, we note during the past millennium three
544 intervals of lower $\delta^{18}\text{O}_{\text{diatom}}$ values between 1275–1350 AD, 1400–1550 AD, and 1700–1850

545 AD coincide with three well-documented episodes of Little Ice Age (LIA) glacier advance on
546 mainland Alaska (Fig. 5 and 7) [Calkin *et al.* 2001; Solomina *et al.* 2015]. Furthermore, the
547 $\delta^{18}\text{O}_{\text{diatom}}$ minimum at 1805 AD (+24.6 ‰) marks the culmination of regional LIA glacial
548 advance [Barclay *et al.* 2009; Calkin *et al.* 2001; Wiles *et al.* 2004; Solomina *et al.* 2015]
549 (Fig. 5 and 7).

550

551 **5.4 Coherency of paleoisotope trends**

552 Several paleoisotope records from Alaska have also been interpreted in terms of synoptic-
553 scale changes in atmospheric circulation and inter-comparison with Heart Lake $\delta^{18}\text{O}_{\text{diatom}}$
554 yields many commonalities and insights [Anderson *et al.* 2005; Fisher *et al.* 2004; 2008;
555 Schiff *et al.* 2009; Jones *et al.* 2014; Hammarlund *et al.* 2015] (Fig. 7). For instance, a strong
556 inverse relationship *ca.* 9.5–4.0 ka is apparent with millennial scale $\delta^{18}\text{O}_{\text{diatom}}$ variations at
557 Mica Lake, in Prince William Sound [Schiff *et al.* 2009] (Fig. 7). Lower Mica Lake $\delta^{18}\text{O}_{\text{diatom}}$
558 values indicate precipitation delivered by zonal flow under a weak Aleutian Low, whereby
559 precipitating systems are subject to increased rainout as they pass over the Kenai Peninsula
560 and coastal mountain ranges. Conversely, increased meridional flow during a strong Aleutian
561 Low delivers locally sourced moisture from nearby Gulf of Alaska, thereby reducing
562 distillation and isotope depletion in precipitation, thus yielding higher Mica Lake $\delta^{18}\text{O}_{\text{diatom}}$
563 values [Schiff *et al.* 2009]. The reciprocal relationship between precipitation-inferred $\delta^{18}\text{O}$
564 values at Heart and Mica Lakes between *ca.* 9.5–4.0 ka conforms to modelling and empirical
565 analyses of spatial patterns of $\delta^{18}\text{O}_{\text{P}}$ [Berkelhammer *et al.* 2012; Bailey *et al.* 2015]. The
566 Horse Trail Fen record from the Kenai lowlands is also comparable to Heart Lake from *ca.*
567 8.0 ka and demonstrates overall higher $\delta^{18}\text{O}$ values during the early Holocene and reflecting
568 generally weak Aleutian Low circulation [Jones *et al.* 2014]. The only other full Holocene
569 paleoisotope record from eastern Beringia is from the Mount Logan ice core [Fisher *et al.*
570 2008], which exhibits strong correspondence with the Jellybean [Anderson *et al.* 2005] and

571 Heart Lake $\delta^{18}\text{O}$ records during the early-mid Holocene (Fig. 7).

572 Secondary, but notable departures between paleoisotope records are evident during
573 the late Holocene (Fig. 7), some of which can be reconciled by considering the detailed, non-
574 linear complexity of atmospheric circulation. For instance between *ca.* 3.0–1.0 ka Heart Lake
575 $\delta^{18}\text{O}_{\text{diatom}}$ does not exhibit marked excursions to the higher $\delta^{18}\text{O}$ values documented in Mt.
576 Logan, Jellybean and Pechora Lakes, interpreted as an interval of pronounced weak Aleutian
577 Low circulation [Anderson *et al.* 2005; Fisher *et al.* 2008; Hammarlund *et al.* 2015]. At Heart
578 Lake, this period is characterized by $\delta^{18}\text{O}_{\text{diatom}}$ values closer to the Holocene mean (Fig. 7).
579 These differences could reflect prevailing atmospheric patterns characterized by a more
580 southerly displaced western centre of low pressure in the northwest Pacific, which typically
581 results in a higher density of storms being steered into the Gulf of Alaska and eastern
582 Kamchatka Peninsula [Mock *et al.* 1998; Rodionov *et al.* 2007]. Under such conditions,
583 precipitation at Mt. Logan, Jellybean and Pechora Lakes would be relatively ^{18}O -enriched
584 [Berkelhammer *et al.* 2012], whereas Heart Lake would fail to exhibit higher $\delta^{18}\text{O}_{\text{diatom}}$ values
585 since these storm systems would track south of the Aleutian Islands [Rodionov *et al.* 2007].

586

587 **6. Conclusions**

588 The new datasets and analysis presented here extend modern observations across Alaska and
589 Siberia back through the Holocene to bridge a critical gap in the regional network of proxy-
590 climate records [Sundqvist *et al.* 2014; Brooks *et al.* 2015; Kaufman *et al.* 2016]. Although
591 GCMs typically emphasize insolation as the key driver of Holocene temperature change in
592 Alaska [Renssen *et al.* 2009], we demonstrate a more complex relationship and emphasise the
593 role of moisture availability and transport within the land–atmosphere–ocean system.

594 The Aleutian Islands straddle a critical zone of cyclogenesis that influences regional
595 temperature and precipitation patterns, including the heat and moisture flux between the

596 extratropical Pacific and Arctic; hence, the variable modes of atmospheric circulation we
597 identify have a wide reaching global influence through atmospheric-oceanic teleconnections.
598 Our empirically-derived understanding of the drivers and magnitude of these past changes
599 provide a means to contextualise contemporary climate trends, along with their potential
600 future trajectory and impact, across Alaska and the wider North Pacific. Specifically, we
601 demonstrate that Holocene shifts in Aleutian Low circulation directly impacted the net mass
602 balance of south-central Alaska's glaciers and ice fields through temperature and
603 precipitation variability [Solomina *et al.* 2015]. Given that Alaska is currently experiencing a
604 period of intensified Aleutian Low circulation, which should be favorable for glacier *growth*,
605 the widespread and well documented 21st century retreat of glaciers and ice cover [Larsen *et*
606 *al.* 2015] would now appear to be unprecedented within the context of long-term Holocene
607 environmental change.

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616

617 **8. Data availability**

618 Key datasets for this study are available in Supplementary Table 1. All data produced by this
619 study (*will be*) available online at the World Data Center for Paleoclimatology (WDC Paleo)

620 (<https://www.ncdc.noaa.gov/data-access/paleoclimatology-data>) and in the NERC National
621 Geoscience Data Centre (NGDC) (<http://www.bgs.ac.uk/services/ngdc/>).

622

623 **9. References**

624 Anderson, L., Abbot, M.B., Finney, B.P., and Burns, S.J., (2005), Regional atmospheric
625 circulation change in the North Pacific during the Holocene inferred from lacustrine
626 carbonate oxygen isotopes, Yukon Territory, Canada, *Quat. Res.*, 64, 21–35, doi:
627 10.1016/j.yqres.2005.03.005.

628 Anderson, R.S., Hallett, D.J., Berg, E., Jass, R.B., Toney, J.L., De Fontaine, C.S., and
629 DeVolder, A., (2006), Holocene development of boreal forests and fire regimes on the Kenai
630 Lowlands of Alaska, *The Holocene*, 16(6), 791–803, doi: 10.1191/0959683606hol966rp.

631 Anderson, L., Berkelhammer, M., Barron, J.A., Steinman, B.A., Finney, B.P., and Abbott,
632 M.B., (2016), Lake oxygen isotopes as recorders of North American Rocky Mountain
633 hydroclimate: Holocene patterns and variability at multi-decadal to millennial timescales,
634 *Glob. Planet. Change*, 137, 131–148, doi: 10.1016/j.gloplacha.2015.12.021.

635 Andrén, E., Klimaschewski, A., Self, A. E., Amour, N. S., Andreev, A. A., Bennett, K. D.,
636 Conley, D.J., Edwards, T.W.D., Solovieva, N., and Hammarlund, D., (2015), Holocene
637 climate and environmental change in north-eastern Kamchatka (Russian Far East), inferred
638 from a multi-proxy study of lake sediments, *Global Planet. Change*, 134, 41–54, doi:
639 10.1016/j.gloplacha.2015.02.013.

640 Baier, J., Lücke, A., Negendank, J.F., Schleser, G.H., and Zolitschka, B., (2004), Diatom and
641 geochemical evidence of mid-to late Holocene climatic changes at Lake Holzmaar, West-
642 Eifel (Germany), *Quat. Int.*, 113(1), 81–96, doi: 10.1016/S1040-6182(03)00081-8.

643 Bailey, H.L., Henderson, A.C.G., Sloane, H.J., Snelling, A., Leng, M.J., and Kaufman, D.S.,

644 (2014), The effects of species on lacustrine $\delta^{18}\text{O}_{\text{diatom}}$ and its implications for environmental
645 reconstructions, *J. Quat. Sci.*, 29, 393–400, doi: 10.1002/jqs.2711.

646 Bailey, H.L., Kaufman, D.S., Henderson, A.C.G., and Leng, M.J., (2015), Synoptic scale
647 controls on the $\delta^{18}\text{O}$ in precipitation across Beringia, *Geophys. Res. Lett.*, 42, 4608–4616,
648 doi: 10.1002/2015GL063983.

649 Barclay, D.J., Wiles, G.C., and Calkin, P.E., (2009), Holocene glacier fluctuations in Alaska,
650 *Quat. Sci. Rev.*, 28, 2034–2048, doi: 10.1016/j.quascirev.2009.01.016.

651 Barker, P.A., Street-Perrott, F.A., Leng, M.J., Greenwood, P.B., Swain, D.L., Perrott, R.A.,
652 Telford, J., and Ficken, K.J., (2001) A 14 ka oxygen isotope record from diatom silica in two
653 alpine tarns on Mt. Kenya, *Science*, 292, 2307–2310, doi: 10.1126/science.1059612.

654 Barr, I.D., and Solomina, O., (2014), Pleistocene and Holocene glacier fluctuations upon the
655 Kamchatka Peninsula, *Glob. Planet Change*, 113, 110–120, doi:
656 10.1016/j.gloplacha.2013.08.005.

657 Berger, A., and Loutre, M.F., (1991), Insolation values for the climate of the last 10 million
658 years, *Quat. Sci. Rev.*, 10, 297–317, doi:10.1016/0277-3791(91)90033-Q.

659 Berkelhammer, M., Stott, L., Yoshimura, K., Johnson, K., and Sinha, A., (2012), Synoptic
660 and mesoscale controls on the isotopic composition of precipitation in the western United
661 States, *Climate Dynamics*, 38 (3-4), 433–454, doi: 10.1007/s00382-011-1262-3.

662 Biskaborn, B.K., Subetto, D.A., Savelieva, L.A., Vakhrameeva, P.S., Hansche, A.,
663 Herzsuh, U., Klemm, J., Heinecke, L., Pestryakova, L.A., Meyer, H., and Kuhn, G.,
664 (2016), Late Quaternary vegetation and lake system dynamics in north-eastern Siberia:
665 Implications for seasonal climate variability, *Quat. Sci. Rev.*, 147, 406–421, doi:
666 10.1016/j.quascirev.2015.08.014.

667 Black, R.F., (1976), Late Quaternary glacial events, Aleutian Islands, Alaska. In:
668 Easterbrook, D.D., Sibrava, V. (Eds.), Quaternary Glaciations in the Northern Hemisphere.
669 IUGSUNESCO International Geological Correlations Program, Project 73-1-24. International
670 Union of Quaternary Research, Bellingham, pp. 285–301.

671 Bradbury, P., Cumming, B., and Laird, K., (2002), A 1500-year record of climatic and
672 environmental change in Elk Lake, Minnesota III: measures of past primary productivity, *J.*
673 *Paleolimnol.*, 27(3), 321–340, doi: 10.1023/A:1016035313101.

674 Bradley, C.C., (1948), Geologic notes on Adak Island and the Aleutian chain, Alaska, *Am. J.*
675 *Sci.*, 246(4), 214-240, doi: 10.2475/ajs.246.4.214.

676 Brooks, S.J., Diekmannb, B., Jones, V.J., and Hammarlund, D., (2015), Holocene
677 environmental change in Kamchatka: A synopsis, *Glob. Planet Change*, 134, 166–174, doi:
678 10.1016/j.gloplacha.2015.09.004.

679 Calkin, P.E., Wiles, G.C., and Barclay, D. J., (2001), Holocene coastal glaciation of Alaska,
680 *Quat. Sci. Rev.*, 20, 449–461, doi: 10.1016/S0277-3791(00)00105-0.

681 Camburn, K.E., and Charles, D.F., (2000), *Diatoms of Low-alkalinity Lakes in the*
682 *Northeastern United States*, ANSP Special Publication 18. Academy of Natural Sciences of
683 Philadelphia, Philadelphia.

684 Cavalieri, D.J., and Parkinson, C.L., (1987), On the relationship between atmospheric
685 circulation and the fluctuations in the sea ice extents of the Bering and Okhotsk Seas, *J.*
686 *Geophys. Res.*, 92, 7141–7162, doi: 10.1029/JC092iC07p07141.

687 Chaplignin, B., Narancic, B., Meyer, H., and Pienitz, R., (2016), Paleo-environmental
688 gateways in the eastern Canadian arctic—Recent isotope hydrology and diatom oxygen

689 isotopes from Nettilling Lake, Baffin Island, Canada, *Quat. Sci. Rev.*, 147, 379–390, doi:
690 10.1016/j.quascirev.2016.03.028.

691 Clayton, R.N., and Mayeda, T.K., (1963), The use of bromine pentafluoride in the extraction
692 of oxygen from oxide and silicates for isotopic analysis, *Geochim. Cosmochim. Acta*, 27, 43–
693 52, doi: 10.1016/0016-7037(63)90071-1.

694 Clegg, B.F., and Hu, F.S., (2010), An oxygen-isotope record of Holocene climate change in
695 south-central Brooks Range, Alaska, *Quat. Sci. Rev.*, 29, 928–939, doi:
696 10.1016/j.quascirev.2009.12.009.

697 Coats, R.R., (1956), *Reconnaissance geology of some western Aleutian Islands, Alaska*, US
698 Geological Survey Bulletin 1028-E, Government Printing Office.

699 Corbett, D., West, D., and Lefevre, C., (2010), *The People at the End of the World: The*
700 *Western Aleutian Project and the Archeology of Shemya Island*, Alaska Anthropological
701 Association Monograph Series VIII.

702 Devlin, J.E., and Finkelstein, S.A., (2011), Local physiographic controls on the responses of
703 Arctic lakes to climate warming in Sirmilik National Park, Nunavut, Canada, *J. Paleolimnol.*,
704 45(1), 23–39, doi: 10.1007/s10933-010-9477-6.

705 Douglas, M.S.V., and Smol, J.P., (2010), *Freshwater diatoms as indicators of environmental*
706 *change in the High Arctic*, In: Smol, J.P., Stoermer, E.F. (Eds.), *The Diatoms: Application for*
707 *the Environmental and Earth Sciences*. Cambridge University Press, Cambridge, pp.
708 249–266.

709 Fisher, D.A., Wake, C., Kreutz, K., Yalcin, K., Steig, E., Mayewski, P., Anderson, L., Zheng,
710 J., Rupper, S., Zdanowicz, C., Demuth, M., Waszkiewicz, M., Dahl-Jensen, D., Goto-Azuma,
711 K., Bourgeois, J.B., Koerner, R.M., Sekerka, J., Osterberg, E., Abbott, M.B., Finney, B.P.,

712 and Burn, S.J., (2004), Stable isotope records from Mount Logan, Eclipse ice cores and
713 nearby Jellybean Lake. Water cycle of the North Pacific over 2000 years and over five
714 vertical kilometres: Sudden shifts and tropical connections, *Geogr. Phys. Quat.*, 58 (2–3),
715 337–352, doi: 10.7202/013147ar.

716 Fisher, D., Osterberg, E., Dyke, A., Dahl-Jensen, D., Demuth, M., Zdanowicz, C., Bourgeois,
717 J., Koerner, R.M., Mayewski, P., Wake, C., Kreutz, K., Steig, E., Zheng, J., Yalcin, K., Goto-
718 Azuma, K., Luckman B., and Rupper, S., (2008), The Mt Logan Holocene-late Wisconsinan
719 isotope record: tropical Pacific--Yukon connections, *The Holocene*, 18, 667–677, doi:
720 10.1177/0959683608092236.

721 Fraser, G.D., and Snyder, G.L., (1959), *Geology of southern Adak Island and Kagalska*
722 *Island, Alaska*, US Geological Survey Bulletin 1028, pp. 371–408.

723 Gibson, C.E., Anderson, N.J., and Haworth, E.Y., (2003), *Aulacoseira subarctica*: taxonomy,
724 physiology, ecology and palaeoecology, *Eur. J. Phycol.*, 38, 83–101, doi:
725 10.1080/0967026031000094102.

726 Grimm, E.C., (1987), Coniss - a Fortran-77 program for stratigraphically constrained cluster-
727 analysis by the method of incremental sum of squares. *Comput. Geosci.*, 13, 13–35.

728 Grimm, E.C., (2015) TILIA software. Version 2.0.41. <https://www.tilait.com/download/>

729 Hammarlund, D., Klimaschewski, A., St. Amour, N.A., Andrén, E., Self, A.E., Solovieva, N.,
730 Andreev, A.A., Barnekowa, L., and Edwards, T.W.D, (2015), Late Holocene expansion of
731 Siberian dwarf pine (*Pinus pumila*) in Kamchatka in response to increased snow cover as
732 inferred from lacustrine oxygen-isotope records, *Glob. Planet. Change*, 134, 91–100, doi:
733 10.1016/j.gloplacha.2015.04.004.

734 Harada, N., Katsuki, K., Nakagawa, M., Matsumoto, A., Seki, O., Addison, J.A., Finney,
735 B.P., and Sato, M., (2014), Holocene sea surface temperature and sea ice extent in the
736 Okhotsk and Bering Seas, *Prog. Oceanogr.*, 126, 242–253, doi:
737 10.1016/j.pocean.2014.04.017.

738 Hausmann, S., and Pienitz, R., (2009), Seasonal water chemistry and diatom changes in six
739 boreal lakes of the Laurentian Mountains (Québec, Canada): impacts of climate and timber
740 harvesting, *Hydrobiologia*, 635(1), 1–14, doi: 10.1007/s10750-009-9855-0.

741 Heusser, C.J., (1978), Post-glacial vegetation on Adak Island, Aleutian Islands, Alaska. *Bull.*
742 *Torrey Bot. Club.*, 105, 18–23, doi: 10.2307/2484259.

743 Hoff, U., Biskaborn, B.K., Dirksen, V.G., Dirksen, O., Kuhn, G., Meyer, H., Nazarova, L.,
744 Roth, A., and Diekmann, B., (2015), Holocene environment of Central Kamchatka, Russia:
745 Implications from a multi-proxy record of Two-Yurts Lake, *Glob. Planet. Change*, 134,
746 101–117, doi: 10.1016/j.gloplacha.2015.07.011.

747 Horn, H., Paul, L., Horn, W., and Petzoldt, T., (2011), Long-term trends in the diatom
748 composition of the spring bloom of a German reservoir: is *Aulacoseira subarctica* favoured
749 by warm winters? *Fresh. Biol.*, 56(12), 2483–2499, doi: 10.1111/j.1365-2427.2011.02674.x.

750 IAEA/WMO, (2017), *Global Network of Isotopes in Precipitation*, The GNIP Database.
751 Accessible at: <http://www.iaea.org/water>.

752 Interlandi, S.J., Kilham, S.S., and Theriot, E.C., (1999), Responses of phytoplankton to
753 varied resource availability in large lakes of the Greater Yellowstone Ecosystem, *Limnol.*
754 *Oceanogr*, 44(3), 668–682, doi: 10.4319/lo.1999.44.3.0668.

755 Itaki, T., and Ikehara, K., (2004), Middle to late Holocene changes of the Okhotsk Sea
756 Intermediate Water and their relation to atmospheric circulation, *Geophys. Res. Lett.*, 31,
757 L24309, doi: 10.1029/2004GL021384.

758 Jones, V.J., Leng, M.J., Solovieva, N., Sloane, H.J. and Tarasov, P., (2004) Holocene climate
759 of the Kola Peninsula; evidence from the oxygen isotope record of diatom silica, *Quat. Sci.*
760 *Rev.*, 23(7–8), 833–839, doi: 10.1016/j.quascirev.2003.06.014.

761 Jones, M. C., Wooller, M., and Peteet, D.M., (2014), A deglacial and Holocene record of
762 climate variability in south-central Alaska from stable oxygen isotopes and plant macrofossils
763 in peat, *Quat. Sci. Rev.*, 87, 1–11, doi: 10.1016/j.quascirev.2013.12.025.

764 Juggins, S., (2014), C2 Data Analysis. Version 1.7.6. University of Newcastle, Newcastle.

765 Kalnay, E. *et al.* (1996), The NCEP/NCAR 40-Year Reanalysis Project. *Bull. Am. Meteorol.*
766 *Soc.*, 77, 437–471.

767 Kaufman, D.S., Ager, T.A., Anderson, N.J., Anderson, P.M., Andrews, J.T., Bartelein, P.J.,
768 Burbaker, L.B., Coats, L.L., Cwynar, L.C., Duval, M.L., Dyke, A.S., Edwards, M.E., Eiser,
769 W.R., Gajewski, K., Geisodottir, A., Hu, F.S., Jennings, A.E., Kaplan, M.R., Kewin, M.W.,
770 Lozhkin, A.V., MacDonald, G.M., Miller, G.H., Mock, C.J., Oswald, W.W., Otto-Blisner,
771 B.L., Porinchu, D.F., Rühland, K., Smol, J.P., Steig, E.J., and Wolfe, B.B., (2004), Holocene
772 thermal maximum in the western Arctic (0-180° W), *Quat. Sci. Rev.*, 23, 529–560, doi:
773 10.1016/j.quascirev.2003.09.007.

774 Kaufman, D.S., Axford, Y.L., Henderson, A.C.G., McKay, N.P., Oswald, W.W., Saenger, C.,
775 Anderson, R.S., Bailey, H.L., Clegg, B., Gajewski, K., Sheng Hu, F., Jones, M.C., Massa, C.
776 Routson, C.C., Werner, A., Wooller, M.J., and Yu, Z., (2016), Holocene climate changes in
777 eastern Beringia (NW North America) — A systematic review of multi-proxy evidence,
778 *Quat. Sci. Rev.*, 147, 312–339, doi: 10.1016/j.quascirev.2015.10.021.

779 Krammer, K., and Lange-Bertalot, H., (1986–1991), *Bacillariophyceae Band 2/2*. Gustav
780 Fischer Verlag, Stuttgart, pp.1–4.

781 Krawiec, A.C.L, and Kaufman, D.S., (2014), Holocene storminess inferred from sediments of
782 two lakes on Adak Island, Alaska, *Quaternary Res.*, 82, 73–84, doi:
783 10.1016/j.yqres.2014.02.007.

784 Krawiec, A.C.L, Kaufman, D.S., and Vaillencourt, D.A., (2013), Age models and
785 tephrostratigraphy from two lakes on Adak Island, Alaska. *Quat. Geochronol.*, 18, 41–53,
786 doi: 10.1016/j.quageo.2013.07.002.

787 Labeyrie, L.D., (1974), New approach to surface seawater palaeotemperatures using $^{18}\text{O}/^{16}\text{O}$
788 ratios in silica of diatom frustules, *Nature*, 248, 40–42, doi: 10.1038/248040a0.

789 Lachniet, M.S., Lawson, D.E., Stephen, H., Sloat, A.R., and Patterson, W.P., (2016),
790 Isoscapes of $\delta^{18}\text{O}$ and $\delta^2\text{H}$ reveal climatic forcings on Alaska and Yukon precipitation, *Water*
791 *Resour. Res.*, 52(8), 6575–6586, doi: 10.1002/2016WR019436.

792 Lamb., A.L., Brewer, T.S., Leng, M.J., Sloane, H.J., and Lamb, H.F., (2007), A geochemical
793 method for removing the effect of tephra on lake diatom oxygen isotope records, *J.*
794 *Paleolimnol.*, 37, 499–516, doi: 10.1007/s10933-006-9034-5.

795 Larsen, C.F., Burgess, E., Arendt, A.A., O'neel, S., Johnson, A.J. and Kienholz, C., (2015),
796 Surface melt dominates Alaska glacier mass balance, *Geophys. Res. Lett.*, 42(14),
797 5902–5908, doi: 10.1002/2015GL064349. .

798 Leclerc, A.J., and Labeyrie, L., (1987), Temperature dependence of the oxygen isotopic
799 fractionation between diatom silica and water, *Earth Planet. Sci. Lett.*, 84(1), 69–74, doi:
800 10.1016/0012-821X(87)90177-4.

801 Leng, M.J., and Barker, P.A., (2006), A review of the oxygen isotope composition of
802 lacustrine diatom silica for paleoclimate reconstruction, *Earth Sci. Rev.*, 75, 5–27, doi:
803 10.1016/j.earscirev.2005.10.001.

804 Leng, M.J., and Sloane, H.J., (2008), Combined oxygen and silicon isotope analysis of
805 biogenic silica, *J. Quat. Sci.*, 23, 313–319, doi: 10.1002/jqs.1177.

806 Leng, M., Barker, P., Greenwood, P., Roberts, N., and Reed, J., (2001), Oxygen isotope
807 analysis of diatom silica and authigenic calcite from Lake Pinarbasi, Turkey, *J. Paleolimnol.*,
808 25(3), 343–349, doi: 10.1023/A:1011169832093.

809 Lepskaya, E.V., Jewson, D.H., and Usoltseva, M.V., (2010), *Aulacoseira subarctica* in
810 Kurilskoye Lake, Kamchatka: a deep, oligotrophic lake and important Pacific salmon
811 nursery, *Diatom Research*, 25(2), 323–335, doi: 10.1080/0269249X.2010.9705853.

812 Lotter, A.F., and Bigler, C., (2000), Do diatoms in the Swiss Alps reflect the length of
813 ice-cover? *Aquatic Sciences*, 62(2), 125–141, doi: 10.1007/s000270050002.

814 Lotter, A.F., Pienitz, R., and Schmidt, R., (2010), *Diatoms as indicators of environmental*
815 *change in subarctic and alpine regions*, In: Smol, J.P., Stoermer, E.F. (Eds.), *The Diatoms:*
816 *Application for the Environmental and Earth Sciences*. Cambridge University Press,
817 Cambridge.

818 Mackay, A.W., Swann, G.E.A., Brewer, T.S., Leng, M.J., Morley, D.W., Piotrowska, N.,
819 Rioual, P., and White, D., (2011), A reassessment of late glacial—Holocene diatom oxygen
820 isotope record from Lake Baikal using a geochemical mass-balance approach, *J. Quat. Sci.*
821 26, 627–634, doi: 10.1002/jqs.1484.

822 Mantua, N.J., Hare, S.R., Zhang, Y., Wallace, J.M., and Francis, R.C., (1997), A Pacific
823 interdecadal climate oscillation with impacts on salmon production, *Bull. Am. Meteorol. Soc.*,
824 78, 1069–1079, doi: 10.1175/1520-0477(1997)078<1069:APICOW>2.0.CO;2.

825 Marcott, S.A., Shakun, J.D., Clark, P.U., and Mix, A.C., (2013), A reconstruction of regional
826 and global temperature for the past 11 300 years, *Science*, 339, 1198–1201, doi:
827 10.1126/science.1228026.

828 Max, L., Riethdorf, J-R., Tiedemann R., Smirnova, M., Lembke-Jene, L., Fahl, K., Nürnberg,
829 D., Matul, A., and Mollenhauer, G., (2012), Sea surface temperature variability and sea-ice
830 extent in the subarctic northwest Pacific during the past 15,000 years, *Paleoceanography*, 27,
831 PA3213, doi:10.1029/2012PA002292.

832 Mayewski, P.A., Rohling, E.E., Stager, J.C., Karlén, K.A., Maasch, W., Meeker, L.D.,
833 Meyerson, E.A., Gasse, F., van Kreveld, S., Holmgren, K., Lee-Thorp, J., Rosqvist, G., Rack,
834 F., Staubwasser, M., Schneider, R.R., and Steiger, E.J., (2004), Holocene climate variability,
835 *Qual. Res.*, 62, 243–255, doi: 10.1016/j.yqres.2004.07.001.

836 Meyer, H., Chapligin, B., Hoff, U., Nazarova, L., and Diekmann, B., (2015), Oxygen isotope
837 composition of diatoms as Late Holocene climate proxy at Two-Yurts-Lake, Central
838 Kamchatka, Russia. *Glob. Planet. Chang.*, 134, 118–128, doi:
839 10.1016/j.gloplacha.2014.04.008.

840 Minoshima, K., Kawahata, H., and Ikehara, K., (2007), Changes in biological production in
841 the mixed water region (MWR) of the northwestern North Pacific during the last 27 kyr,
842 *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, 254, 430–447, doi:10.1016/j.palaeo.2007.06.022.

843 Mock, C. J., Bartlein, P. J., and Anderson, P. M., (1998), Atmospheric circulation patterns
844 and spatial climatic variations in Beringia, *Int. J. Climatol.*, 10, 1085–1104,
845 doi:10.1002/(SICI)1097-0088(199808)18:10<1085::AID-JOC305>3.0.CO;2-K.

846 Morley, D.W., Leng, M.J., Mackay, A.W., Sloane, H.J., Rioual, P. and Battarbee, R.W.,
847 (2004), Cleaning of lake sediment samples for diatom oxygen isotope analysis, *J.*
848 *Paleolimnol.*, 31(3), 391–401, doi: 10.1023/B:JOPL.0000021854.70714.6b.

849 Nazarova, L., de Hoog, V., Hoff, U., Dirksen, O., and Diekmann, B., (2013), Late Holocene
850 climate and environmental changes in Kamchatka inferred from the subfossil chironomid
851 record, *Quat. Sci. Rev.*, 67, 81–92, doi: 10.1016/j.quascirev.2013.01.018.

852 NOAA, (2017), National Oceanic and Atmospheric Administration. National Climatic Data
853 Centre. <https://www.ncdc.noaa.gov/land-based-station-data>.

854 Rehfeld, K., Münch, T., Ho, S.L., and Laepple, T., (2018), Global patterns of declining
855 temperature variability from the Last Glacial Maximum to the Holocene, *Nature*, 554 (7692),
856 356–359, doi:10.1038/nature25454.

857 Renssen, H., Seppä, H., Heiri, O., Goosse, H., and Fichefet, T., (2009), The spatial and
858 temporal complexity of the Holocene thermal maximum, *Nat. Geosci.*, 2, 411–414, doi:
859 10.1038/ngeo513.

860 Renssen, H., Seppä, H., Crosta, X., Goosse, H., and Roche, D.M., (2012), Global
861 characterization of the Holocene Thermal Maximum, *Quat. Sci. Rev.*, 48, 7–19, doi:
862 10.1016/j.quascirev.2012.05.022.

863 Rioual, P., Andrieu-Ponel, V., Rietti-Shati, M., Battarbee, R.W., de Beaulieu, J.L., Cheddadi,
864 R., Reille, M., Svobodova, H., and Shemesh, A., (2001), High-resolution record of climate
865 stability in France during the last interglacial period, *Nature*, 413(6853), 293–296, doi:
866 10.1038/35095037.

867 Rioual, P., Andrieu-Ponel, V., de Beaulieu, J.L., Reille, M., Svobodova, H., and Battarbee,
868 R.W., (2007), Diatom responses to limnological and climatic changes at Ribains Maar

869 (French Massif Central) during the Eemian and Early Würm, *Quat. Sci. Rev.*, 26(11),
870 1557–1609, doi: 10.1016/j.quascirev.2007.03.009.

871 Roberts, S., Jones, V.J., Allen, J.R., and Huntley, B., (2015), Diatom response to mid-
872 Holocene climate in three small Arctic lakes in northernmost Finnmark, *The Holocene*, 25(6),
873 911–920, doi: 10.1177/0959683615572853.

874 Rodionov, S. N., Bond, N.A., and Overland, J.E., (2007), The Aleutian Low, storm tracks,
875 and winter climate variability in the Bering Sea, *Deep Sea Res. II.*, 54, 2560–2577, doi:
876 10.1016/j.dsr2.2007.08.002.

877 Rosqvist, G., Jonsson, C., Yam, R., Karlén, W., and Shemesh, A., (2004), Diatom oxygen
878 isotopes in pro-glacial lake sediments from northern Sweden: a 5000 year record of
879 atmospheric circulation, *Quat. Sci. Rev.*, 23(7), 851–859, doi:
880 10.1016/j.quascirev.2003.06.009.

881 Rühland, K., Priesnitz, A., and Smol, J.P., (2003), Paleolimnological evidence from diatoms
882 for recent environmental changes in 50 lakes across Canadian Arctic treeline, *Arct. Antarct.*
883 *Alp. Res.*, 35(1), 110–123, doi: 10.1657/1523-0430(2003)035[0110:PEFDFR]2.0.CO;2.

884 Rühland, K., Paterson, A.M., and Smol, J.P., (2008), Hemispheric-scale patterns of
885 climate-related shifts in planktonic diatoms from North American and European lakes, *Glob.*
886 *Change Biol.*, 14(11), 2740–2754, doi: 10.1111/j.1365-2486.2008.01670.x.

887 Saros, J.E. and Anderson, N.J., 2015. The ecology of the planktonic diatom *Cyclotella* and its
888 implications for global environmental change studies, *Biol. Rev.*, 90(2), 522–541, doi:
889 10.1111/brv.12120.

890 Savoskul, O.S., (1999), Holocene glacier advances in the headwaters of Sredniaya Avacha,
891 Kamchatka, Russia, *Qual. Res.*, 52, 14–26, doi: 10.1006/qres.1999.2051.

892 Schiff, C.J., Kaufman, D.S., Wolfe, A.P., Dodd, J., and Sharp, Z., (2009), Late Holocene
893 storm-trajectory changes inferred from the oxygen isotope composition of lake diatoms, south
894 Alaska, *J. Paleolimnol.*, 41, 189–208, doi: 10.1007/s10933-008-9261-z.

895 Smol, J. P., Wolfe, A. P., Birks, H. J. B., Douglas, M. S., Jones, V. J., Korhola, A., Pienitz,
896 R., Rühland, K., Sorvari, S., Antoniades, D., and Brooks, S. J., (2005), Climate-driven regime
897 shifts in the biological communities of arctic lakes., *PNAS*, 102(12), 4397–4402, doi:
898 10.1073/pnas.0500245102.

899 Solomina, O.N., Bradley, R.S., Hodgson, D.A., Ivy-Ochs, S., Jomelli, V., Macintosh, A.N.,
900 Nesje, A., Owen, L.A., Wanner, H., Wiles, G.C., and Young, N.E., (2015), Holocene glacier
901 fluctuations, *Quat. Sci. Rev.*, 111, 9–34, doi: 10.1016/j.quascirev.2014.11.018.

902 Solovieva, N., Klimaschewski, A., Self, A. E., Jones, V. J., Andrén, E., Andreev, A. A.,
903 Hammarlund, D., Lepskaya, E.V., and Nazarova, L., (2015), The Holocene environmental
904 history of a small coastal lake on the north-eastern Kamchatka Peninsula, *Glob. Planet.*
905 *Change*, 134, 55–66, doi: 10.1016/j.gloplacha.2015.06.010.

906 Spaulding, S.A., Lubinski, D.J. and Potapova, M., (2017), Diatoms of the United States.
907 <http://westerndiatoms.colorado.edu>

908 Streletskiy, D.A., Tananaev, N.I., Opel, T., Shiklomanov, N.I., Nyland, K.E., Streletskaya,
909 I.D. and Shiklomanov, A.I., (2015), Permafrost hydrology in changing climatic conditions:
910 seasonal variability of stable isotope composition in rivers in discontinuous permafrost.
911 *Environ. Res. Lett.*, 10 (9), p.095003, doi: 10.1088/1748-9326/10/9/095003.

912 Sundqvist, H.S., Kaufman, D.S., McKay, N.P., Balascio, N.L., Briner, J.P., Cwynar, L.C.,
913 Sejrup, H.P., Seppä, H., Subetto, D.A., Andrews, J.T. and Axford, Y., (2014), Arctic
914 Holocene proxy climate database—new approaches to assessing geochronological accuracy

915 and encoding climate variables *Clim. Past*, 10(4), 1605–1631, doi: 10.5194/cp-10-1605-
916 2014.

917 Swann, G.E.A., and Patwardhan, S.V., (2011), Application of Fourier Transform Infrared
918 Spectroscopy (FTIR) for assessing biogenic silica sample purity in geochemical analyses and
919 palaeoenvironmental research, *Clim. Past*, 7, 65–74, doi: 10.5194/cp-7-65-2011.

920 TDX (2013), TDX Power - Adak Reconnaissance Study. *Hatch*.
921 <http://akenergyinventory.org/hyd/SSH-2013-0004.pdf>

922 ter Braak, C.J.F., and Prentice, I.C., (1988), A theory of gradient analysis, *Adv. Ecol. Res.*,
923 18, 271–317, doi: 10.1016/S0065-2504(03)34003-6.

924 Trenberth, K.E., and Hurrell, J.W., (1994), Decadal atmosphere-ocean variations in the
925 Pacific, *Clim. Dyn.*, 9, 303–319, doi: 10.1007/BF00204745.

926 USGS, (2017), USGS *Landsat* 8 Images. Available at: <https://landsat.gsfc.nasa.gov/>

927 Vachula, R.S., Chipman, M.L., and Hu, F.S., (2017), Holocene climatic change in the
928 Alaskan Arctic as inferred from oxygen-isotope and lake-sediment analyses at Wahoo Lake,
929 *The Holocene*, 27 (4), 1–14, doi: <https://doi.org/10.1177/0959683617702230>.

930 Walker, M., Berhelhammer, M., Björck, S., Cwynar, L.C., Fisher, D.A., Long, A.J., Lowe,
931 J.J., Newnham, R.M., Rasmussen, S.O., and Weis, H., (2012), Formal subdivision of the
932 Holocene Series/Epoch: a discussion Paper by a Working Group of INTIMATE (Integration
933 of ice-core, marine and terrestrial records) and the Subcommittee on Quaternary
934 Stratigraphy (International Commission on Stratigraphy), *J. Quat. Sci.*, 27, 649–659, doi:
935 10.1002/jqs.2565.

936 Wang, L., Lu, H., Liu, J., Gu, Z., Mingram, J., Chu, G., Li, J., Rioual, P., Negendank, J.F.,
937 Han, J., and Liu, T., (2008), Diatom-based inference of variations in the strength of Asian

938 winter monsoon winds between 17,500 and 6000 calendar years BP, *J. Geophys. Res. Atmos.*,
939 113, D2101, doi: 10.1029/2008JD010145.

940 Welker, J.M., (2000), Isotopic ($\delta^{18}\text{O}$) characteristics of weekly precipitation collected
941 across the USA: an initial analysis with application to water source studies, *Hydrol.*
942 *Process.*, 14(8), 1449–1464, doi: 10.1002/1099-1085(20000615)14:8<1449::AID-
943 HYP993>3.0.CO;2-7.

944 Wiles, G., D'Arrigo, R., Villalba, R., Calkin, P., and Barclay, D.J., (2004), Century-scale
945 solar variability and Alaskan temperature change over the past millennium, *Geophys.*
946 *Res. Lett.* 31, L15203, doi: 10.1029/2004GL020050.

947 Zander, P.D., Kaufman, D.S., Kuehn, S.C., Wallace, K.L., and Anderson, R.S., (2013), Early
948 and late Holocene glacial fluctuations and tephrostratigraphy, Cabin Lake, Alaska, *J. Quat.*
949 *Sci.*, 28, 761–771, doi: 10.1002/jqs.2671

950 Zhang, Y., Renssen, H., Seppä, H., and Valdes, P.J., (2017), Holocene temperature
951 evolution in the Northern Hemisphere high latitudes—Model-data comparisons, *Quat. Sci.*
952 *Rev.*, 173, 101–113, doi: 10.1016/j.quascirev.2017.07.018.

953 **Author contributions**

954 D.S.K was PI, led the fieldwork and retrieved the sediment cores. D.S.K., H.L.B, H.J.S.,
955 A.C.G.H. and M.J.L developed the study concept. H.L.B conducted the research, sample
956 preparation, SEM, diatom and statistical analyses, interpreted the results, produced the
957 figures, and wrote the original manuscript. D.S.K. and A.L.H. critically revised the original
958 manuscript, and together with H.M. and J.M.W. provided technical advice and comments.
959 H.J.S performed the FTIR and diatom isotope measurements. M.J.L. supervised the diatom
960 isotope measurements and undertook the isotope corrections. H.J.S., A.C.G.H., M.J.L., H.M.
961 and J.M.W provided minor editorial revisions. All authors approved the final manuscript.

962 The authors declare no competing financial interests.

963

964 **Figure captions**

965

966 **FIG1.JPG** [Image size: 1.5 page width]

967 Figure 1. Location of (a) Adak Island in the central Aleutian Islands, (b) Heart Lake and
968 Andrew Lake, (c) oblique north west view of Heart Lake with the inflow channel visible in
969 the foreground [credit: Yarrow Axford], and (d) monthly mean precipitation (blue bars) and
970 surface air temperature at Adak airport (1949–2016), whereby solid lines depict mean
971 (black), minimum (blue) and maximum (red) temperatures [NOAA, 2017]. Numbered circles
972 in 1a indicate key sites referred to in text: (1) LV29-114-3 [Max *et al.* 2012], (2) Pechora
973 Lake [Hammarlund *et al.* 2015], (3) SO201-12-77KL [Max *et al.* 2012], (4) Horse Trail Fen
974 [Jones *et al.* 2014], (5) Mica Lake [Schiff *et al.* 2009], (6) Mount Logan [Fisher *et al.* 2008],
975 and (7) Jellybean Lake [Anderson *et al.* 2005]

976

977 **FIG2.JPG** [Image size: Column width]

978 Figure 2. Mean winter (December–February) sea level pressure associated with the six most
979 positive (a) and negative (b) North Pacific Index (NPI) values between 1950 and 2017
980 [Trenberth and Hurrell, 1994]. A negative (positive) NPI is a strong (weak) Aleutian Low.
981 Arrows highlight the direction of the primary storm tracks delivering precipitation to our site
982 on Adak Island (yellow star) [Bailey *et al.* 2015]. SLP data obtained from NCEP/NCAR V1
983 reanalysis [Kalnay *et al.* 1996]. Numbered yellow circles in (a) indicate locations of the (1)
984 LV29-114-3 [Max *et al.* 2012], (2) Pechora Lake [Hammarlund *et al.* 2015], (3) SO201-12-
985 77KL [Max *et al.* 2012], (4) Horse Trail Fen [Jones *et al.* 2014], (5) Mica Lake [Schiff *et al.*
986 2009], (6) Mount Logan [Fisher *et al.* 2008], and (7) Jellybean Lake [Anderson *et al.* 2005]
987 climate records discussed in text

988

989 **FIG3.PDF** [Image size: Full page width]

990 Figure 3. Heart Lake diatom stratigraphy and Principal Components Analysis (PCA) scores
991 of the 11 dominant diatom species (>5 % abundance), grouped by habitat preference. Diatom
992 zone demarcation (dashed lines 1–4) is guided by the CONISS cluster analysis. Variables are
993 plotted on a linear timescale (ka BP) and the depth scale refers to depth below lake floor

994

995 **FIG4.PDF** [Image size: Column width]

996 Figure 4. Loadings of the 11 dominant diatom taxa from Heart Lake and their corresponding
997 PCA scores. Sample scores (circles) are coloured according to their down core diatom
998 assemblage Zone (1–4). Dashed coloured ellipses group diatom species by their habitat
999 preference

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1001 **FIG5.PDF** [Image size: Full page width]

1002 Figure 5. Time series of Heart Lake $\delta^{18}\text{O}_{\text{diatom}}$ during (a) the past millennium and (b) the
1003 Holocene. Horizontal dashed grey lines indicate the Holocene and the 21st century mean
1004 $\delta^{18}\text{O}_{\text{diatom}}$ value. Orange diamonds and white triangles indicate previously published
1005 radiocarbon ages and tephra beds, respectively [Krawiec *et al.* 2013]. Vertical blue bars
1006 correspond to three intervals of Little Ice Age glacier advance in mainland Alaska [Solomina
1007 *et al.* 2015]

1008

1009 **FIG6.PDF** [Image size: Column width]

1010 Figure 6. Heart Lake bottom water $\delta^{18}\text{O}$ (2009 and 2010) on the local meteoric water line
1011 (LMWL) and the global meteoric water line (GMWL). LMWL data are derived from Adak
1012 monthly composite precipitation samples collected by the Global Network of Isotopes in
1013 Precipitation (GNIP) [IAEA/WMO, 2017]

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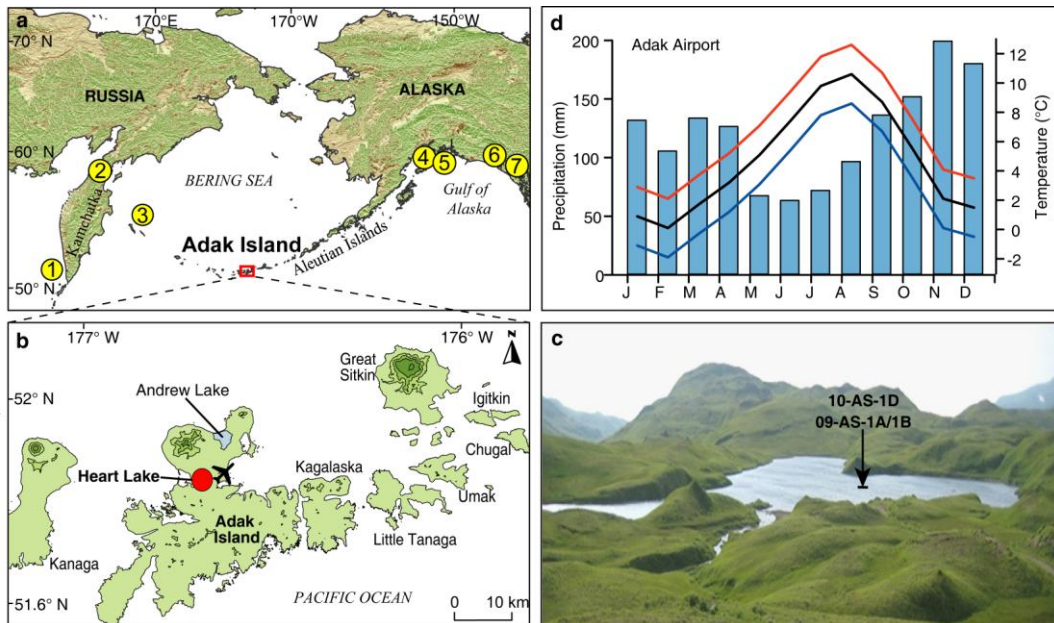
1015 **FIG7.PDF** [Image size: Column width]

1016 Figure 7. Holocene time series of (a) summer (JJA) insolation at 65°N [Berger and Loutre,
1017 1991], (b) alkenone SSTs from LV29-114-3 in the Okhotsk Sea [Max et al. 2012], (c)
1018 Pechora Lake $\delta^{18}\text{O}$ [Hammarlund et al. 2015], (d) Heart Lake $\delta^{18}\text{O}_{\text{diatom}}$ (this record), (e)
1019 intervals of expanded mountain glaciers in eastern Beringia [Solomina et al. 2015], (f) Mica
1020 Lake $\delta^{18}\text{O}$ [Schiff et al. 2009], (g) Mount Logan ice $\delta^{18}\text{O}$ [Fisher et al. 2008], (h) Horse Trail
1021 Fen $\delta^{18}\text{O}$ [Jones et al. 2014], and (i) Jellybean Lake $\delta^{18}\text{O}$ [Anderson et al. 2005]. Black lines
1022 in (g) and (i) represent 40-yr smoothed intervals. Vertical red shading indicates the eastern
1023 Beringia mid-Holocene Thermal Maximum [Kaufman et al. 2016], blue shading indicates the
1024 Little Ice Age (LIA) [Solomina et al. 2015]

1025 **Figures**

1026

FIG1



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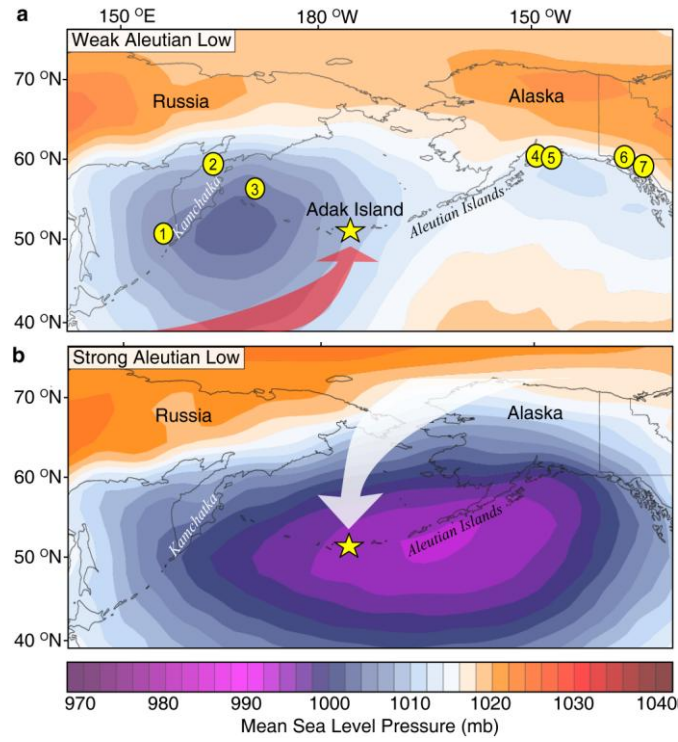
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FIG2



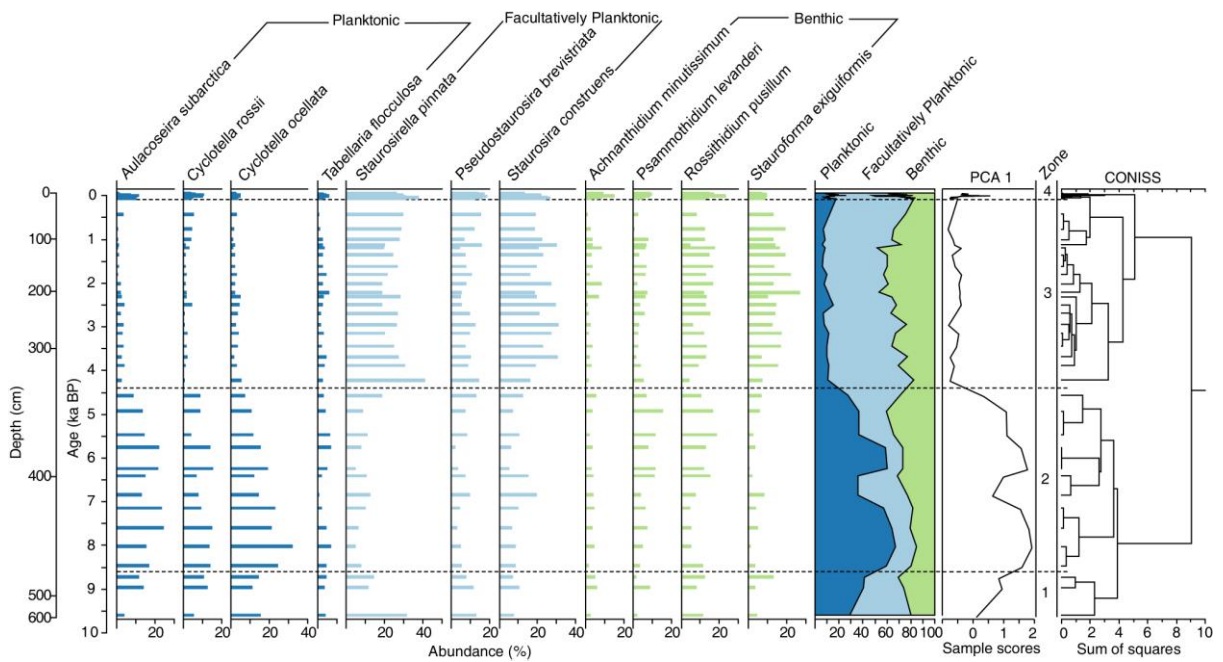
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FIG3



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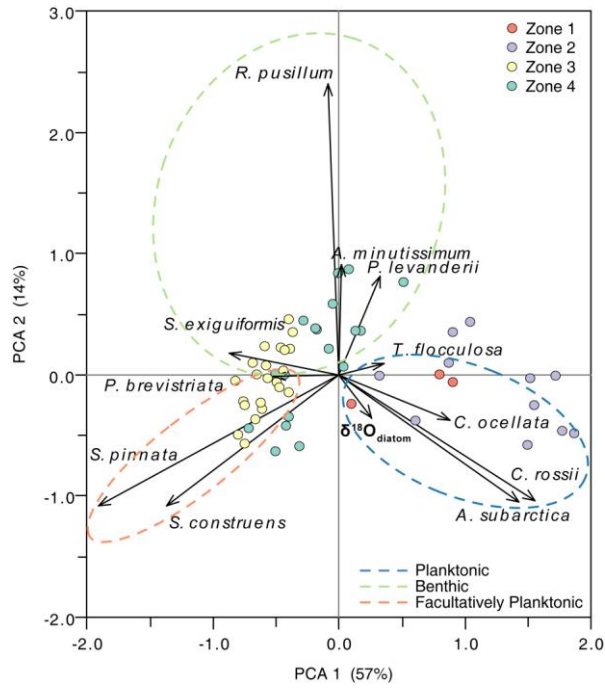
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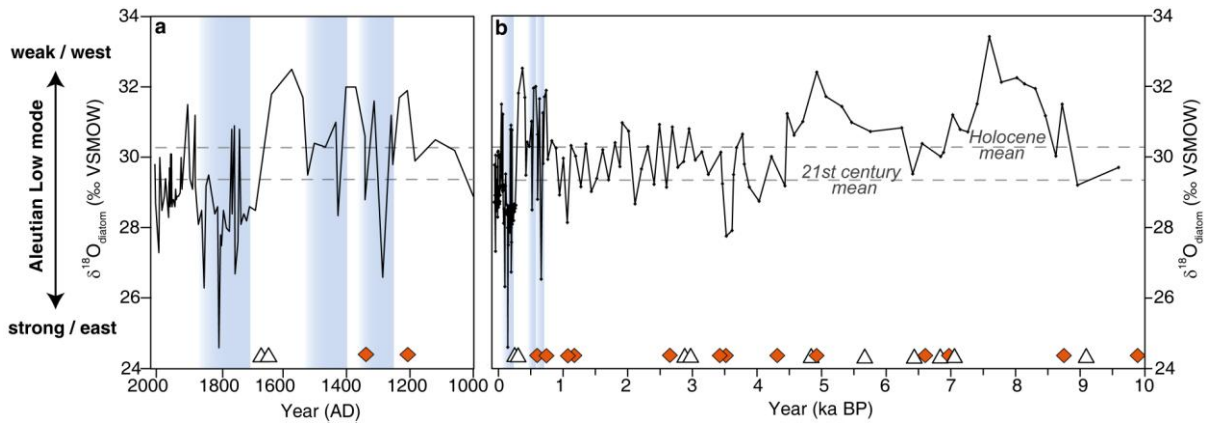


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FIG5



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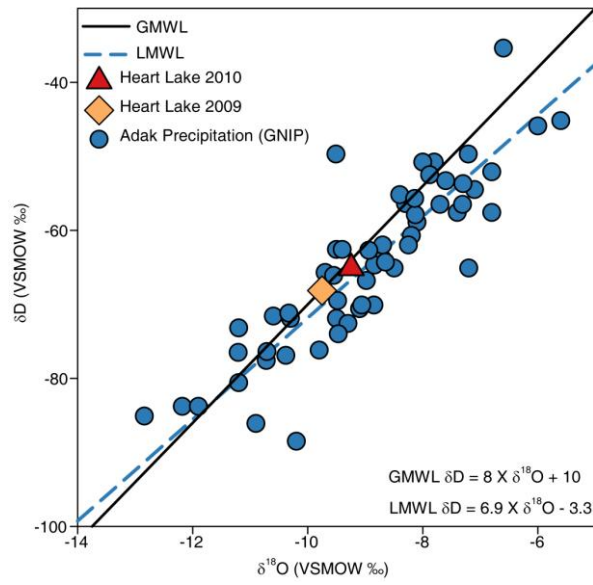
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FIG6



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FIG7

