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1 **Sea ice control on winter subsurface temperatures of the North Iceland**
2 **Shelf during the Little Ice Age: A TEX₈₆ calibration case study**

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25 **Key Points:**

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27 1) Our local Icelandic GDGT calibration produces more realistic temperature
28 estimates with substantially lower uncertainty over broader spatial calibrations
29 2) Periods of thick sea ice during the Little Ice Age likely warmed the subsurface as
30 a result of winter insulation

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33 **Keywords:** North Iceland Shelf, marine sediment, Little Ice Age, lipid biomarkers, IP₂₅,
34 GDGTs

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46 **Abstract**

47 Holocene paleoceanographic reconstructions along the North Iceland Shelf (NIS) have
48 employed a variety of sea-surface temperature (SST) and sea-ice proxies. However, these
49 surface proxies tend to have a seasonal bias toward spring/summer, and thus, only provide
50 a discrete snapshot of surface conditions during one season. Furthermore, SST proxies can
51 be influenced by additional confounding variables resulting in markedly different
52 Holocene temperature reconstructions. Here, we expand Iceland’s marine paleoclimate
53 toolkit with TEX₈₆^L; a temperature proxy based on the distribution of archaeal glycerol
54 dibiphytanyl glycerol tetraether (GDGT) lipids. We develop a local Icelandic calibration
55 from 21 surface sediment samples covering a wide environmental gradient across Iceland’s
56 insular shelves. Locally calibrated GDGT results demonstrate that: 1) TEX₈₆^L reflects
57 winter subsurface (0-200 m) temperatures on the NIS, and 2) our calibration produces more
58 realistic temperature estimates with substantially lower uncertainty (S.E. ±4 °C) over global
59 calibrations. We then apply this new calibration to a high-resolution marine sediment core
60 (last millennium) collected from the central NIS (B997-316 GGC, 658 m depth) with age
61 control constrained by ¹⁴C-dated mollusks. To test the veracity of the GDGT subsurface
62 temperatures, we analyze quartz and calcite wt% and a series of highly branched isoprenoid
63 alkenes, including the sea ice biomarker IP₂₅, from the same core. The sediment records
64 demonstrate that the development of thick sea ice during the Little Ice Age warmed the
65 subsurface due to winter insulation. Importantly, this observation reflects a seasonal
66 component of the sea-ice/ocean feedback to be considered for the non-linear cooling of the
67 Little Ice Age in and around Iceland.

68

69 **1. Introduction**

70 The steep oceanographic (temperature, salinity, and nutrient) gradients caused by the
71 presence of Arctic and Atlantic Ocean currents surrounding Iceland have made the insular
72 shelves targets for northern North Atlantic climate change studies since the last
73 deglaciation (Knudsen et al., 2004; Ólafsdóttir et al., 2010). Throughout the Holocene, the
74 strength and latitudinal position of these currents has varied on centennial timescales,
75 impacting terrestrial climate (Larsen et al., 2012; Geirsdóttir et al., 2013, 2019; Harning et
76 al., 2018), as well as the status of Icelandic ice caps (Larsen et al., 2011; Brynjólfsson et
77 al., 2015; Harning et al., 2016a, 2016b; Anderson et al., 2018). As the North Atlantic is the
78 region that exhibits the largest meridional heat flux of the Northern Hemisphere (Wunsch,
79 1980), and the area of deep-water formation that drives the Atlantic Meridional
80 Overturning Circulation (AMOC), changes in local climate also have widespread
81 hemispheric relevance (Denton & Broecker, 2008; Buckley & Marshall, 2016). Thus,
82 gaining a more comprehensive understanding of the past oceanographic conditions in this
83 region of the North Atlantic is not only key to understanding past episodes of climate
84 change, but also critical to contextualize circulation changes under a currently warming
85 climate (Spielhagen et al., 2011; Caesar et al., 2018; Thornalley et al., 2018).

86 Over recent decades, numerous marine sediment core studies have generated
87 surface and bottom water temperature proxy records based on Mg/Ca and $\delta^{18}\text{O}$ of benthic
88 and planktic foraminifera, calcite wt%, the alkenone unsaturation index (U_{37}^k) and biotic
89 species assemblages, such as dinoflagellates and diatoms (Andersen et al., 2004; Castañeda
90 et al., 2004; Giraudeau et al., 2004; Smith et al., 2005; Solignac et al., 2006; Bendle &
91 Rosell-Melé, 2007; Justwan et al., 2008; Ran et al., 2008; Ólafsdóttir et al., 2010; Jiang et
92 al., 2015; Moossen et al., 2015; Kristjánsdóttir et al., 2016). Sea surface temperature (SST)

93 proxies derived from phytoplankton result in a bias toward spring/summer SST and are
94 influenced by additional confounding variables (i.e., salinity, nutrients, and depth habitat
95 of biota, e.g., [Prah1 et al., 2006](#); [Chival et al., 2014](#)), resulting in markedly different
96 Holocene temperature reconstructions around Iceland ([Kristj1nsd1ttir et al., 2016](#)). As an
97 example, the Little Ice Age (LIA, 1250-1850 CE) is believed to be the coldest multi-
98 centennial climate anomaly of the Holocene in Iceland, yet the coldest Holocene conditions
99 inferred from alkenones ([Kristj1nsd1ttir et al., 2016](#)), dinocysts ([Solignac et al., 2006](#)) and
100 diatoms ([Andersen et al., 2004](#); [Justwan et al., 2008](#)) occur earlier, between 4 and 2 ka.
101 Although the cooling observed in some proxies between 4 and 2 ka may be linked to long-
102 term changes in the AMOC ([Hall et al., 2004](#)) and/or North Atlantic Oscillation ([Orme et
103 al., 2018](#)), expanding Iceland’s quantitative proxy toolkit may help reconcile proxy
104 discrepancies.

105 In this study, we focus on quantifying the distribution of archaeal glycerol
106 dibiphytanyl glycerol tetraethers (GDGTs) archived in marine sediment from the North
107 Iceland Shelf (NIS). Although yet to be used to reconstruct marine paleoclimate on the
108 NIS, GDGT distributions have been shown to reflect modern winter subsurface
109 temperature (subT, 0-200 m) around Iceland ([Rodrigo-G1miz et al., 2015](#)), the North Sea
110 ([Herfort et al., 2006](#)), Skagerrak ([Rueda et al., 2009](#)), and Antarctica ([Kim et al., 2010,
111 2012](#)). Assuming temperature is the dominant control on the distribution of GDGTs on the
112 NIS ([Schouten et al., 2013 and references therein](#)), but acknowledging that at least part of
113 the variability could also be explained by confounding effects such as ammonia oxidation
114 rates ([Hurley et al., 2016](#)), we improve absolute temperature estimates by developing a
115 local Icelandic calibration based on the analysis of surface sediments. We then apply this

116 local calibration to our late Holocene marine sediment core record. A suite of additional
117 oceanographic surface climate proxies from the same core allow us to test the veracity of
118 and to explore controls on GDGTs-based temperatures around Iceland.

119

120 **2. Regional Setting**

121 Today, the NIS represents the boundary where Arctic and Atlantic Ocean currents intercept
122 (Fig. 1a-b, [Stefánsson, 1962](#); [Hopkins, 1991](#); [Belkin et al., 2009](#)). This front separates the
123 cool, low salinity and sea-ice-bearing East Icelandic Current (EIC, 1 to 4 °C) to the north
124 from the warmer and more saline Atlantic waters carried by the North Iceland Irminger
125 Current (NIIC, 5 to 8 °C) on the inner and mid-shelf ([Orvik et al., 2001](#)). The current
126 density differences between the two water masses result in vertical stratification along the
127 NIS, such that the NIIC overlies the denser and cooler Upper Arctic Intermediate Waters
128 (<0 °C, UAIW) (Fig. 1c). Within 70-100 km from Iceland’s northern coastline, freshwater
129 run-off and summer heating modify the NIIC surface waters and form “coastal surface
130 waters” (Fig. 1c), which then disintegrate during the following winter ([Stefánsson, 1962](#);
131 [Ólafsson et al., 2008](#)). The onset of this stratification in early spring triggers the spring
132 bloom of phytoplankton ([Zhai et al., 2012](#)).

133 Atlantic waters provide the primary source of nutrients (i.e., phosphate, nitrate,
134 silica) to the Icelandic shelves. Due to the greater influence of nutrient-deficient polar
135 waters, NIS nutrient concentrations are considerably lower compared to those along the
136 south of Iceland, where Atlantic waters dominate ([Stefánsson, 1968](#); [Stefánsson &](#)
137 [Ólafsson, 1991](#)). Although the freshwater run-off from Iceland is key for the seasonal
138 stratification and phytoplankton blooms along the NIS, it has negligible direct effects on

139 nitrate and phosphate concentrations throughout the water column (Stefánsson & Ólafsson,
140 1991). In terms of modern oxygen saturation, the eastern NIS has similar values to those
141 of waters south of Iceland, which may suggest relatively high rates of productivity for both
142 locations (Stefánsson & Ólafsson, 1991). However, given that the NIS is rather limited in
143 available nutrients, the relatively high oxygen saturation on the NIS may also relate to
144 higher solubility of the colder Arctic waters.

145 Sea ice is also an integral component of the NIS. Iron oxide data on detrital grains
146 suggest that drift ice is predominately sourced from east and southeast Greenland but also
147 from as far as Canada and Russia, with the latter distal sources dramatically increasing in
148 abundance over the last 1 ka (Andrews et al., 2009a; Darby et al., 2017). The presence of
149 and correlation between quartz and the IP₂₅ biomarker – proxies for sea ice extent - in core-
150 top sediment along the NIS, and their absence from Iceland’s southern and western shelves
151 further supports the dominance of drift ice origins over local sea ice production (Axford et
152 al., 2011; Cabedo-Sanz et al., 2016a). When present, sea ice limits the exchange of heat,
153 gases and moisture between the ocean and atmosphere, in addition to insulating the colder
154 polar atmosphere from the relatively warmer ocean during winter (Thorndike et al., 1975;
155 Maykut, 1978, 1982). Due to Iceland’s close proximity to the historical (post-1870 CE) sea
156 ice edge (Divine & Dick, 2007), past changes in sea ice advection along the EIC have
157 resulted in profound changes in local marine and terrestrial climate (Ogilvie & Jónsson,
158 2001; Moros et al., 2006; Massé et al., 2008; Miller et al., 2012; Cabedo-Sanz et al., 2016a).

159

160 **3. Methods**

161 *3.1. Surface and marine core sediments*

162 During July 1997, the cooperative USA/Icelandic *Bjarni Sæmundsson* B997 research
163 cruise visited 30 locations across Iceland’s western and northern shelves ([Helgadóttir,](#)
164 [1997](#)). At each location, marine surface sediments were collected using a grab sampler.
165 Previous studies have used these surface samples to describe the regional distributions of
166 foraminifera $\delta^{18}\text{O}$ ([Smith et al., 2005](#)), quartz wt % ([Andrews & Eberl, 2007](#)), and the sea
167 ice biomarker IP₂₅ ([Axford et al., 2011](#); [Cabedo-Sanz et al., 2016a](#)). We selected a subset
168 (n=11) of these marine surface sediment samples for GDGT analyses to help construct a
169 local Icelandic calibration (Fig. 1). As many of the 30 surface sediment locations were
170 spatially clustered, our selection provides a representative sample from each geographical
171 location the cruise covered, and, thus optimizes our local calibration by spanning the full
172 range of oceanographic conditions present around Iceland today. The B997 cruise also
173 recovered a suite of piston and gravity sediment cores. In this study, we focus on giant
174 gravity core B997-316 GGC (2.47 m long) from the central North Iceland Shelf (66.75°N,
175 18.79°W, 658 mbsl, Fig. 1) ([Helgadóttir, 1997](#)). Sediment (~1 cm³) was subsampled every
176 six cm for mineralogical and biomarker analyses. In order to minimize the degradation of
177 biomarkers (e.g., [Cabedo-Sanz et al., 2016b](#)), samples were taken from cores stored at 4
178 °C. All samples were subsequently freeze-dried and kept in the freezer (-20 °C) prior to
179 biomarker extraction.

180

181 *3.2. Age control*

182 Four radiocarbon-based age control points are derived from a combination of mollusks (*T.*
183 *equalis*) and benthic foraminifera (*N. labradorica* and *G. auriculata arctica*) sampled from
184 the B997-316 GGC core (Table 1). As the B997-316 GGC core lacked datable material in

185 the uppermost sediment, two additional mollusks (*T. equalis*) were sampled from the near
186 surface sediment of an adjacent short gravity core, B997-316 SGC (Table 1), to confirm
187 that the tops of the sediment cores are modern and that no surface sediment was lost during
188 coring. Samples were prepared for AMS radiocarbon dating at the Institute of Arctic and
189 Alpine Research (INSTAAR) ¹⁴C Preparation Lab and analyzed at the University of
190 California Irvine.

191

192 *3.3. Mineralogical analyses*

193 Quantitative X-ray diffraction (qXRD) analysis was conducted on the <2 mm sediment
194 fraction using the method developed by Eberl (2003) and used extensively on sediment
195 samples on other B997- cores (Andrews & Eberl, 2007; Andrews et al., 2009a; Andrews,
196 2009). Comparison between qXRD weight percent estimates on known mineral mixtures
197 and replicate analyses indicate that the errors on the weight percent estimates of both are
198 in the range of ± 1 %. For B997-316 GGC, we focus on the identification of quartz as a
199 proxy to reconstruct the incursion of drift ice into Icelandic waters (i.e., sea ice and/or
200 icebergs), and calcite as a proxy of ocean productivity (Andrews et al., 2001).

201

202 *3.4. Biomarker analyses*

203 At the University of Plymouth, freeze-dried subsamples (~1-2 g) from core B997-316 GGC
204 were extracted for biomarkers by ultrasonication using dichloromethane:methanol (2:1,
205 v/v). Samples were initially spiked with an internal standard (9-octylheptadec-8-ene, 9-
206 OHD, 10 μ L; 10 μ g mL⁻¹) to permit quantification of highly branched isoprenoid (HBI)
207 alkenes. Total lipid extracts (TLEs) were separated into three fractions (F1-F3) using silica

208 column chromatography, after elution with hexane (6 mL), hexane:methylacetate (80:20,
209 v/v, 6 mL), and methanol (4 mL), respectively. The F1 fraction contained aliphatic
210 hydrocarbons including highly branched isoprenoids (HBIs; i.e., IP₂₅, C_{25:2} and C_{25:3}),
211 whereas F2 contained lipids with hydroxyl functional groups, including GDGTs. At the
212 University of Colorado Boulder, freeze dried marine surface sediment samples (~3-7 g)
213 were extracted three times on a Dionex accelerated solvent extractor (ASE 200) using
214 dichloromethane:methanol (9:1, v/v) at 100 °C and 2000 psi, and kept as TLEs for the
215 GDGT analysis.

216 The IP₂₅ (C_{25:1}), diene II (C_{25:2}) and triene Z (C_{25:3}) biomarkers were analyzed at the
217 University of Plymouth as described by [Belt et al. \(2012, 2015\)](#). Analysis of the F1 was
218 performed via gas chromatography-mass spectrometry (GC-MS) following the methods
219 and operating conditions of [Belt et al. \(2012\)](#) on an Agilent 7890A GC coupled to a 5975
220 series mass selective detector fitted with an Agilent HP-5ms column (30 m x 0.25 mm x
221 0.25 mm). Mass spectrometric analyses were carried out in selected ion monitoring mode.
222 The identification of IP₂₅ ([Belt et al., 2007](#)), diene II ([Belt et al., 2007](#)) and triene Z ([Belt](#)
223 [et al., 2000](#)) was based on their characteristic GC retention indices (RI_{HP5MS} = 2081, 2082,
224 and 2044 for IP₂₅, diene II, and triene Z, respectively) and mass spectra ([Belt, 2018](#)).
225 Quantification of lipids was achieved by comparison of mass spectral responses of selected
226 ions (IP₂₅, *m/z* 350; diene II, *m/z* 348; triene Z, *m/z* 346) with those of the internal standard
227 (9-OHD, *m/z* 350) and normalized according to their respective response factors and
228 sediment masses ([Belt et al., 2012](#)). Analytical reproducibility was monitored using a
229 standard sediment with known abundances of biomarkers for every 14-16 sediment
230 samples extracted (analytical error 4%, n = 31).

231 For GDGTs, we analyzed aliquots of the F2 from B997-316 GGC and aliquots of
232 the TLE from marine surface sediment samples in the Organic Geochemistry Laboratory
233 at the University of Colorado Boulder. Dry samples were dissolved in hexane:isopropanol
234 (99:1, v/v), sonicated, vortexed, and then filtered using a 0.45 µm PTFE syringe filter. Prior
235 to analysis samples were spiked with 10 ng of the C₄₆ GDGT internal standard (Huguet et
236 al., 2006). Isoprenoid GDGTs were identified and quantified via high performance liquid
237 chromatography – mass spectrometry (HPLC-MS) following modified methods of
238 Hopmans et al. (2016) on a Thermo Scientific Ultimate 3000 HPLC interfaced to a Q
239 Exactive Focus Orbitrap-Quadrupole MS. Rather than starting at 18% hexane:isopropanol
240 (9:1, v/v) (Hopmans et al., 2016), we began our eluent gradient with 30%
241 hexane:isopropanol (9:1, v/v) to shorten retention and overall run times without
242 compromising the chromatographic separation of GDGTs. The HPLC was conditioned for
243 20 minutes between runs. Samples were analyzed on full scan mode with a mass range of
244 500-1500 *m/z* at 70,000 mass resolution. GDGTs were identified based on their
245 characteristic masses and elution patterns. We adopt the TEX₈₆^L index to reflect relative
246 changes in temperature, which is a modification of the original TEX₈₆ index (Schouten et
247 al., 2002) constructed for temperatures <15 °C (Kim et al., 2010, 2012):

$$248 \quad \text{TEX}_{86}^L = \log\left(\frac{[\text{GDGT} - 2]}{[\text{GDGT} - 1] + [\text{GDGT} - 2] + [\text{GDGT} - 3]}\right)$$

249

250 3.5. Local Icelandic TEX₈₆^L calibration

251 The largest uncertainty in the temperature relationship of GDGTs in global calibrations is
252 at the low end of the temperature spectrum (<5 °C, Kim et al., 2010, 2012), which may
253 reflect a reduced sensitivity of Thaumarchaeota to temperature in cooler climates (Wuchter

254 [et al., 2004](#)) or different regional oceanographic effects. Although a recent spatially-
255 varying, Bayesian TEX₈₆-temperature calibration model was developed to capture regional
256 oceanography variability (Tierney & Tingley, 2014), it excludes high-latitude settings and
257 does not include any core top samples within a ~1000-km radius of Iceland. Hence, we
258 targeted a network of local marine surface sediments (Fig. 1) to develop a local calibration
259 that innately reflects the nuances of Icelandic oceanography and low local temperatures.
260 We supplemented our 11 surface sediment samples with 10 previously published surface
261 sediment samples from around Iceland (Table 2, [Rodrigo-Gámiz et al., 2015](#)) to generate
262 a more comprehensive GDGT calibration that spans a larger geographical area and
263 temperature gradient than obtainable using B997 samples alone. We note that although
264 [Rodrigo-Gámiz et al. \(2015\)](#) sampled the surface of sediment cores, our samples were
265 collected used a grab sampler, which may disturb the original sedimentary structure.
266 However, natural factors such as sea floor mixing and variable sedimentation rates will
267 always introduce uncertainty in the temperature embedded no matter how the sample is
268 collected. Thus, we contend that the datasets can be merged for calibration purposes.

269 To calibrate the TEX₈₆^L index, *in situ* decadal mean temperatures from 1995-2004
270 CE were obtained from the World Ocean Atlas (WOA09, [Locarnini et al., 2010](#)) at the
271 quarter-degree pixel where each surface sediment site is located. Subsequently, seasonal
272 (spring, summer, fall, winter) and annual SST, in addition to 0-10 m, 0-20 m, 0-30 m, 0-
273 40 m, 0-50 m, 0-60 m, 0-70 m, 0-80 m, 0-90 m, 0-100 m, 0-125 m, 0-150 m, 0-175 m, 0-
274 200 m depth subsurface temperature integrations, were each regressed against the 21 core
275 top TEX₈₆^L index values ([Rodrigo-Gámiz et al., 2015](#); [this study](#)) to assess which portion

276 of the water column and which season the GDGT distributions most closely correspond to
277 around Iceland. We calculated p -values for each regression to determine their significance.

278

279 **4. Results and Interpretations**

280 *4.1. Age model*

281 An age model for the B997-316 GGC sediment core was generated in the CLAM software
282 using the Marine13 calibration curve ($\Delta R=0$, [Reimer et al., 2013](#)) and a smooth spline
283 regression over 1000 iterations ([Blaauw, 2010](#)). The calibrated benthic foraminifera date
284 from 49.5 cm depth produced an age reversal in the initial model, and thus, was identified
285 as an outlier and removed from the final age model (Fig. 2). The ~400-year difference
286 between the calibrated age of the foraminifera and that estimated from the model may relate
287 to changes in ΔR resulting from variable water masses ([Eiríksson et al., 2004](#); [Wanamaker](#)
288 [et al., 2012](#)). The two mollusks from the adjacent short gravity core (B997-316 SGC) both
289 returned conventional ^{14}C ages ≤ 400 years (Table 1), confirming modern sediment at the
290 core top of the SGC.

291 Based on several lines of reasoning, we argue that the modern ages of the SGC can
292 be used to validate the extrapolation of the B997-316 GGC age model to the surface (Fig.
293 2). First, given that both cores were collected from the same location, we can exclude any
294 impacts from geographic-dependent factors, such as variable sedimentation rates or
295 oceanographic currents, that may cause the age-depth relationships to differ between the
296 two cores. Second, coring-dependent factors can also be excluded as the two cores were
297 collected in succession of each other using the same equipment ([Helgadóttir, 1997](#)). Even
298 though coring operations can result in the loss of saturated or poorly consolidated surface

299 sediment, the identical coring techniques employed suggests that if the SGC captures
300 modern surface sediment, so should the GGC.

301 Given that our age model only uses the three lowermost mollusks from the GGC
302 and the uppermost mollusk samples from the adjacent SGC, increased age uncertainty
303 undoubtedly exists where no datable material could be obtained (i.e., ~1400-2000 CE).
304 However, we argue that our age estimates throughout the entire GGC core are reasonably
305 strong. First, NIS sedimentation rate slopes only change significantly between
306 deglacial/nonglacial periods ([Andrews et al., 2002](#); [Xiao et al., 2017](#)) due to the
307 presence/absence of the Icelandic Ice Sheet. Following the rapid demise of the Icelandic
308 Ice Sheet ~15 thousand years ago ([Norðdahl and Ingólfsson, 2015](#); [Patton et al., 2017](#)),
309 sedimentation rates have remained linear across the NIS ([Castañeda et al., 2004](#)), consistent
310 with the linearity of B997-316 GGC’s sedimentation rate over the last millennium (Fig. 2).
311 Second, when our age model is applied to the proxy datasets (Fig. 3), the interpreted period
312 of the Little Ice Age (LIA, see following section) is consistent with previous age ranges in
313 Iceland that are derived from high-resolution and precisely-dated terrestrial archives
314 ([Geirsdóttir et al., 2009](#); [Larsen et al., 2011, 2012](#)).

315

316 *4.2. Sediment core B997-316 GGC*

317 *4.2.1. Minerological analyses*

318 In Icelandic waters, the two minerals quartz and calcite are qualitative indicators that reflect
319 the incursion of drift ice (i.e., sea ice and/or icebergs) and marine surface productivity,
320 respectively ([Andrews et al., 2001, 2009a](#)). In years when cold low-salinity Arctic water
321 dominates, sea ice (% quartz) increases and surface productivity (% calcite) decreases due
322 to the development of a well stratified water column. The opposite is seen in the proxies

323 during years characterized by warm and saline Atlantic waters, which reduces sea ice
324 presence and mixes the water column resulting in higher productivity. Not surprisingly, %
325 quartz and calcite generally show an inverse relationship over the last millennium in B997-
326 316 GGC, which can be interpreted as the relative dominance of Arctic versus Atlantic
327 waters at this location (Fig. 3a-b).

328 Percent quartz ranges from 1.4 to 2.7 %, whereas calcite ranges from 5.4 to 8.3 %
329 (Fig. 3a-b). Recent analyses using mineral mixtures with known quartz wt % of 3.5 and 1.5
330 (Andrews et al., 2018) confirm that these small amounts of quartz can be correctly
331 measured. Prior to ~1250 CE, quartz is relatively low, and calcite is the highest of the
332 record, suggesting a dominance of warmer Atlantic waters at this time. Subsequently,
333 quartz begins a gradual yet quasi-episodic rise towards its peak abundance at ~1900 CE.
334 On the other hand, calcite appears to decline more sharply to lower values after ~1250 CE
335 and remain relatively low through ~1900 CE, when it rises to levels near its pre-1250 CE
336 state. Based on these two minerals, the period between 1250 and 1900 CE was likely
337 characterized by cooler Arctic waters that favored the advection of drift ice, vertical
338 stratification and lower surface productivity on the NIS. Following 1900 CE, the conditions
339 reverted back to a dominance of warmer Atlantic waters that favored restricted sea drift
340 transport and higher surface productivity (Fig. 3).

341 4.2.2. *Highly-branched isoprenoid (HBI) alkenes*

342 The analysis of the biomarker IP₂₅ (Belt et al., 2007), a monounsaturated C₂₅ HBI
343 biosynthesized by Arctic sea ice diatoms (Belt et al., 2008; Brown et al., 2014), has gained
344 recent traction as a novel proxy for spring/summer sea ice conditions around Iceland
345 (Massé et al., 2008; Andrews et al., 2009b; Sicre et al., 2013; Cabedo-Sanz et al., 2016a;

346 [Xiao et al., 2017](#)). Although the IP₂₅ biomarker is well-preserved in Arctic and sub-Arctic
347 marine sediment and routinely applied in paleo sea ice reconstructions as old as 5.3 Ma
348 ([Stein et al., 2016](#)), questions remain regarding its vertical transport, degradation processes,
349 and environmental controls (see reviews by [Belt & Müller, 2013](#); [Belt, 2018](#)). Notably, the
350 interpretation of its presence (or lack thereof) can be ambiguous. IP₂₅ below the limit of
351 detection has often been interpreted as reflecting either a lack of seasonal sea ice cover, or
352 permanent and thick sea ice that inhibits light penetration needed for sea ice diatoms to
353 photosynthesize and grow. However, this is likely an over-simplification of a broader range
354 of scenarios that result in absent IP₂₅ ([Belt, 2018](#)). In any case, further information may be
355 obtained by the complementary analysis of certain open-water phytoplankton biomarkers
356 (i.e., brassicasterol or dinosterol, [Müller et al., 2011](#)).

357 Based on a distinctively heavy stable carbon isotopic composition, in addition to
358 similar concentration profiles to IP₂₅ across Arctic marine surface sediment, the di-
359 unsaturated HBI diene II also has an Arctic sea ice diatom source ([Belt et al., 2008](#); [Cabedo-
360 Sanz et al., 2013](#); [Brown et al., 2014](#)), and is made by some Antarctic sea ice algae as well
361 ([Belt et al., 2016](#)). In contrast, a tri-unsaturated HBI (hereafter, triene Z) is biosynthesized
362 by certain open-water diatoms ([Belt et al., 2000, 2008, 2015](#); [Rowland et al., 2001](#)), and
363 sources for the Arctic and Antarctic have recently been identified ([Belt et al., 2017](#)).
364 Importantly, the presence (or lack thereof) of triene Z, like certain phytoplankton sterols,
365 may help differentiate between open-water or thick sea ice conditions inferred from IP₂₅
366 and diene II in the Arctic ([Cabedo-Sanz et al., 2013](#); [Smik et al., 2016](#); [Köseoğlu et al.,
367 2018](#)). However, since sterols may also be derived from other (e.g., terrestrial) sources in

368 addition to sea ice algae (Huang & Meinschein, 1976; Volkman, 1986; Volkman et al.,
369 1998; Belt et al., 2013, 2018), we limit our analysis here to triene Z only.

370 HBIs were detected in all downcore samples, with the exception of 96.5 cm depth
371 (1509 CE), where no triene Z, diene II or IP₂₅ were detected (Fig. 3c-e). Concentrations
372 ranged from near detection up to 1.6 ng/g sediment for triene Z, up to 19 ng/g sediment for
373 diene II, and up to 4 ng/g sediment for IP₂₅. Triene Z exhibited the highest concentrations
374 prior to 1200 CE, while its abundance diminished to very low or undetectable between
375 ~1200 and 1800 CE (Fig. 3c). Triene Z then rises to higher concentrations up through 2000
376 CE. The similar relative trends of diene II and IP₂₅ concentrations suggest that both HBIs
377 are likely sourced from sea ice algae around Iceland, similar to other Arctic (Brown et al.,
378 2014) and Antarctic locations (Collins et al., 2013). Periods of synchronous reductions of
379 diene II and IP₂₅ concentrations occur at ~1170-1290 CE, 1450-1650 CE, and 1880 CE-
380 present.

381 The similar overall trends between % calcite and triene Z abundance suggest that
382 both proxies indicate temperate water surface productivity (Fig. 3b-c). Hence, in years
383 where warmer Atlantic waters dominate, both % calcite and triene Z abundance increase,
384 while the opposite trend dominates during years characterized by cooler Arctic waters. The
385 detection of both IP₂₅ and % quartz throughout the record suggests that sea ice has been a
386 persistent feature at this location of the NIS over the last millennium, even during intervals
387 when elevated % calcite and triene Z suggest an increased influence of warmer Atlantic
388 waters. We interpret the reduction of diene II and IP₂₅ at ~1170-1290 CE and 1880 CE-
389 present to reflect diminished seasonal sea ice because of higher concentrations of triene Z

390 at the same time. In contrast, the reduction of diene II and IP₂₅ from 1450-1650 CE likely
391 reflects a period of thick, perennial sea ice as triene Z was mostly undetectable (Fig. 3c-e).

392 4.2.3. *Glycerol dibiphytanyl glycerol tetraethers (GDGTs)*

393 Changes in the degree of cyclization (number of cyclopentane moieties) in GDGTs have
394 classically been interpreted to represent a physiological response of marine ammonia
395 oxidizing Thaumarchaeota to changes in *in situ* temperature (e.g., [Schouten et al., 2002](#)).

396 Thus, the TEX₈₆ paleothermometer index has been empirically linked to annual or winter
397 subT (0-200 m depth) in global data sets ([Schouten et al., 2002](#); [Kim et al., 2010, 2012](#)).

398 This presumption is supported by a recent study along a latitudinal transect in the western
399 Atlantic Ocean, which demonstrated that the most likely water depths where GDGTs are
400 produced from and exported to marine sediment is around 80-250 m ([Hurley et al., 2018](#)),

401 similar to evidence for archaea abundance maxima at 200 m depths in the Pacific Ocean
402 ([Karner et al., 2001](#)). Considering that Thaumarchaeota are chemolithoautotrophs that

403 perform ammonia oxidation (conversion of NH₄⁺ to NO₂⁻), they are typically more
404 abundant around the primary NO₂⁻ maximum near the base of the photic zone ([Francis et](#)

405 [al., 2005](#); [Church et al., 2010](#); [Hurley et al., 2018](#)), and are thus most productive when there
406 is minimized phytoplanktic competition over NH₄⁺ ([Schouten et al., 2013](#)). In the case of

407 the Arctic region, the latter occurs during the less productive winter months when
408 photosynthesis for sea surface species is inhibited, which may explain the seasonal winter

409 temperature bias of GDGTs previously observed in this region ([Rodrigo-Gámiz et al.,](#)
410 [2015](#)). However, in addition to subT, recent studies have shown that several other

411 environmental and geochemical factors can influence the degree of cyclization, such as
412 growth phase ([Elling et al., 2014](#)), ammonia oxidation rates ([Hurley et al., 2016](#)), and

413 oxygen concentrations (Qin et al., 2015). In contrast to some other marine temperature
414 proxies, such as the $\delta^{18}\text{O}$ of planktic foraminifera, GDGTs do not seem to be influenced by
415 variations in salinity (Wuchter et al., 2004, 2005; Elling et al., 2015). Finally, GDGTs
416 appear to be relatively resistant to oxic degradation (Schouten et al., 2004), and thus, likely
417 reflect original living conditions once deposited in the sedimentary record.

418 GDGTs were present above the detection limits in all marine sediment core
419 samples, and substantially increase in concentration at the core top (Supplement Fig S1).
420 The calculated $\text{TEX}_{86}^{\text{L}}$ index ranged from -0.71 to -0.63 (Fig. 3f). The record displays high
421 variability and a rather constant first order trend towards the present, in addition to the
422 occurrence of two intervals of substantial decreases in $\text{TEX}_{86}^{\text{L}}$ values during 1350-1530
423 CE and 1745-1975 CE. Both periods are preceded by periods of relatively higher $\text{TEX}_{86}^{\text{L}}$
424 values during 1110-1350 CE and 1530-1745 CE, respectively. A full paleoceanographic
425 interpretation of the $\text{TEX}_{86}^{\text{L}}$ results is provided in the discussion.

426

427 *4.3. Local Icelandic $\text{TEX}_{86}^{\text{L}}$ calibration*

428 GDGTs were also detected and above detection limits in all B997 marine surface sediments
429 samples (n=11, Supplemental Fig S2). $\text{TEX}_{86}^{\text{L}}$ values of these samples ranged from -0.72
430 to -0.61 (Table 2). The 10 marine surface sediment samples from Rodrigo-Gámiz et al.
431 (2015) span a greater geographical and environmental range around Iceland, and hence
432 exhibit a greater range of $\text{TEX}_{86}^{\text{L}}$ values (-0.71 to -0.49, Table 2). When we use the
433 combined set of Icelandic marine surface sediment samples (n=21), the regression analysis
434 demonstrates that the integration of winter temperatures from the top 200 m of the water
435 column provides the best regression coefficients ($R^2=0.73$, $p<0.001$) compared to the

436 integration of other seasonal temperatures and the mean annual value (Figs. 4). Thus,
437 sedimentary values around Iceland most likely represent winter subT that integrate a signal
438 of the uppermost 200 m of the water column, consistent with the findings of Rodrigo-
439 Gámiz et al. (2015).

440

441 **5. Discussion**

442 *5.1. Local Icelandic TEX_{86}^L vs. regional Arctic calibration*

443 If we supplement the combined Icelandic data set with more marine surface sediment
444 samples from the greater northern North Atlantic region (Kim et al., 2010), the correlation
445 coefficients of our winter subT (0-200 m) regression is substantially reduced ($R^2 = 0.43$ vs.
446 0.73 , Supplemental Fig S3). This suggests that a local Icelandic calibration is optimal over
447 larger regional calibrations, and perhaps, more accurately captures the nuances of local
448 Icelandic oceanography. We hypothesize that the poorer performance of a more regional
449 GDGT calibration for the North Atlantic region may relate to the inclusion of: 1) surface
450 sediment samples from distal locations that feature different oceanographic environments
451 than Iceland (e.g., Hudson Bay), and/or 2) samples from higher latitude (e.g., Svalbard and
452 the Barents Sea) that are less “responsive” in terms of GDGT cyclization as they fall under
453 the colder end of the spectrum in the global TEX_{86}^L calibration, which is characterized by
454 a higher uncertainty and deviation from linearity (Kim et al., 2010). The standard error in
455 our Icelandic winter subT calibration for 0-200 m (± 0.4 °C), is also an order of magnitude
456 lower than the error derived from global low temperature calibrations (e.g., 4.0 °C, Kim et
457 al., 2010; 2.8 °C, Kim et al., 2012). The reduced uncertainty achieved in our Icelandic
458 calibration highlights the growing need for the continued development and application of

459 regional calibrations in future biomarker-based paleoclimate reconstructions (e.g., [Kaiser](#)
460 [et al., 2015](#); [Foster et al., 2016](#); [Russell et al., 2018](#)). This is particularly important in areas
461 where the temperature relationship of GDGTs deviates from the overall linear correlation
462 observed in global calibrations (i.e., cold and warm regions).

463 Despite the reduced uncertainty compared to global calibrations, the regression
464 coefficient for the Icelandic winter subT calibration ($R^2=0.73$) is comparatively lower than
465 the global calibration ($R^2=0.86-0.87$; [Kim et al., 2010, 2012](#)). We hypothesize that the
466 unconstrained confounding influence of ammonia-oxidation on the degree of GDGT
467 cyclization (e.g., [Hurley et al., 2016](#)) may contribute to the scatter of our dataset (Fig. 4b).
468 Although specific ammonia (NH_4^+) and nitrite (NO_2^-) information for this region is
469 currently unavailable, reduced (enhanced) ammonia oxidation rates in the water column
470 throughout the year would result in increased (decreased) degree of cyclization, thus
471 yielding higher and lower temperatures, respectively ([Hurley et al., 2016](#)). If ammonia
472 oxidation rates are driven by changes in ammonia supply and utilization (e.g., reduced
473 nutrient availability in Arctic waters, or competition with phytoplankton), we cannot
474 separate the influence of nutrient variability on the Icelandic $\text{TEX}_{86}^{\text{L}}$ values with our
475 current dataset. While oxygen availability has also been shown to influence the degree of
476 cyclization in GDGTs ([Qin et al., 2015](#)), this factor is unlikely to affect the distribution of
477 GDGTs around Iceland as these waters are relatively well-mixed and ventilated today
478 ([Stefánsson & Ólafsson, 1991](#)), and presumably have been since the early Holocene
479 ([Kristjánsson et al., 2016](#)). With all known controlling factors considered, we suggest
480 that our local $\text{TEX}_{86}^{\text{L}}$ calibration improves the temperature estimates for Icelandic winter

481 subsurface waters. Future work in constraining the effects of ammonia-oxidation around
482 Iceland would undoubtedly benefit the application of TEX₈₆^L on the NIS.

483 By applying temperature calibrations to our down core TEX₈₆^L record, our data
484 reveal rapid and abrupt temperature variability on the NIS during the last millennium (Fig.
485 5). If the existing annual SST (Kim et al., 2010) and annual subT TEX₈₆^L calibrations
486 developed for polar regions (Kim et al., 2012) are applied, the GDGT distributions suggest
487 that subT fluctuated up to 5 °C over the course of decades. These observations are
488 considerably higher than expected, especially given that they are comparable to the
489 magnitude of SST changes observed in other NIS proxy records over the entire Holocene
490 (e.g., Andersen et al., 2004; Bendle & Rosell-Melé, 2007; Jiang et al., 2015; Kristjánsson
491 et al., 2016). As originally hypothesized, this exercise demonstrates that global calibrations
492 that feature greater uncertainty for low temperatures and that do not include sites proximal
493 to Iceland are not appropriate for the NIS. In contrast, by applying our local winter subT
494 calibration, the magnitude of estimated subT is not only reduced to ranges more
495 comparable to other proxy records for the last millennium but, importantly, also captures
496 the modern instrumental winter subT (within calibration uncertainty) at the B997-316 GGC
497 site (4 °C, Fig. 5), further reinforcing the application of our local Icelandic TEX₈₆^L
498 calibration.

499

500 *5.2. NIS surface and subsurface climate variability during the Little Ice Age*

501 The NIS represents one of the few global examples where paleo-IP₂₅ abundance in marine
502 cores has been calibrated against observational and documentary records (Massé et al.,
503 2008; Andrews et al., 2009b). As a result, the variability of IP₂₅ has been routinely applied

504 to marine sediment around Iceland as a robust indicator for seasonal sea ice (Massé et al.,
505 2008; Andrews et al., 2009b; Sicre et al., 2013; Cabedo-Sanz et al., 2016a). Similar to these
506 previous studies, IP₂₅ concentrations in B997-316 GGC increase abruptly during the 13th
507 century, and with the exception of the period 1450-1650 CE, remain elevated until the 19th
508 century when concentrations begin to diminish (Fig. 6a). By employing statistical analyses
509 on IP₂₅ abundances and 11 other marine climate proxy datasets from marine sediment core
510 MD99-2263, Andrews et al. (2009b) showed that a major regime shift in the marine climate
511 off NW Iceland commenced after 1200 CE, possibly linked to a strengthening high-
512 pressure ridge over Greenland in winter/spring that favored stronger north/northwesterly
513 winds and increased drift ice export to Iceland. Our mineral and HBI records consistently
514 reflect major shifts in surface conditions at a similar time and in the same direction (Fig.
515 3a-e), reinforcing the observed regime shift in marine climate, and increase of sea ice in
516 particular, during the 13th century (Bergthórsson, 1969; Ogilvie & Jónsson, 2001; Massé
517 et al., 2008; Andrews et al., 2009b; Sicre et al., 2013; Cabedo-Sanz et al., 2016a). The
518 consistency of our surface productivity and sea ice proxy records in reflecting the
519 established understanding of marine climate over the last millennium on the NIS supports
520 the fidelity of the B997-316 GGC marine sediment record, and therefore, the interpretation
521 of the GDGT record.

522 When the GDGT record is converted to winter subT, two pronounced centennial-
523 scale cold anomalies exhibit mean winter subT below the record average of 4.34 °C; at
524 1350-1530 CE (3.99 °C) and at 1745-1975 CE (4.19 °C). These two cold anomalies are
525 consistent with low surface productivity (% calcite and triene Z) and increased seasonal
526 sea ice (% quartz, diene II, IP₂₅), which suggest greater dominance of colder Arctic surface

527 waters between ~1250 and 1900 CE (Fig. 6a-b). In addition, alkenone-derived SST from
528 marine core MD99-2275 50 km to the east (Fig. 1a) document steady cooling throughout
529 this interval (Fig. 6c, [Sicre et al., 2011](#)), further supporting the presence of cool, Arctic
530 surface waters in the vicinity of B977-316 at this time. However, the timing for the onset
531 (1350 CE) and termination (1975 CE) of LIA cooling observed in the subsurface during
532 winter appears to lag that of the surface (1250 and 1900 CE, respectively) (Fig. 6). A
533 variety of model and data-based studies have demonstrated that the LIA was triggered by
534 a combination of sustained stratospheric volcanic sulfate injection ([Zhong et al., 2010](#);
535 [Miller et al., 2012](#); [Sicre et al., 2013](#); [Slawinska & Robock, 2018](#)), low total solar irradiance
536 ([Shindell et al., 2001](#)) and changes in the North Atlantic Oscillation, one of the major
537 modes of internal climate variability in the North Atlantic ([Trouet et al., 2009](#)). On the NIS,
538 these radiative forcings directly impact the ocean surface, as manifested in the immediate
539 and abrupt increase in seasonal sea ice, reduced northward heat transport and suppression
540 of SSTs ([Miller et al., 2012](#)). The phase relationship between the B997-316 GGC surface
541 proxies and GDGTs suggests that it may have taken up to a century for the radiative forcing
542 in contact with the surface to propagate to the subsurface.

543 The subsurface warming observed between the two subsurface cold anomalies
544 (~1530-1745 CE, Fig. 6d) may suggest a reduced influence of the colder Arctic water mass
545 that generally characterized the LIA. However, similar to our surface proxies (Fig. 6a-b)
546 and the alkenone-derived SST from MD99-2275 (Fig. 6c), a local sclerochronological ^{14}C
547 record (ΔR_{shell}) constructed from mollusk shells reflects the continued dominance of older
548 Arctic waters in the benthos as well (Fig. 6e, [Wanamaker et al., 2012](#)). In fact, between
549 1450-1650 CE, the combination of our IP₂₅ and triene Z datasets suggest thicker and more

550 permanent sea ice above the B997-316 site (Fig. 6a-b), an interpretation supported by
551 additional LIA sea ice proxy (IP₂₅ and quartz) records from the NIS (Massé et al., 2008;
552 Andrews et al., 2009b; Cabedo-Sanz et al., 2016a). If thick sea ice conditions are
553 maintained throughout the year, the insulating effects of sea ice would warm the subsurface
554 waters during winter, as reflected by our GDGT record (Fig. 6d). The thickening of sea ice
555 that we observe at 1450 CE coincides with the local intensification of LIA terrestrial
556 cooling manifested in the synchronous advance of local Icelandic ice caps (Larsen et al.,
557 2011; Harning et al., 2016a), and reduced Icelandic lake productivity (Geirsdóttir et al.,
558 2013, 2019; Harning et al., 2018). A previous data-modeling comparison showed that this
559 LIA intensification was likely forced by another episode of high stratospheric sulfate
560 loading from explosive tropical volcanism and sustained by sea-ice/ocean feedbacks that
561 reflected incoming solar radiation during summer (Miller et al., 2012). We suggest that the
562 winter sea ice insulation may be another important component of the sea-ice/ocean
563 feedback to consider in the non-linear nature of Little Ice Age cooling around Iceland.

564 Following the dissipation of thick sea ice conditions at 1650 CE, rising IP₂₅ and
565 triene Z concentrations suggest the return to seasonal sea ice conditions that favored the
566 co-productivity of sea ice and open water algae at the B997-316 site (Fig. 6a-b). The change
567 in sea ice conditions is the likely mechanism for the return of lower subT at 1745 CE,
568 inferred from low GDGT temperature anomalies (Fig. 6d). We hypothesize that as the sea
569 ice thinned out during spring months, possibly spurred by the previously stored subsurface
570 heat, the winter ice pack would have also thinned accordingly. Consequently, a more open
571 winter sea ice pack would have facilitated increased heat flux from the ocean to the colder
572 overlying atmosphere, as reflected by the lower GDGT-based subsurface temperatures.

573

574 **6. Conclusion**

575 Consistent with the community’s growing comprehension of GDGT-based temperature
576 records at high latitudes, we show that archaeal isoprenoid GDGT distributions ($\text{TEX}_{86}^{\text{L}}$)
577 around Iceland predominately reflect winter subsurface temperatures (0-200 m).
578 Furthermore, by developing a local calibration based on a network of surface sediment
579 samples, reconstructed NIS subsurface temperature estimates and uncertainty are improved
580 upon those obtained from regional and global calibrations. Our $\text{TEX}_{86}^{\text{L}}$ subsurface
581 paleotemperature record from the NIS captures the cooling likely associated with the LIA
582 (1250-1900 CE), as seen in additional surface proxies (sea ice and marine productivity)
583 from the same sediment core. However, the LIA onset, intensification, and termination of
584 the subsurface lags those changes of the surface, suggesting that it may have up to a century
585 for changes at the surface to propagate to the subsurface during the late Holocene. We
586 propose that the development of thick sea ice conditions during the intensification of the
587 LIA around 1450 CE insulated the subsurface in winter, resulting in apparently warmer
588 seasonal subsurface waters. This mechanism likely represents another seasonal component
589 of the sea-ice/ocean feedback to be considered in the abrupt cooling manifested in and
590 around Iceland during the LIA.

591

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604

605

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1105

1106 **TABLES and FIGURES (8 units):**

1107

1108 **Table 1:** Radiocarbon information. ¹⁴C ages calibrated in Calib 7.1 (Stuiver et al., 2018) using the
1109 MARINE13 calibration curve (Reimer et al., 2013) and a ΔR of 0. Note that ¹⁴C and calibrated
1110 ages are presented as BP (Before Present) in this table, and as CE (Common Era) in the main text.

Sediment core	Sediment depth (cm)	Lab ID	Material	δ ¹³ C (‰)	Conventional ¹⁴ C date BP ± σ	ΔR	Calibrated age BP ± σ
B997-316 SGC	7.5	GRL-1691-S	mollusk (<i>T. equalis</i>)	0.62	294 ± 91	0	<400
B997-316 SGC	18	GRL-1690-S	mollusk (<i>T. equalis</i>)	-7.2	402 ± 38	0	<400
B997-316 GGC	49.5	CURL-18624	foraminifera (<i>N. labradorica</i> and <i>G. auriculata arctica</i>)	-14	1030 ± 15	0	600 ± 35
B997-316 GGC	135	CURL-19693	mollusk (<i>T. equalis</i>)	-9	1040 ± 15	0	620 ± 25
B997-316 GGC	160	CURL-19511	mollusk (<i>T. equalis</i>)	-8.5	1075 ± 15	0	645 ± 15
B997-316 GGC	212.5	CURL-20191	mollusk (<i>T. equalis</i>)	-5	1245 ± 15	0	780 ± 35

1111

1112

1113 **Table 2:** Surface sediment calibration information. * indicates data from Rodrigo-Gámiz et al.
1114 (2015).

Site ID	Latitude	Longitude	Water depth (mbsl)	TEX ₈₆ ^L
B997-313	66.617000°	-23.933000°	213	-0.67
B997-315	66.736000°	-24.332000°	217	-0.69
B997-316	66.746000°	-18.792000°	658	-0.68
B997-319	66.447000°	-18.843000°	422	-0.72
B997-324	66.527000°	-21.152000°	281	-0.65
B997-334	66.410000°	-21.880000°	112	-0.70
B997-329	65.965000°	-21.294000°	112	-0.68
B997-331	66.136000°	-21.591000°	165	-0.71
B997-344	64.836000°	-24.369000°	284	-0.61
B997-346	64.927000°	-24.129000°	320	-0.63

B997-347	63.928000°	-24.482000°	327	-0.64
Station 1*	62.000317°	-15.999183°	2255	-0.49
Station 7*	61.498550°	-24.172250°	1628	-0.51
Station 3*	63.366200°	-16.628267°	240	-0.59
Station 5*	63.583267°	-22.143733°	188	-0.62
Station 6*	63.238233°	-22.561417°	315	-0.61
Station 8*	64.293183°	-24.147083°	260	-0.62
Station 10*	66.677450°	-24.179500°	241	-0.64
Station 11*	66.633317°	-20.833433°	367	-0.63
Station 13*	67.501633°	-15.069217°	884	-0.71
Station 14*	66.303100°	-13.972817°	262	-0.68

1115 **Fig. 1:** A) Overview maps of modern Icelandic oceanography. A) February 2014 and B) May 2014
1116 50 m depth *in situ* temperature integrated from local CTD stations. Marine sediment cores (black
1117 dots) and used B997 surface sediment sample locations (black + and B997-316 GGC core site) are
1118 marked. C) May 2014 S-N trending cross section of NIS bathymetry and vertical *in situ* temperature
1119 structure along the Siglunes transect (A-A' in panels A and B) and through the B997-316 GGC
1120 marine sediment core site. Data from Hafrannsóknastofnun ([Marine and Freshwater Research
1121 Institute, <http://www.hafro.is/Sjora>](http://www.hafro.is/Sjora)).

1122

1123 **Fig. 2:** CLAM age model. Gray shaded area denotes the 95% confidence envelope ([Blaauw, 2010](#)).
1124 Teal and asterisked mollusk ages are from the adjacent short gravity core, B997-316 SGC.
1125 Radiocarbon information provided in Table 1.

1126

1127 **Fig. 3:** B997-316 GGC marine sediment core climate proxies over the last millennium. A) % quartz,
1128 B) % calcite, C) triene Z concentrations, D) diene II concentrations, E) IP₂₅ concentrations, and F)
1129 TEX₈₆^L. Blue boxes highlight colder, LIA-like conditions reflected in the surface climate proxies
1130 (A-E) and the subsurface proxy (F).

1131

1132 **Fig. 4:** Regression analysis summary of surface sediment GDGT calibration. A) Correlation
1133 coefficient (R²) of all 21 surface sediment TEX₈₆^L values ([Rodrigo-Gámiz et al., 2015](#); [this study](#))

1134 against seasonal and annual temperature depth integrations. B) Calibration of Icelandic marine
1135 surface sediment TEX_{86}^L values against winter 0-200 m temperature, where gray lines denote the
1136 95% confidence envelope. Surface sediment data shown as closed circles ([this study](#)) and open
1137 circles ([Rodrigo-Gámiz et al., 2015](#)).

1138

1139 **Fig. 5:** Comparison of the available TEX_{86}^L temperature calibrations on the B997-316 GGC
1140 sediment record. Icelandic winter subsurface temperature ([this study](#)), annual SST ([Kim et al.,](#)
1141 [2010](#)) and annual subsurface temperature ([Kim et al., 2012](#)). Modern (1995-2004 CE) winter
1142 subsurface temperature at the B997-316 GGC site marked with gray dashed line ([Locarnini et al.,](#)
1143 [2010](#)).

1144

1145 **Fig. 6:** Comparison of select B997-316 GGC marine climate proxies to other well-dated Icelandic
1146 NIS marine climate records. A) B997-316 GGC IP_{25} concentrations ([this study](#)), B) triene Z
1147 concentrations ([this study](#)), C) MD99-2275 alkenone-inferred SST ([Sicre et al., 2011](#)), D) B997-
1148 316 GGC GDGT-inferred subsurface temperatures, with values below the record mean (4.34 °C)
1149 highlighted in blue ([this study](#)), and E) schlerochronological ΔR record, where increases in ΔR_{shell}
1150 values reflect the incursion of older, Arctic waters ([Wanamaker et al., 2012](#)). Vertical yellow bars
1151 highlight the period of interpreted thick sea ice, and the associated insulation/warming of the
1152 subsurface. Dashed blue lines bound the inferred periods of LIA-like conditions for the surface (A-
1153 C) and subsurface (D).

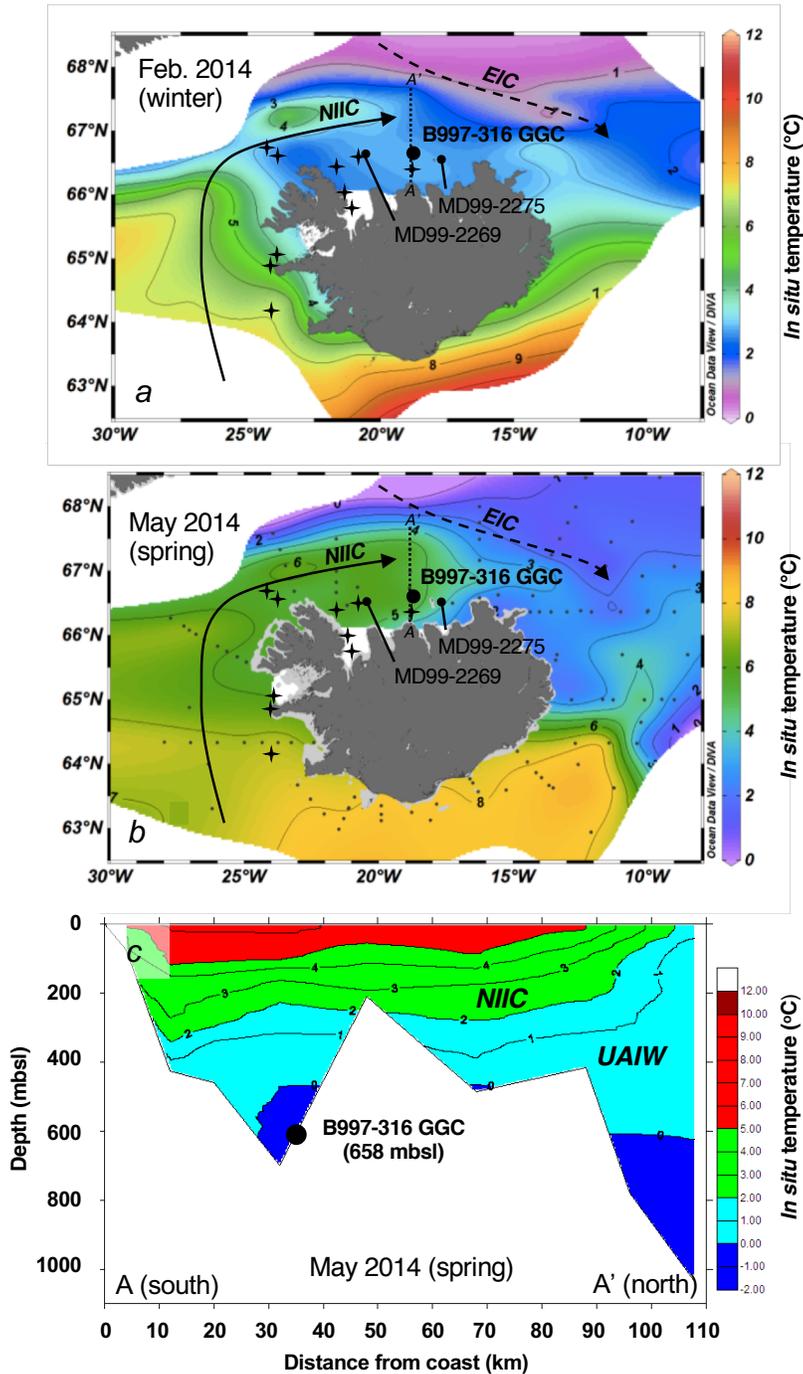


Fig. 1: A) Overview maps of modern Icelandic oceanography. A) February 2014 and B) May 2014 50 m depth potential temperature integrated from local CTD stations. Marine sediment cores (black dots) and used B997 surface sediment sample locations (black + and B997-316 GGC core site) are marked. C) May 2014 S-N trending cross section of NIS bathymetry and vertical potential temperature structure along the Siglunes transect and through the B997-316 GGC marine sediment core site. Data from Hafrannsóknastofnun ([Marine and Freshwater Research Institute, http://www.hafro.is/Sjora/](http://www.hafro.is/Sjora/)).

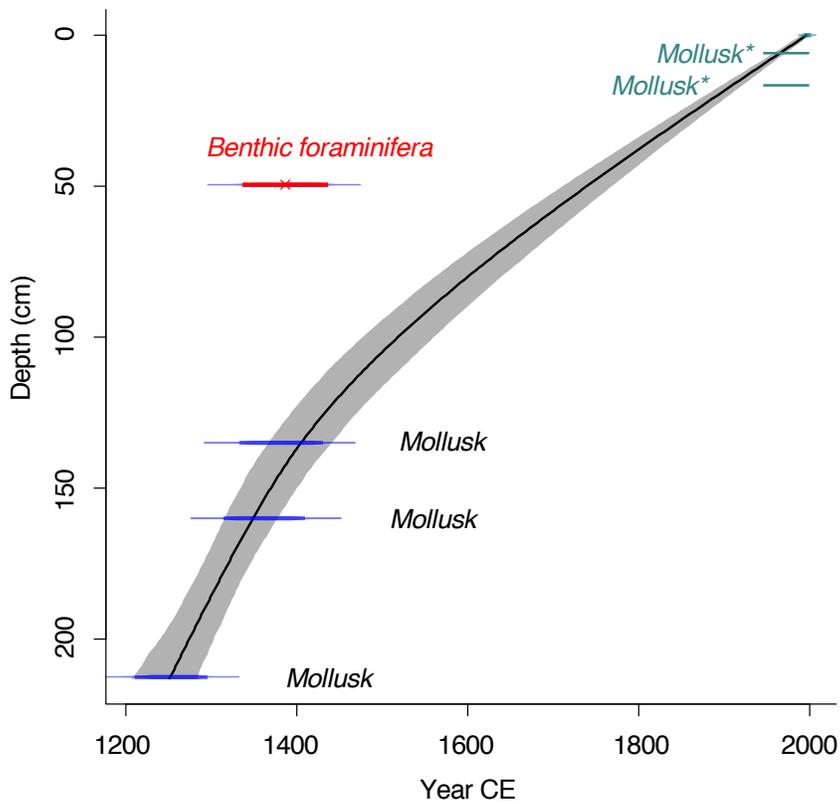


Fig 2: CLAM age model. Gray shaded area denotes the 95% confidence envelope (Blaauw, 2010). Teal and asterisked mollusk ages are from the adjacent short gravity core, B997-316 SGC, and not used as age control points in this model. Radiocarbon information provided in Table 1.

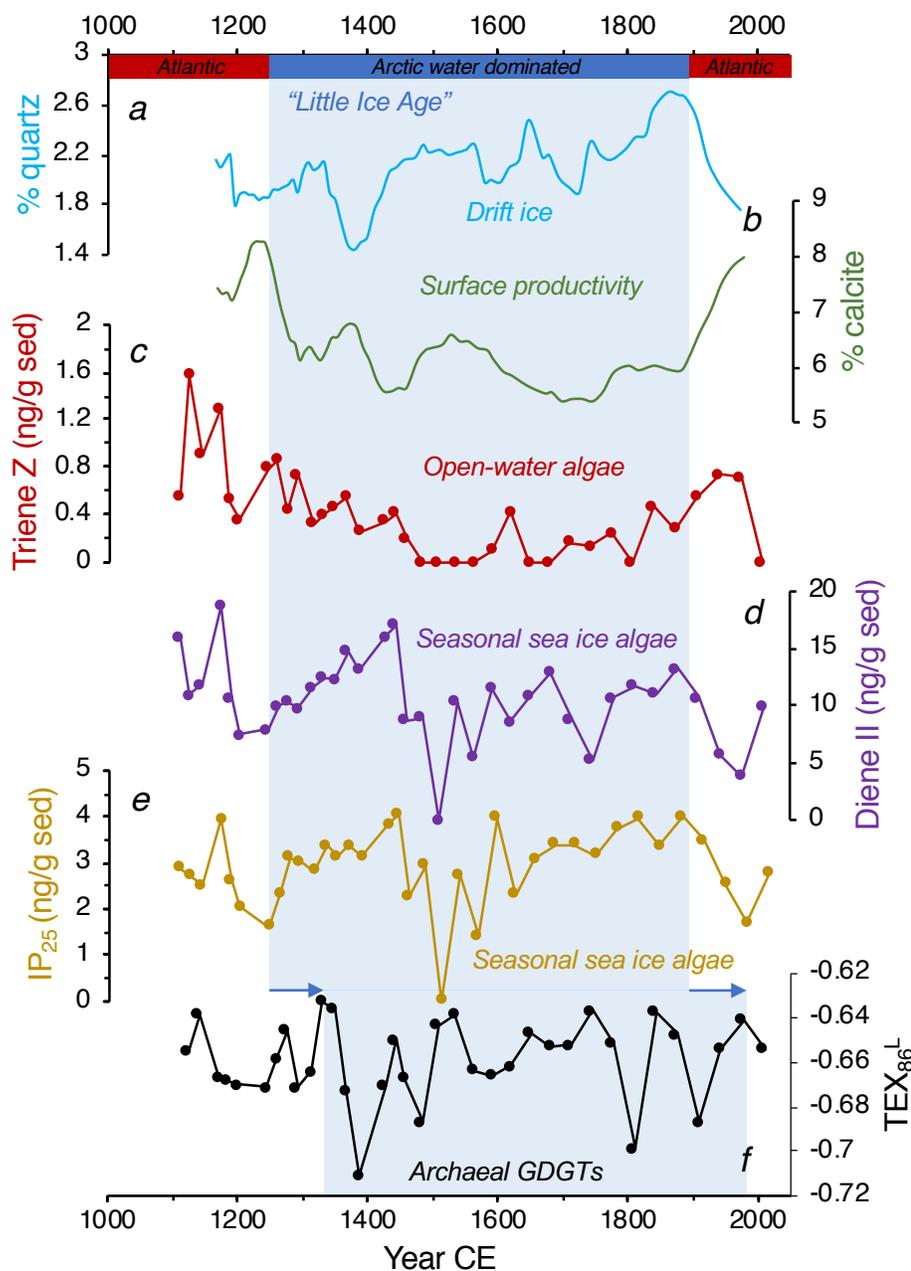


Fig. 3: B997-316 GGC marine sediment core climate proxies over the last millennium. A) % quartz, B) % calcite, C) triene Z concentrations, D) diene II concentrations, E) IP₂₅ concentrations, and F) TEX₈₆^L. Blue boxes highlight colder, LIA-like conditions reflected in the surface climate proxies (A-E) and the subsurface proxy (F).

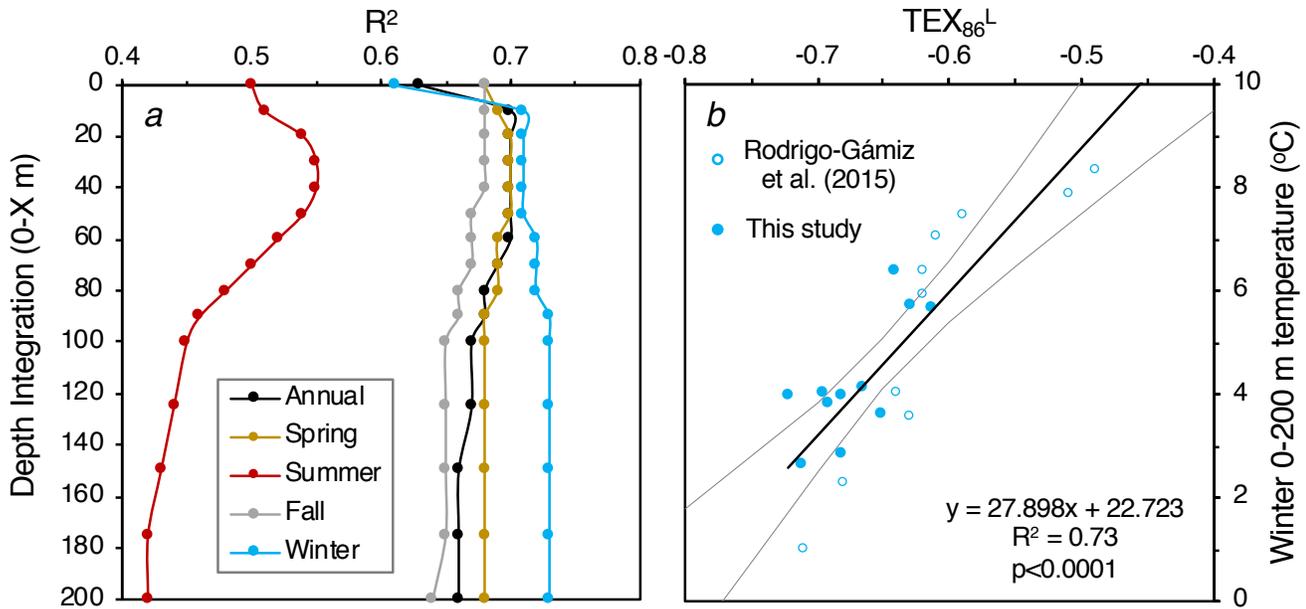


Fig. 4: Regression analysis summary of surface sediment GDGT calibration. A) Correlation coefficient (R^2) of all 21 surface sediment TEX_{86}^L values against seasonal and annual temperature depth integrations. B) Calibration of Icelandic marine surface sediment TEX_{86}^L values against winter 0-200 m temperature, where gray lines denote the 95% confidence envelope. Surface sediment data shown as closed circles ([this study](#)) and open circles ([Rodrigo-Gámiz et al., 2015](#)).

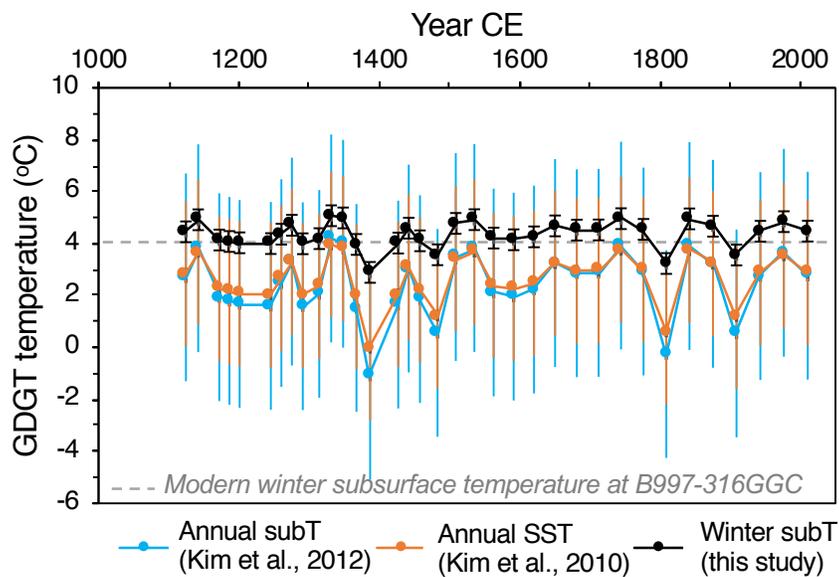


Fig. 5: Comparison of the available TEX_{86}^L temperature calibrations on the B997-316 GGC sediment record. Icelandic winter subsurface temperature ([this study](#)), annual SST ([Kim et al., 2010](#)) and annual subsurface temperature ([Kim et al., 2012](#)). Modern winter subsurface temperature at the B997-316 GGC site marked with gray dashed line.

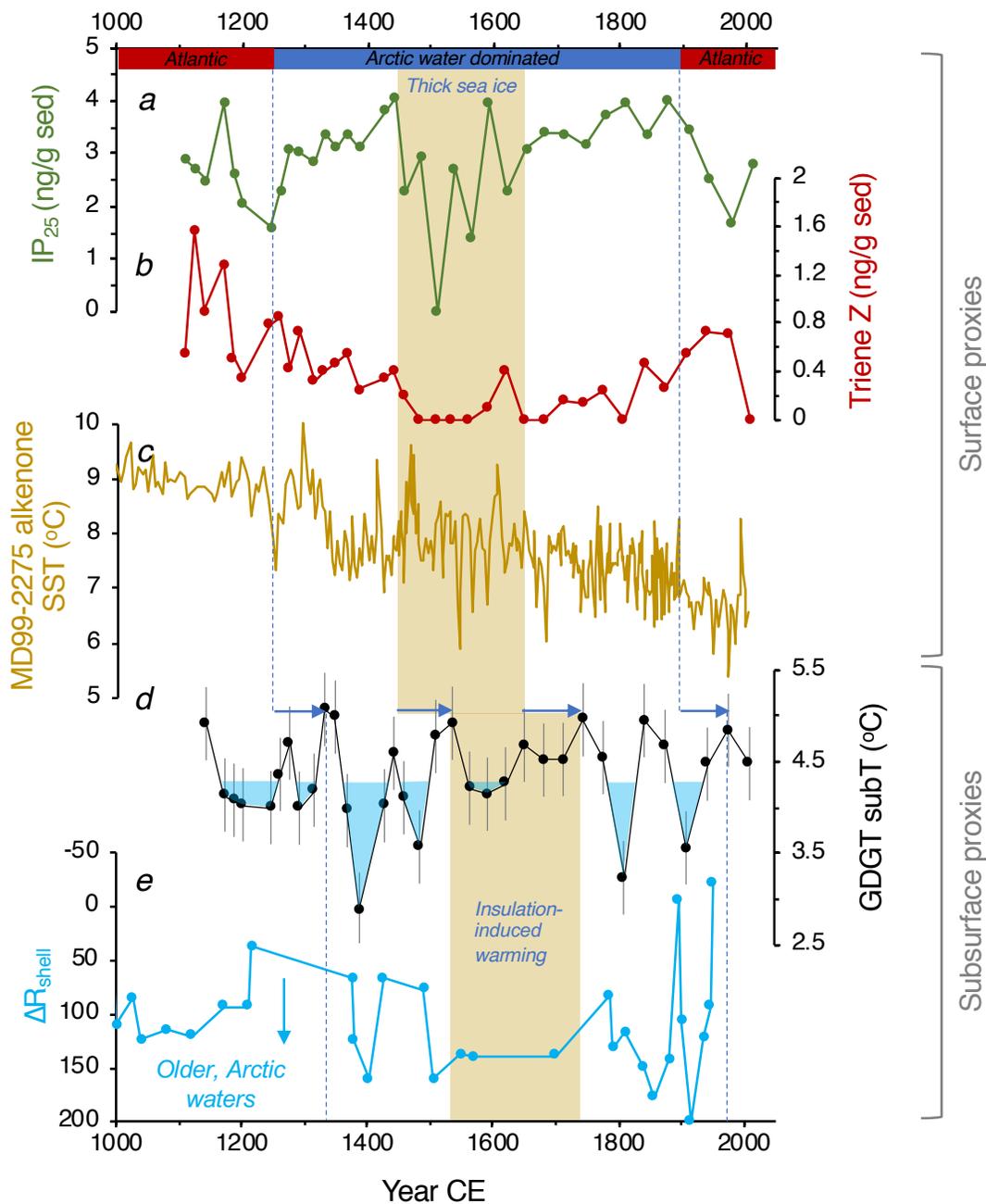
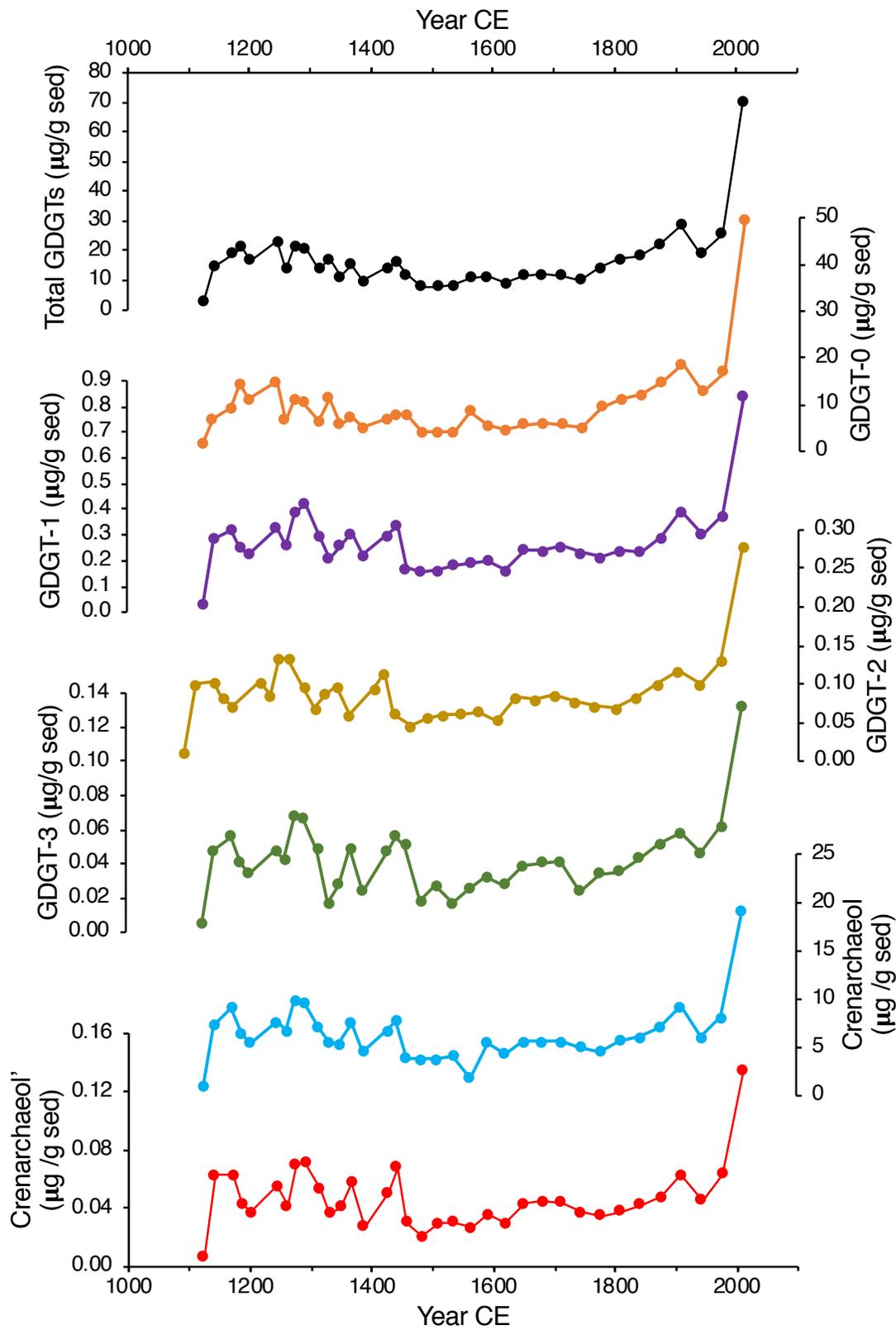
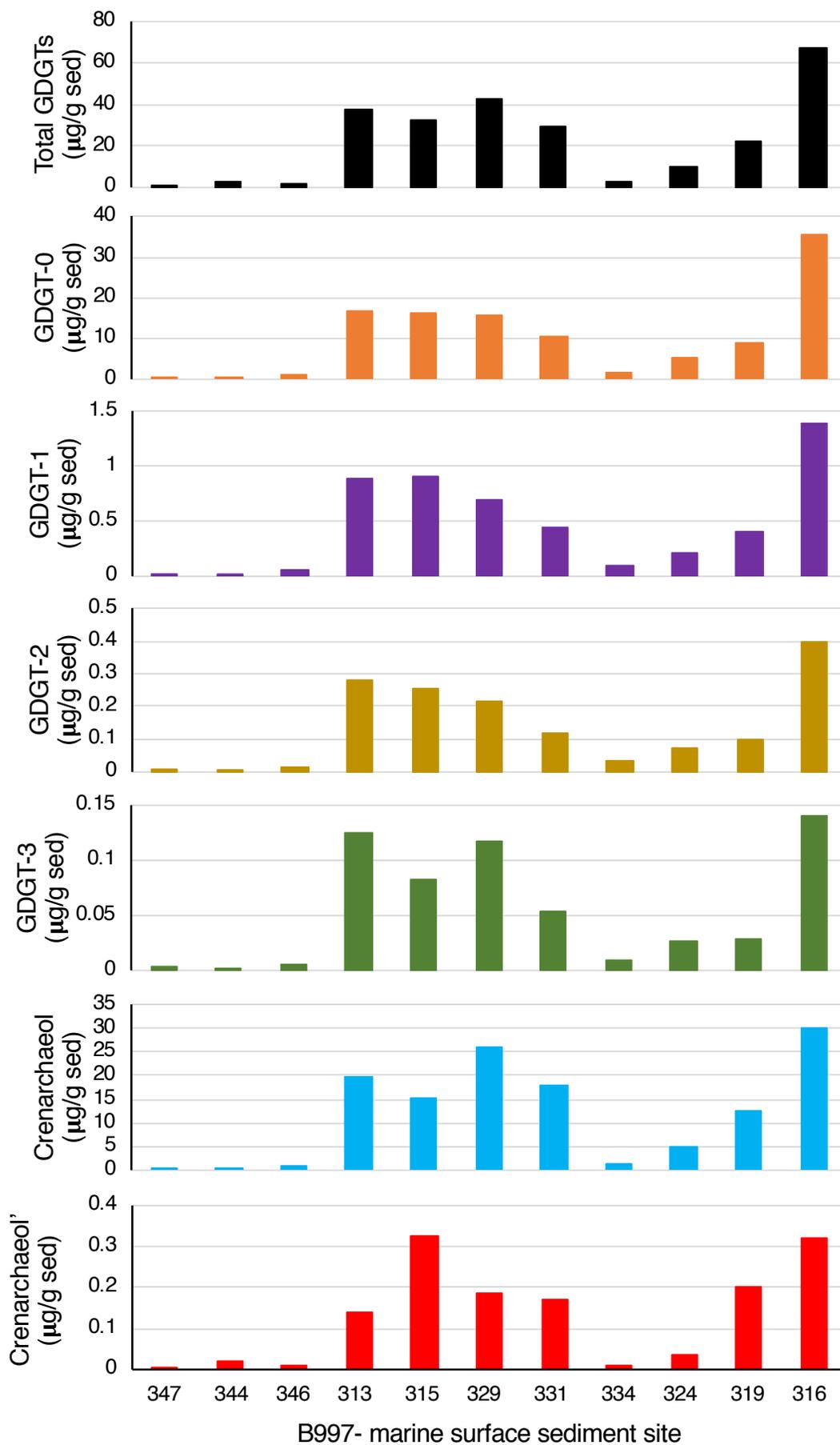


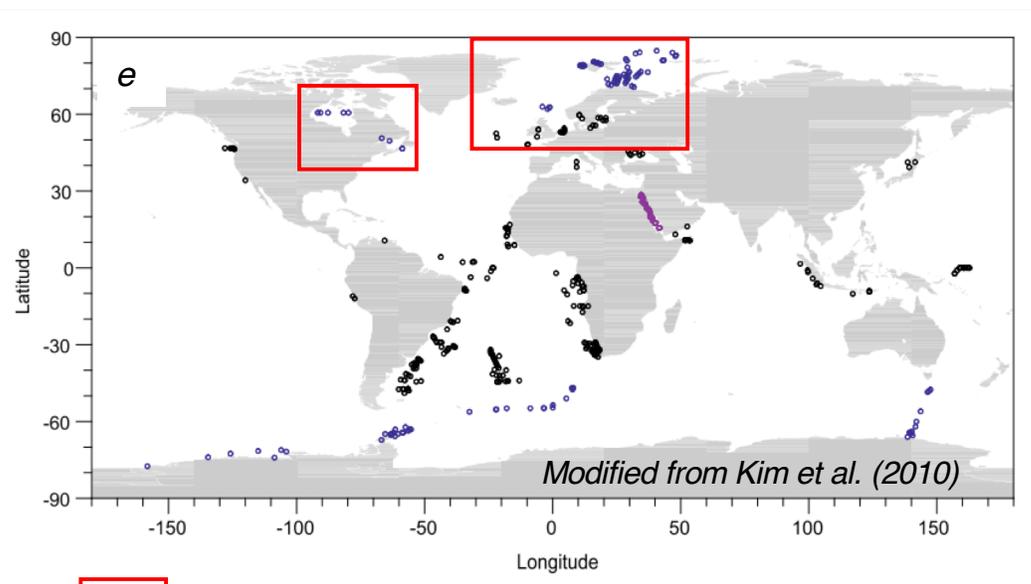
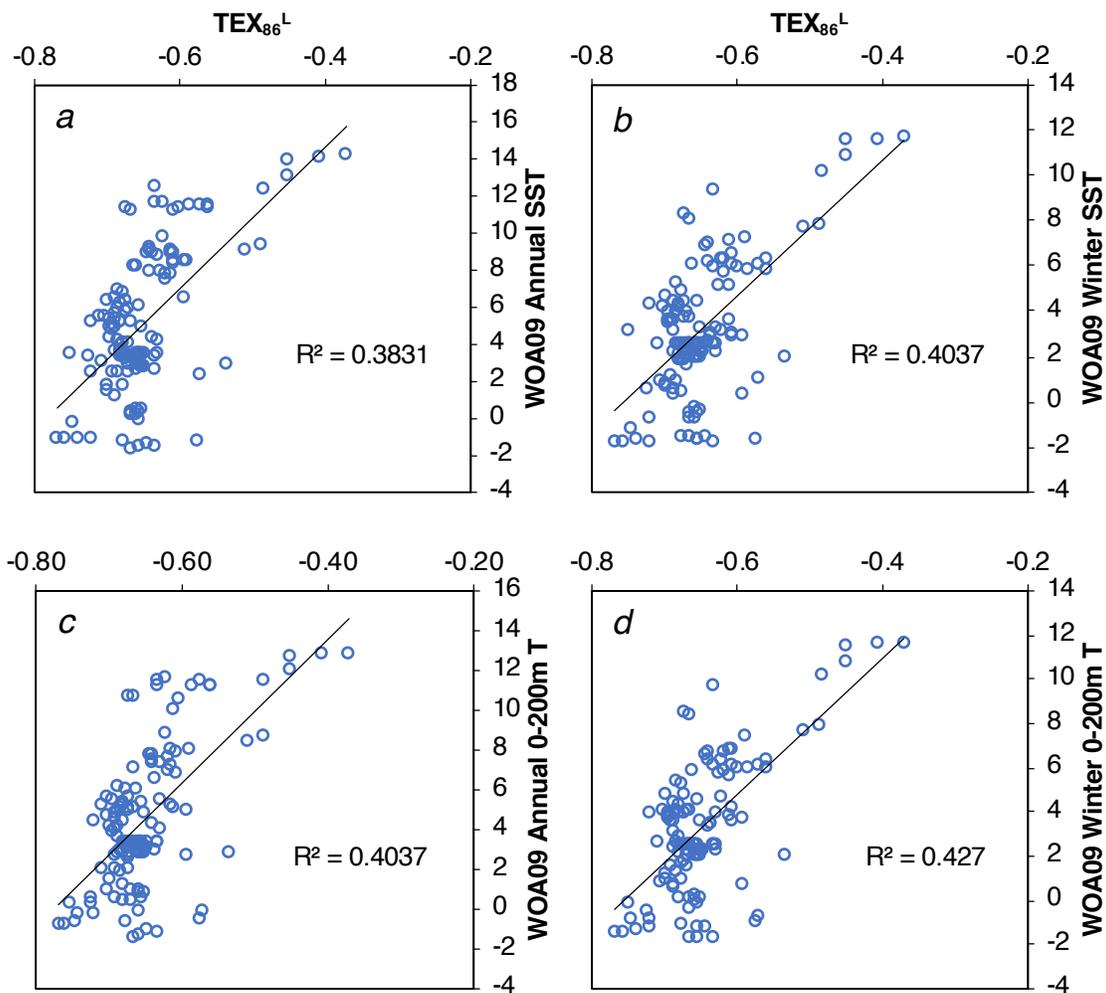
Fig. 6: Comparison of select B997-316 GGC marine climate proxies to other well-dated, high-resolution Icelandic NIS marine climate records. A) B997-316 GGC IP_{25} concentrations ([this study](#)), B) Triene Z concentrations ([this study](#)), C) MD99-2275 alkenone-inferred SST ([Sicre et al., 2011](#)), D) B997-316 GGC GDGT-inferred subsurface temperatures, with values below the record mean highlighted in blue ([this study](#)), and E) sclerochronological ΔR_{shell} record, where increases in ΔR_{shell} values reflect the incursion of older, Arctic waters, and a weaker AMOC ([Wanamaker et al., 2012](#)). Vertical yellow bars highlight the period of interpreted thick sea ice, and then the delayed associated insulation/warming of the subsurface. Dashed blue lines bound the inferred periods of LIA-like conditions for the surface (A-C) and subsurface (D) proxies.



Supplemental Fig S1: GDGT concentrations in B997-316 GGC marine sediment samples.



Supplemental Fig S2: GDGT concentrations in B997 marine surface sediment samples. Sample labels are abbreviated (i.e., 347 = B997-347) and ordered geographically from the southwestern-most (347, left) to the northeastern-most (316, right).



 Northern Hemisphere sites added to our calibration

Supplemental Fig S3: Using samples selected from the global calibration of Kim et al. (2010) we tested whether we could improve the local calibration by extending the range of samples, and thus, the environmental gradient. A) Annual SST, B) Winter SST, C) Annual 0-200 m T, and D) 0-200m T. Panel E highlights the northern hemisphere samples included from the the global calibration of Kim et al. (2010).