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**Measuring multiple cosmogenic nuclides in glacial cobbles
sheds light on Greenland Ice Sheet processes**

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1 **Abstract**

2 The behavior of the Greenland Ice Sheet during the Pleistocene remains uncertain due to the
3 paucity of evidence predating the Last Glacial Maximum. Here, we employ a novel approach,
4 cosmogenic nuclide analysis of individual subglacial-derived cobbles, which allows us to make
5 inferences about ice sheet processes and subglacial erosion. From three locations in western
6 Greenland, we collected 86 cobbles from the current ice sheet margin and nine cobbles
7 exposed on the modern proglacial land surface. We measured the concentration of in situ ^{10}Be
8 in all cobbles ($n = 95$) and ^{26}Al and ^{14}C in a subset ($n = 14$). Cobbles deposited during Holocene
9 retreat have ^{10}Be exposure ages generally consistent with the timing of ice retreat determined
10 by other methods. Conversely, most of the 86 subglacial cobbles contain very low
11 concentrations of ^{10}Be (median 1.0×10^3 atoms g^{-1}), although several have $\sim 10^4$ and one has
12 $\sim 10^5$ atoms g^{-1} . The low concentrations of ^{10}Be in most subglacial cobbles imply that their
13 source areas under the Greenland Ice Sheet are deeply eroded, preserving minimal evidence of
14 surface or near-surface exposure. The presence of measurable ^{14}C in ten of the cobbles requires
15 that they experienced cosmogenic nuclide production within the past ~ 30 ka; however, $^{14}\text{C}/^{10}\text{Be}$
16 ratios of ~ 6 suggest that nuclide production occurred during shielding by overlying material.
17 Only two of the 86 subglacial cobbles definitively have cosmogenic nuclide concentrations
18 consistent with prior surface exposure. Overall, isotopic analysis of subglacial cobbles indicates
19 that western Greenland's subglacial landscape is characterized by deep erosion and minimal
20 subaerial exposure.

21

22 **Keywords** ($n = 6$): Cosmogenic nuclides; Greenland; Geochemistry; Isotopes; Pliocene; Pleistocene

23 **1. Introduction**

24 The Greenland Ice Sheet is an erosive agent that has shaped the underlying landscape
25 for millions of years (Bierman et al., 2016). However, the depth of scouring, spatial distribution
26 of erosive versus non-erosive areas, and mechanisms of sediment entrainment and transport
27 remain uncertain due to the inaccessibility of the subglacial landscape. Ice retreat during
28 interglacial periods can expose a limited view of surfaces that are usually covered by ice, and
29 studies of sediments deposited in the marine realm (Bierman et al., 2016; Christ et al., 2019;
30 Flesche-Kleiven et al., 2002; Helland and Holmes, 1997; Larsen et al., 1994) provide an offshore
31 view of glacial processes. Analysis of bedrock at the bottom of ice cores (Schaefer et al., 2016)
32 provides a direct sampling of the subglacial landscape, albeit at a single point in space. But
33 overall, studies of the subglacial landscape remain limited and often rely on fragmentary and
34 indirect evidence.

35 Here, we seek to assess long-term ice sheet behavior and erosivity using a novel
36 approach: analysis of multiple cosmogenic nuclides in individual detrital cobbles sourced from
37 beneath the Greenland Ice Sheet and transported to the ice margin by ice and/or subglacial
38 water. We use 86 subglacial cobbles collected from the modern-day ice sheet margin and
39 proximal outwash streams as well as nine cobbles from the proglacial landscape (Fig. 1 and
40 Supplementary Data Fig. S1), the cosmogenic nuclide concentrations of which record ice sheet
41 processes over both time and space. The cobbles come from three regions in western
42 Greenland with distinct glaciological and erosive conditions.

43 Quantifying cosmogenic nuclide concentrations in detrital cobbles instead of at a single
44 location (e.g., bedrock from the GISP2 ice core, Schaefer et al. (2016)) provides wide spatial

45 coverage, allowing us to infer ice sheet processes across extensive subglacial sediment source
46 areas. Although the source location and flow path of any individual cobble is unknowable and
47 the exposure and erosion history cannot be modeled uniquely, investigating a large number of
48 cobbles yields patterns and trends in cosmogenic nuclide concentrations. We assess these
49 patterns in subglacial cobble cosmogenic nuclide concentrations to make inferences about
50 western Greenland Ice Sheet processes.

51

52 **2. Background**

53 *2.1. Using Multiple Cosmogenic Nuclides to Infer Cobble Exposure/Erosion History*

54 In situ produced cosmogenic nuclides, such as ^{10}Be , ^{26}Al , and ^{14}C , have been employed
55 for several decades to reconstruct glacial histories of bedrock surfaces and moraine boulders
56 (Balco, 2011). These nuclides, produced predominately by secondary neutron spallation
57 reactions (Lal and Peters, 1967) but also by muon interactions (Heisinger et al., 2002a;
58 Heisinger et al., 2002b) build up in rock at known rates over time. Because production by
59 neutron spallation decreases exponentially with depth, subglacial erosion of several meters of
60 rock strips most pre-existing spallation-produced nuclides (Balco, 2011). Production by muon
61 interactions (Heisinger et al., 2002a; Heisinger et al., 2002b) occurs at lower rates but to a
62 greater depth, such that muon-produced nuclides are present in rocks at low but measurable
63 concentrations even after glacial erosion has stripped tens of meters of surface material
64 (Bierman et al., 2016; Briner et al., 2016; Davis et al., 1999).

65 When multiple cosmogenic nuclides with different half-lives (e.g. ^{10}Be , 1.4 Ma; ^{26}Al , 0.7
66 Ma; and ^{14}C , 5.7 ka) are analyzed in the same sample, ratios of their concentrations provide

67 information about both exposure and burial (Briner et al., 2014). Nuclide production dominates
68 the former, whereas loss through radioactive decay dominates the latter. Pairing two longer-
69 lived nuclides, such as $^{26}\text{Al}/^{10}\text{Be}$, provides information about exposure and burial integrated
70 over the past 10^5 to 10^6 years (Balco et al., 2014), but is relatively insensitive to short durations
71 of burial (less than several hundred kyr). Shorter-lived nuclides, such as ^{14}C , are more sensitive
72 to recent exposure, burial, and erosion (Briner et al., 2014; Miller et al., 2006). Because the ^{14}C
73 half-life is two orders of magnitude less than those of ^{10}Be and ^{26}Al , appreciable ^{14}C mandates
74 spallation and/or muon production within ~ 30 ka.

75 Such multi-isotope approaches depend on knowing the ratio of the isotopes during
76 surface production. For $^{26}\text{Al}/^{10}\text{Be}$, the ratio of surface production is generally assumed to be
77 ~ 6.75 (Balco et al., 2008), although recent modeling (Argento et al., 2013) and empirical data
78 (Corbett et al., 2017) suggest that the value may be higher in certain locations. For $^{14}\text{C}/^{10}\text{Be}$, the
79 ratio of surface production, albeit less well-constrained, is ~ 3 -4 (Argento et al., 2013; Briner et
80 al., 2014; Schimmelpfennig et al., 2012). Ratios lower than production are indicative of burial
81 following initial exposure, or of prolonged surface exposure because the ratio decreases over
82 time due to preferential decay of the shorter-lived nuclide.

83 Because the production rates and ratios are dependent on depth, the ratios change as a
84 result of the relative proportion of the spallogenic and muogenic production (as reviewed in
85 Marrero et al. (2016)). Accounting for their respective instantaneous production pathways, the
86 $^{26}\text{Al}/^{10}\text{Be}$ production ratio increases only slightly with depth, whereas the $^{14}\text{C}/^{10}\text{Be}$ and $^{14}\text{C}/^{26}\text{Al}$
87 production ratios increase more appreciably with depth (Fig. 2). This occurs because the
88 muogenic fraction of the total is greater for ^{14}C than for ^{10}Be and ^{26}Al (Lupker et al., 2015).

89 Therefore, although $^{14}\text{C}/^{10}\text{Be}$ and $^{14}\text{C}/^{26}\text{Al}$ are generally used to assess burial following initial
90 exposure (Briner et al., 2014; Miller et al., 2006), nuclide ratios can also be used to assess
91 shielding depth during exposure (Rand and Goehring, 2019).

92

93 *2.2. Greenland Ice Sheet Processes and Implications for Subglacial Cobble Nuclide*

94 *Concentrations*

95 *2.2.1. Ice Sheet History and Possible Subglacial Cobble Exposure Periods*

96 Although continental glaciation occurred as early as the middle to late Miocene in East
97 Greenland (Helland and Holmes, 1997; Larsen et al., 1994), expansive ice likely first occupied
98 Greenland ~2.5 Ma, coincident with overall northern hemisphere cooling (Bierman et al., 2016;
99 Flesche-Kleiven et al., 2002). Since the inception of a large Greenland Ice Sheet at the beginning
100 of the Pleistocene, Greenland's landscape has been dominated by burial and progressive
101 erosion rather than exposure (Bierman et al., 2016).

102 In addition to being exposed to cosmic rays before the onset of glaciation (with varying
103 degrees of partial shielding based on their depth of burial), the cobbles or their source outcrops
104 could have been re-exposed when ice extent was reduced during warm periods of the
105 Pleistocene (Schaefer et al., 2016). Interglacial sediments indicative of a warm Arctic exist in
106 several locations around Greenland and are thought to be early Pleistocene in age (Funder et
107 al., 2001). During MIS11 at ~400 ka, pollen evidence (De Vernal and Hillaire-Marcel, 2008) and
108 sediment provenance studies (Reyes et al., 2014) show that much of the southern Greenland
109 Ice Sheet disappeared. The Eemian Period ~130 ka was characterized by appreciably reduced
110 ice extent as suggested by marine sediment provenance (Colville et al., 2011) and modeling

111 (Helsen et al., 2013). Finally, the Greenland Ice Sheet was smaller than present for several
112 thousand years during the warm middle Holocene as evidenced by lake sediment core records
113 (Briner et al., 2010; Larsen et al., 2015).

114 Although nuclide production in the cobbles we analyzed could have occurred via
115 neutron spallation during exposure in conjunction with these warm periods, muon production
116 also decidedly occurred at depths of meters to tens of meters in the absence of surface
117 exposure. Because muon production extends many meters through overlying material
118 (Heisinger et al., 2002a; Heisinger et al., 2002b), nuclides would have been produced deep
119 within cobble source outcrops before the onset of glaciation and in cobbles or source outcrops
120 beneath overlying till, ice, or snow.

121

122 *2.2.2. Glaciological Processes, Subglacial Erosion, and Cobble Plucking*

123 In addition to recording ice sheet history, the cosmogenic nuclide concentrations of the
124 cobbles we sampled are the result of ice sheet processes including subglacial erosion, plucking,
125 freeze-on, and transport. The Greenland Ice Sheet likely began eroding parts of the underlying
126 landscape as soon as glaciation began, and erosion continued throughout the Pliocene and
127 Pleistocene, progressively eroding through the preglacial regolith and into bedrock at least in
128 certain areas (Bierman et al., 2016). Cobbles were likely sourced from areas that at some point
129 had basal temperatures at or near the pressure-melting point, in order for regelation to
130 incorporate the cobbles into the ice matrix (Alley et al., 1997). However, basal conditions of the
131 Greenland Ice Sheet are not well documented over space and time, and closely juxtapose
132 warm-based (erosive) and cold-based (non-erosive) ice (Petrunin et al., 2013). This spatial and

133 temporal heterogeneity means that cobble source areas likely changed over time along with
134 bed conditions.

135 The residence time of cobbles and sediment in the basal ice is largely determined by the
136 relationships between the basal thermal state, the ice thickness, and the vertical and horizontal
137 flow velocity. Cobbles are either incorporated into or shed from the ice depending on the
138 thermal state at the ice-bed interface (Cuffey and Paterson, 2010). The along-flow advection
139 rates of material within the ice varies from near zero at the ice sheet center to hundreds of
140 meters per year near the margin (Cuffey and Paterson, 2010). These rates of advection and the
141 incorporation or deposition of cobbles evolve through time owing to changes in ice sheet
142 conditions, which are constantly re-adjusting in response to changing climate.

143

144 **3. Study Sites**

145 Our study focuses on three locations in western Greenland: Kangerlussuaq, Ilulissat, and
146 Upernavik (Fig. 1). We chose these three sites because their contrasting landscapes are
147 indicative of different glacial processes and erosive regimes. The chronology of ice retreat and
148 the likely timing of surficial cobble exposure is constrained by previous work at each of the
149 three sites.

150 In Kangerlussuaq (67°N, -50°E), the landscape morphology and existing cosmogenic
151 nuclide data are suggestive of deep glacial erosion. The region is characterized by a glacially-
152 sculpted landscape of NE-SW elongated hills and lakes, carved parallel to the direction of ice
153 flow. Rounded, striated bedrock is common. Till fills the valleys, whereas hilltops are typically
154 bare rock. Numerous Holocene moraines are preserved; ¹⁰Be analyses of moraine boulders

155 (Levy et al., 2012) cluster and are consistent with the local organic radiocarbon chronology,
156 suggesting no or minimal inheritance of nuclides from previous exposure periods. Similarly,
157 $^{26}\text{Al}/^{10}\text{Be}$ analyses of high-elevation bedrock surfaces in the area generally indicate deep
158 erosion followed by a single period of exposure during the Holocene (Beel et al., 2016).
159 Meteoric ^{10}Be concentrations in the fine-grained glacial sediment at Kangerlussuaq (n=17) are
160 significantly lower than in other regions of Greenland, suggestive of effective subglacial erosion
161 (Graly et al., 2018). The deposition of the three surficial cobbles we measured likely occurred
162 ~ 6.8 ka based on the moraine chronology of Levy et al. (2012).

163 In Ilulissat (69°N, -50°E), the land surface and previously published cosmogenic nuclide
164 data both suggest extensive glacial erosion, similar to in Kangerlussuaq. The landscape is
165 glacially sculpted, heavily striated, and cut by numerous fjords. Most of the ice in the region
166 drains through a large ice stream, Jakobshavn Isbræ. Holocene moraines are present and
167 dozens of cosmogenic nuclide analyses (reviewed in Young et al. (2013)) suggest that ^{10}Be
168 inheritance from previous exposure is minimal. The deposition of the three surficial cobbles we
169 measured likely occurred ~ 7.8 ka based on the ages of a bedrock sample (GL080; 7.9 ± 0.2 ka)
170 and a boulder sample (GL081; 7.6 ± 0.1 ka) collected from the same location as the cobbles
171 (Corbett et al., 2011).

172 In Upernavik ($\sim 72^\circ\text{N}$, -54°E), unlike Kangerlussuaq and Ilulissat, the landscape
173 morphology and existing cosmogenic nuclide data indicate that glacial erosion was
174 heterogeneous and limited. The region has large relief, characterized by table-top highlands cut
175 by deep fjords. Although some low-elevation bedrock surfaces exhibit glacial rounding, the
176 highlands show evidence of prolonged subaerial weathering including exfoliation, tors, and

177 weathering pits. Analysis of ^{10}Be and ^{26}Al in Upernavik (Corbett et al., 2013) and south of
178 Upernavik (Beel et al., 2016) indicates that the ice was cold-based and non-erosive at times in
179 the past. At high elevations, multi-isotope analysis shows that surfaces preserve total histories
180 of $\sim 10^5$ - 10^6 years, and even some low-elevation surfaces contain nuclides inherited from
181 periods of exposure prior to the Holocene (Beel et al., 2016; Corbett et al., 2013). Meteoric ^{10}Be
182 concentrations of fine-grained sediment entrained in the ice margin are higher here than in
183 Kangerlussuaq or Ilulissat (Graly et al., 2018). The deposition of the three surficial cobbles we
184 measured (collected from an upland surface proximal to the ice margin) may have occurred
185 ~ 12.1 ka based on the mean of ages from a bedrock sample (GU001; 13.6 ± 0.3 ka) and a
186 boulder sample (GU002; 10.6 ± 0.3 ka) collected from the same location, or ~ 11 ka based on
187 our best deglaciation estimate for the region (Corbett et al., 2013), although all of these
188 estimates may be inflated by the presence of inherited ^{10}Be .

189

190 **4. Methods**

191 *4.1. Study Design and Sample Collection*

192 We measured in situ cosmogenic nuclides (^{10}Be in all; ^{26}Al and ^{14}C in a subset) in 86
193 subglacial cobble-sized rocks from Kangerlussuaq (n=33), Ilulissat (n=20), and Upernavik (n=33)
194 (Fig. 1, Supplementary Data Table S1). Most of these cobbles (“icebound cobbles”, n = 62) were
195 sourced from sediment-laden basal ice exposed at the ice sheet margin or supraglacial debris
196 bands not near nunataks (Figs. S1A, S1B, S1C), while a smaller portion (“outwash cobbles”, n =
197 24) were sourced from ice-proximal channels just outside of large outwash tunnels (Fig. S1D).
198 We recorded information about cobble size, lithology, and angularity, and measured the

199 location of collection with a hand-held GPS (Table S1). Cobble lithologies vary, although most
200 are quartz-rich crystalline rocks including granite and gneiss; a much smaller portion are
201 quartzite, breccia, and schist.

202 We also analyzed ^{10}Be in an additional three surficial cobbles from each site ($n = 9$ total,
203 detailed in Table S1, Figs. S1E and S1F) from the modern proglacial landscape to assess whether
204 inherited ^{10}Be is detectable in samples that have been exposed since deglaciation. At two sites,
205 Ilulissat and Upernavik, we collected the surficial cobbles directly adjacent to a bedrock-boulder
206 sample pair. We purposefully selected cobbles of similar sizes to those collected at the ice
207 margin so that the two populations are comparable. At all three sites, the surficial cobbles came
208 from local topographic highpoints with little/no present-day sediment or vegetation cover. At
209 each site, the cobbles were collected within close proximity of each other, usually a few meters.

210

211 *4.2. Sample Preparation and Analysis*

212 Additional methodological detail can be found in the Supplementary Data and in Tables
213 S1-S5. For ^{10}Be and ^{26}Al , Samples were prepared at University of Vermont using procedures
214 described in Corbett et al. (2016). $^{10}\text{Be}/^9\text{Be}$ ratios were measured by accelerator mass
215 spectrometry (AMS) at Lawrence Livermore National Laboratory and corrected for backgrounds
216 using a $^{10}\text{Be}/^9\text{Be}$ blank ratio of $(4.2 \pm 1.6) \cdot 10^{-16}$ ($n = 24$, Table S3). We chose a threshold ^{10}Be
217 concentration of $3 \cdot 10^3$ atoms g^{-1} (exceeded by 14 of the 86 subglacial cobbles), above which we
218 also analyzed ^{26}Al and ^{14}C . This analytical threshold was chosen to select for samples with
219 sufficient ^{26}Al to be measurable above detection limits and to yield meaningful $^{26}\text{Al}/^{10}\text{Be}$ ratios.
220 $^{26}\text{Al}/^{27}\text{Al}$ ratios were measured by AMS at Purdue Rare Isotope Measurement (PRIME)

221 Laboratory and corrected for backgrounds using a $^{26}\text{Al}/^{27}\text{Al}$ blank ratio of $(7.6 \pm 7.0) \cdot 10^{-16}$ ($n =$
222 14, Table S3). For ^{14}C , sample preparation and measurement by AMS were conducted at
223 University of Cologne (Fulop et al. (2015), Table S4).

224 For the nine surficial cobbles, we calculated exposure ages (Table S5) with the CRONUS
225 Earth online exposure age calculator (Balco et al., 2008). We used the northeastern North
226 American production rate calibration dataset and Lal/Stone scaling (see Supplementary Data
227 for details).

228

229 *4.3. Theoretical Models*

230 To explore the concentrations of ^{10}Be and ^{26}Al under a long-lived, erosive ice sheet, we
231 modeled the evolution of their concentrations in a bedrock profile for various erosion,
232 exposure, and ice thickness scenarios. We assumed sea-level high-latitude production rates,
233 including production by muons, calculated using the MATLAB implementation in Balco et al.
234 (2008) of the method of Heisinger et al. (2002b) and a $^{26}\text{Al}/^{10}\text{Be}$ surface production ratio of 7.3
235 (Corbett et al., 2017). The simulations were initialized with a bedrock profile in steady state
236 (i.e., nuclide production equal to nuclide loss from decay and erosion), based on various
237 constant preglacial erosion rates (5, 20, and 50 m Myr⁻¹) assuming no ice cover prior to the
238 Pleistocene. We then considered two exposure scenarios following the onset of glaciation at 2.5
239 Ma: (1) continuous ice cover and (2) 8 kyr of interglacial exposure every 100 kyr (i.e., scenario 2
240 in Schaefer et al. (2016), their Fig. 3). The sensitivity of results to muon production through ice
241 was examined for a range of ice thicknesses (50 m, 200 m, 1000 m, and infinite) during intervals
242 of cover. We modeled erosion by removing surface material and shifting the profile up in

243 proportion to the subglacial erosion rate (5, 20, and 50 m Myr⁻¹) when ice covered, and
244 assumed zero erosion when ice free. Nuclides experience continuous decay throughout each
245 simulation.

246

247 5. Results

248 For the 73 subglacial cobbles with ¹⁰Be/⁹Be detectable above background values, ¹⁰Be
249 concentrations are $(2.0 \pm 1.0) \cdot 10^2$ to $(1.12 \pm 0.02) \cdot 10^5$ atoms g⁻¹ (Table S1), ranging over three
250 orders of magnitude. The ¹⁰Be concentrations form a right-skewed distribution with a median
251 of $1.0 \cdot 10^3$ atoms g⁻¹ and a mean of $4.2 \cdot 10^3$ atoms g⁻¹ (Fig. 3). The ¹⁰Be concentrations do not
252 form statistically separable populations (Fig. 4) based on type (icebound vs. outwash; t-test, p =
253 0.29), location (Kangerlussuaq, Ilulissat, Upernavik; ANOVA, p = 0.67), shape (angular,
254 subangular, subrounded, rounded; ANOVA, p = 0.46), or lithology (gneiss, granite, other;
255 ANOVA, p = 0.81).

256 The 14 subglacial cobbles with the highest ¹⁰Be concentrations ($>3 \cdot 10^3$ atoms g⁻¹, the
257 analytical threshold used to determine which samples we analyzed for ²⁶Al and ¹⁴C) also do not
258 cluster with regards to cobble characteristics (Table S1). They have varying lithologies and
259 varying shapes, ranging from subrounded to angular, although none of the 14 were well-
260 rounded. They came from both the ice margin itself as well as outwash deposits, with no
261 systematic bias toward one or the other.

262 ²⁶Al concentrations are $(2.3 \pm 0.3) \cdot 10^4$ to $(7.7 \pm 0.3) \cdot 10^5$ atoms g⁻¹ and the resulting
263 ²⁶Al/¹⁰Be ratios are 5.0 ± 1.1 to 8.4 ± 1.2 (n = 13, Tables 1 and S2, Fig. 5). In reference to the
264 empirically-determined Greenland ²⁶Al/¹⁰Be surface production ratio of 7.3 (Corbett et al.,

265 2017), none of the cobbles are above the production ratio by $>1\sigma$, 7 are indistinguishable, and 6
266 are below by $>1\sigma$. Conversely, in reference to the commonly-assumed $^{26}\text{Al}/^{10}\text{Be}$ surface
267 production ratio of 6.75, 5 cobbles are above by $>1\sigma$, 5 are indistinguishable, and 3 are below
268 by $>1\sigma$.

269 Of the cobbles analyzed for ^{14}C , three were below detection limit; the remaining ten ^{14}C
270 concentrations are $(5.1 \pm 1.4) \cdot 10^4$ to $(1.5 \pm 0.2) \cdot 10^5$ atoms g^{-1} ($n = 13$, Tables 1 and S4). All but
271 two samples (GK015 and GU010) have significant ^{14}C in excess of the instantaneous $^{14}\text{C}/^{10}\text{Be}$
272 surface production ratio (Fig. 6). When $^{14}\text{C}/^{10}\text{Be}$ and $^{14}\text{C}/^{26}\text{Al}$ are plotted as regressions, all but
273 the same two of the samples lie along a trendline forming a significant linear relationship ($R^2 =$
274 0.92 for both regressions); the slope of the regression is 5.92 for $^{14}\text{C}/^{10}\text{Be}$ and 0.76 for $^{14}\text{C}/^{26}\text{Al}$
275 (Fig. 7), both well above the surface production ratios. The two samples not on the regression
276 (GK015 and GU010) have higher ^{10}Be and ^{26}Al concentrations but lower ^{14}C concentrations than
277 the remainder of the dataset (Fig. 6).

278 For the nine surficial cobbles, ^{10}Be concentrations are $(3.9 \pm 0.1) \cdot 10^4$ to $(6.9 \pm 0.1) \cdot 10^4$
279 atoms g^{-1} (Table S5). The average ^{10}Be concentrations by site are $(4.4 \pm 0.6) \cdot 10^4$ for
280 Kangerlussuaq, $(5.4 \pm 0.5) \cdot 10^4$ for Ilulissat, and $(5.6 \pm 1.2) \cdot 10^4$ for Upernavik (1SD, $n = 3$ for
281 each), representing relative standard deviations of 12.5, 8.5, and 20.9%, respectively. When
282 considered as exposure ages assuming constant exposure and no erosion, these translate to 6.8
283 ± 0.8 ka (Kangerlussuaq), 8.1 ± 0.7 ka (Ilulissat), and 7.8 ± 1.6 ka (Upernavik; $n = 3$, average, 1SD
284 for each, Table S5).

285

286

287 **6. Discussion**

288 *6.1. Subglacial Cobbles Generally Record Deep Subglacial Erosion*

289 Most of the subglacial cobbles we measured have cosmogenic nuclide concentrations
290 indicating deep erosion under ice without subsequent surface exposure. Long-exposed, high-
291 latitude landscapes, such as the Tertiary preglacial Greenland land surface, would have had ^{10}Be
292 concentrations of $\sim 10^5$ - 10^6 atoms g^{-1} , depending on subaerial erosion rates (see Bierman et al.
293 (2016), their Figs. 2a and 4). The cobbles we measured had much lower ^{10}Be concentrations (13
294 below detection limit and an additional 37 with $<10^3$ atoms g^{-1} , Fig. 3), suggesting they were
295 sourced from deeply-eroded outcrops with little exposure to cosmic radiation. Some of the
296 cobbles with low ^{10}Be concentrations may have never experienced exposure at the surface;
297 their nuclide inventories could be due to muogenic production at depth (Heisinger et al., 2002a;
298 Heisinger et al., 2002b) during shielding by overlying rock, sediment, and/or ice. The general
299 lack of ^{10}Be in the samples demonstrates that an uneroded or minimally eroded, subglacially-
300 preserved, Tertiary landscape was not the source material for the cobbles we collected.

301 However, all glacial detrital sediment records, including the cobbles we studied, are
302 biased toward areas of the ice sheet that generate significant volumes of sediment. The cobbles
303 were most likely derived from areas of warm-based ice, or areas that had warm-based ice
304 during at least one time, in order for plucking or freeze-on to have occurred. This over-
305 representation of erosive areas is similar to biases in records developed from marine sediment
306 cores (Bierman et al., 2016; Christ et al., 2019; Flesche-Kleiven et al., 2002; Helland and Holmes,
307 1997; Larsen et al., 1994) and studies of sediment emanating from glacial drainages (Nelson et
308 al., 2014). The cobbles therefore resulted from processes operating in sediment source areas

309 beneath the ice sheet, presumably areas of warm-based and erosive ice, rather than the
310 subglacial landscape as a whole.

311 Our estimates of subglacial erosion suggest that at least tens to more likely hundreds of
312 meters of rock have been removed from the cobble source landscapes since the onset of
313 glaciation. Theoretical models (Figs. 8, S2, S3) demonstrate that subglacial erosion rates of ~20-
314 50 m Myr⁻¹ are needed to drive ¹⁰Be concentrations in exhumed material down to 10³ atoms g⁻¹
315 assuming thick, continuous ice cover since 2.5 Ma (the resulting ¹⁰Be concentrations are also
316 sensitive to the assumed preglacial erosion rate, which we varied from 5-50 m Myr⁻¹, Figs. 8,
317 S2). These erosion rates are minimum estimates, however, because our measured ²⁶Al/¹⁰Be
318 ratios (~5-8, close to that of surface production) and the ²⁶Al/¹⁰Be of bedrock from the bottom
319 of the GISP2 ice core (~4; Schaefer et al. (2016)) indicate that interior Greenland experienced
320 exposure during the Pleistocene, which would increase nuclide concentrations (Figs. 8, S3). The
321 ²⁶Al/¹⁰Be ratios we measured are also consistent with muon production through thin,
322 continuous ice cover for the past 2.5 Myr, but greater erosion rates (> 50 m Myr⁻¹) in this case
323 would be required to reduce ¹⁰Be concentrations to 10³ atoms g⁻¹ (Figs. 8, S2).

324 Overall, these findings are consistent with deep subglacial erosion and nuclides
325 produced by muons (whether through thin ice or overlying bedrock) during the Pleistocene.
326 These findings agree well with the quantitative estimates of subglacial erosion of Strunk et al.
327 (2017), who used paired ²⁶Al/¹⁰Be data from Greenland's coastal landscapes and a Monte Carlo
328 approach tuned to the oxygen isotope record to conclude that low-lying landscapes (i.e.
329 probable source areas for the cobbles we collected) have eroded at >50 m Ma⁻¹ during the
330 duration of ice cover. Similarly, Goehring et al. (2010) inferred 2-34 m of erosion during the last

331 glacial cycle based on ^{10}Be in East Greenland ice-contact delta sediments, implying erosion rates
332 of tens to hundreds of m Myr^{-1} .

333

334 *6.2. The Greenland Ice Sheet: An (Imperfect) Erosion Machine*

335 Within the spectrum of previously-published cosmogenic measurements of glacial
336 materials in Greenland (see previous studies plotted on Fig. 1), our subglacial cobbles (median
337 $1.0 \cdot 10^3$ atoms g^{-1} , mean $4.2 \cdot 10^3$ atoms g^{-1}) have ^{10}Be concentrations similar to sand emerging
338 from the present-day subglacial drainage system via outwash streams in southern Greenland
339 (mean = $6.5 \pm 4.1 \cdot 10^3$ atoms g^{-1} , $n = 19$, 1SD, Nelson et al. (2014)) and inheritance calculated
340 from eastern Greenland glacial delta depth profiles (error-weighted mean = $6.9 \pm 1.0 \times 10^3$
341 atoms g^{-1} , $n = 5$, 1SD, Goehring et al. (2010), their Fig. 8). The detrital sediment assessed by
342 both Nelson et al. (2014) and Goehring et al. (2010) is more similar to the mean than the
343 median of the cobble data we present, likely because subglacial erosion and transport
344 homogenizes sediment, combining material from more- and less-eroded areas, an effect that
345 we mimic with a large number of cobbles.

346 The ^{10}Be concentrations of Greenland's subglacial sediments (this study, Nelson et al.
347 (2014), and Goehring et al. (2010)) are lower today than they were in the past. Analyses of East
348 Greenland marine cores (Bierman et al., 2016) suggest that sediments shed off Greenland had
349 appreciably more ^{10}Be in the late Miocene and Pliocene ($\sim 10^5$ atoms g^{-1} with decay correction)
350 and somewhat more ^{10}Be throughout the late Pliocene and early Pleistocene ($\sim 10^4$ atoms g^{-1}
351 with decay correction), not reaching $\sim 10^3$ atoms g^{-1} until about the past 1-2 Ma. Glacial diamict
352 recovered in a west Greenland marine core from ~ 1.8 Ma has ^{10}Be concentrations as low as our

353 cobbles (mean= $4.6 \pm 2.0 \cdot 10^3$ atoms g^{-1} , $n = 5$, 1SD, (Christ et al., 2019)). The general decrease
354 in ^{10}Be concentration over time and the low concentrations of ^{10}Be in Pleistocene and modern
355 subglacial materials portrays the Greenland Ice Sheet as an erosive system that has (at least in
356 certain locations) progressively excavated down into deep, seldom-exposed bedrock or
357 sediments that are now being transported to the margin.

358 Although cosmogenic nuclide analyses of detrital sediments generally indicate deep
359 glacial erosion, cosmogenic nuclide analyses of in-place bedrock around Greenland
360 demonstrate that the ice sheet erodes its bed in some areas and not in others, highlighting that
361 the cobbles are biased toward deeply eroded sediment source areas. The bottom of the GISP2
362 ice core from central Greenland preserves bedrock with an order of magnitude more ^{10}Be (9.8 -
363 $24.8 \cdot 10^3$ atoms g^{-1} , Schaefer et al. (2016)) than the subglacial cobbles considered here. Around
364 Greenland's margins there are isolated regions of non-eroded bedrock and resulting high
365 concentrations of ^{10}Be , particularly at high elevations where the ice was likely cold-based and
366 non-erosive, as demonstrated in both west (Beel et al., 2016; Corbett et al., 2013) and east
367 (Håkansson et al., 2009) Greenland. Less erosive areas of the ice sheet do exist, both around
368 the margins and in the interior, but are not represented in the detrital record since those
369 uneroded, high-cosmogenic-nuclide-concentration landscapes are still in place.

370

371 *6.3. Nuclide Production During Shielding*

372 The presence of measurable in situ cosmogenic ^{14}C in the cobbles unequivocally
373 demonstrates recent (within several tens of ka) production of cosmogenic nuclides. Because
374 the half-life of ^{14}C is so short (5.7 ka), any ^{14}C from before ~ 30 ka has largely decayed away;

375 thus, the ^{14}C we measured indicates nuclide production during the latest Pleistocene or
376 Holocene. Recent nuclide production is further evidenced by the significant linear relationship
377 between ^{14}C with ^{10}Be and ^{26}Al (Fig. 7), as recent surface or near-surface exposure and the
378 resulting co-production of ^{14}C , ^{10}Be , and ^{26}Al is the only mechanism to produce a correlation
379 between the short-lived and the long-lived nuclides.

380 It is likely that this nuclide production occurred when the cobbles were partially
381 shielded. The slope of the $^{14}\text{C}/^{10}\text{Be}$ regression formed by most of the cobbles is 5.9 ($R^2 = 0.92$, n
382 $= 8$, Fig. 7), appreciably higher than the commonly accepted surface $^{14}\text{C}/^{10}\text{Be}$ production ratio of
383 ~ 3 -4 (Argento et al., 2013; Briner et al., 2014; Schimmelpfennig et al., 2012). The higher than
384 expected $^{14}\text{C}/^{10}\text{Be}$ is explained best by muogenic production during shielding (Rand and
385 Goehring, 2019), likely under a minimum shielding mass of at least ~ 200 -250 g cm^{-2} (Fig. 2).
386 Even four cobbles with no measurable ^{10}Be contain ^{14}C (Fig. 7, see also Supplementary Data for
387 detail), supporting the idea that the cobbles were initially sourced from depth and remained
388 shielded. Because muon production is a larger component of the total production in ^{14}C than in
389 ^{10}Be , it allows for the production of ^{14}C in the near-absence of ^{10}Be production under shielding.

390 Our results do not provide direct information about the composition of the material that
391 shielded the samples during the past ~ 30 ka, but we infer that ice is the most likely. The
392 observation that most cobbles (all except GK015 and GU010) fall so closely along a line (Fig. 7)
393 indicates that the cosmogenic nuclides they contain were produced at the same time and under
394 similar shielding conditions. One possibility is that nuclide production occurred under partial
395 shielding during the cobbles' journeys along glacial flow lines that took them upward and
396 outward toward the glacial margin. But regardless of whether shielding occurred by rock,

397 sediment, ice, or a combination thereof, the high $^{14}\text{C}/^{10}\text{Be}$ ratios generally reflect a cobble life
398 cycle that is dominated by shielding and nuclide production at depth by muons.

399

400 *6.4. Interglacial Ice Sheet Retreat and Exposure Preceding the Last Glacial Period*

401 Based on multiple nuclide data, only two of the 86 cobbles we analyzed unambiguously
402 experienced surface or near-surface exposure prior to ~30 ka: GK015 (subangular gneiss from
403 Kangerlussuaq) and GU010 (subangular granodiorite from Upernavik), both of which we
404 collected directly from the ice sheet margin. Both cobbles' $^{14}\text{C}/^{10}\text{Be}$ ratios can be explained with
405 spallogenic rather than exclusively muogenic production (Figs. 6 and 7). The cobbles must have
406 experienced at least some recent exposure, likely at depth as discussed above and causing
407 them to contain measurable ^{14}C , but they were also exposed prior to ~30 ka.

408 To assess exposure before ~30 ka, we can correct for recent nuclide production at depth
409 by removing the ^{10}Be and ^{26}Al that would have been co-produced with the measured ^{14}C . We
410 made this correction by using the measured ^{14}C concentration and the slopes of the $^{14}\text{C}/^{10}\text{Be}$
411 and $^{14}\text{C}/^{26}\text{Al}$ regressions to infer the concentrations of ^{10}Be and ^{26}Al that were co-produced with
412 ^{14}C , and subtracting in order to estimate ^{10}Be and ^{26}Al before ~30 ka. This correction yields
413 $^{26}\text{Al}/^{10}\text{Be}$ ratios of 5.76 ± 0.36 for GK015 and 6.80 ± 0.26 for GU010.

414 After correcting the ^{10}Be and ^{26}Al concentrations for recent production, the pre-~30 ka
415 exposure/burial histories of these two cobbles likely differ in both duration and timing. For
416 sample GK015, the corrected $^{26}\text{Al}/^{10}\text{Be}$ ratio (5.76 ± 0.36) is below the $^{26}\text{Al}/^{10}\text{Be}$ production ratio
417 (regardless of whether we assume a surface production ratio of 7.3 or 6.75) beyond 1σ
418 uncertainties, which could be explained by either burial following exposure or prolonged

419 exposure that caused the $^{26}\text{Al}/^{10}\text{Be}$ ratio to drop due to the shorter half-life of ^{26}Al (Balco et al.,
420 2014). In the case of the former, initial exposure could not have occurred exclusively during
421 MIS5e because the duration of intervening burial would be insufficient to cause a detectable
422 departure from the production ratio, although it could have been re-exposed during MIS5e in a
423 multi-stage exposure scenario. Sample GU010 has a higher corrected $^{26}\text{Al}/^{10}\text{Be}$ ratio ($6.80 \pm$
424 0.26), likely indicating surface exposure during a more recent warm period (and perhaps during
425 older periods as well). Its ^{10}Be concentration is the highest of the 86 cobbles we measured,
426 requiring at least ~ 25 ka of surface exposure at sea level (less at higher elevations); accordingly,
427 this inventory of ^{10}Be could not have built up only during the Holocene Climatic Optimum. Its
428 cosmogenic nuclide inventory is a product of multiple periods of exposure, likely including
429 MIS5e and perhaps previous interglacials as well.

430

431 *6.5. Cosmogenic Nuclide Inheritance and Implications for Dating Studies*

432 Measurable, if low, in situ cosmogenic nuclide concentrations in most subglacial cobbles
433 imply that even extensive, long-lived glacial erosion is unable to fully “reset” the cosmogenic
434 clock (Briner et al., 2016; Davis et al., 1999; Rand and Goehring, 2019). Our dataset provides
435 two complementary lines of evidence for the presence of muon-produced nuclides: (1) low but
436 pervasive ^{10}Be concentrations (median 10^3 atoms g^{-1}) and (2) $^{14}\text{C}/^{10}\text{Be}$ and $^{14}\text{C}/^{26}\text{Al}$ co-
437 production at ratios definitively higher than those of surface production. Small concentrations
438 of primarily muon-produced nuclides are likely present in many glacial environments even after
439 extensive subglacial erosion, not only in moraine boulders and glacially-sculpted bedrock as is

440 frequently observed (Balco, 2011), but also in detrital sediment (Goehring et al., 2010; Nelson
441 et al., 2014).

442 Although inherited ^{10}Be is probably present in most cobbles transported to the ice sheet
443 margin, its impact on inferred exposure ages depends on the relative portion of inherited ^{10}Be
444 versus ^{10}Be produced during the current period of exposure (Fig. 9). The median subglacial
445 cobble ^{10}Be concentration is $\sim 10^3$ atoms g^{-1} , equivalent to about 250 years of surface exposure
446 at high latitude; this represents a relatively small effect on the age of a latest Pleistocene
447 moraine, but a very appreciable effect on the age of a Little Ice Age moraine. However, a subset
448 of the subglacial cobbles we analyzed contained higher ^{10}Be concentrations, which would
449 appreciably skew exposure ages for any dating application; for example, GU010 contains $\sim 10^5$
450 atoms g^{-1} of ^{10}Be , the equivalent of ~ 25 ky of surface exposure at sea level and high latitude.

451 The nine surficial cobbles we collected and analyzed exhibit scatter in their ^{10}Be
452 concentrations and inferred ages (Fig. 9, Table S5). ^{10}Be concentrations of these cobbles ($n = 3$
453 per site) are not consistent within 1σ analytic uncertainties (Fig. 9), despite being collected
454 within close proximity. Such variance could reflect shielding (e.g., by snow, ice, or till since
455 deposition, which may be more important for small cobbles than for large boulders) and/or the
456 presence of inherited produced ^{10}Be .

457 Cobble exposure ages (Fig. 9, Table S5) agree with independent estimates of
458 deglaciation timing better in certain locations than in others. In Kangerlussuaq, the mean
459 surficial cobble age (6.8 ± 0.8 ka; Fig. 9) is consistent with the age of deglaciation inferred from
460 the moraine chronology of Levy et al. (2012). In Ilulissat, the mean surficial cobble age (8.1 ± 0.7
461 ka; Figs. S1E and 9) is indistinguishable from a bedrock sample (GL080; 7.9 ± 0.2 ka) and a

462 boulder sample (GL081; 7.6 ± 0.1 ka) collected from the same location (Corbett et al., 2011).
463 These observations are consistent with the findings of Briner et al. (2013), who reported cobble
464 exposure ages similar to boulder and bedrock ages in central western Greenland, a landscape
465 deeply scoured by erosive warm-based ice. However, in Upernavik, the mean surficial cobble
466 age (7.8 ± 1.6 ka; Figs. S1F and 9) is younger than a bedrock sample (GU001; 13.6 ± 0.3 ka) and
467 a boulder sample (GU002; 10.6 ± 0.3 ka) collected from the same location (Corbett et al., 2013).
468 This offset likely reflects the effect of snow or till cover on the small cobble samples and/or ^{10}Be
469 inheritance in the bedrock and boulder samples. The Upernavik area shows strong evidence for
470 non-erosive, cold-based ice (Beel et al., 2016; Corbett et al., 2013), which may lead to
471 disagreement between exposure ages inferred from bedrock, boulders, and cobbles.

472

473 **7. Conclusions**

474 Analysis of cosmogenic nuclides in 86 subglacial cobbles and 9 surficial cobbles from
475 western Greenland demonstrates that detrital material currently emerging at the ice sheet
476 margin has cosmogenic nuclide concentrations generally indicative of deep erosion. Most
477 subglacial cobbles contain little ^{10}Be , only $\sim 10^3$ atoms g^{-1} , suggesting they were sourced from
478 depth and have experienced little exposure since they were quarried. Although less erosive
479 areas of the ice sheet exist and are documented in other studies, they are not well-represented
480 in detrital sediment samples, which originate from areas of warm-based, erosive ice.
481 Measurable ^{14}C in some subglacial cobbles indicates recent nuclide production, within the past
482 ~ 30 ka; however, $^{14}\text{C}/^{10}\text{Be}$ ratios above that of surface production indicate that nuclide
483 production occurred under shielding. Only two subglacial cobbles have $^{14}\text{C}/^{10}\text{Be}$ and $^{14}\text{C}/^{26}\text{Al}$

484 ratios indicative of excess longer-lived nuclides; their ^{10}Be and ^{26}Al concentrations can be
485 explained by surface or near-surface exposure predating the timespan recorded by ^{14}C . Surficial
486 cobbles exhibit scatter in their ^{10}Be concentrations beyond analytic uncertainties, and match
487 deglaciation age estimates better in certain areas than in others. Overall, the nuclide
488 concentrations of 95 glacial cobbles demonstrate that muon-produced nuclides are pervasive
489 even in long-buried and deeply-eroded landscapes. Although inherited ^{10}Be is generally present
490 in small concentrations, it is occasionally present in concentrations high enough to influence
491 exposures ages.

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Table and Figure Captions

Table 1. Isotopic concentrations and uncertainties for the cobbles with ^{10}Be , ^{26}Al , and ^{14}C data. All 95 ^{10}Be measurements are shown in Table S1. Analysis details including measured ratios, background-corrected ratios, AMS cathode numbers, and primary standards are shown in Tables S1 (for ^{10}Be), S2 (for ^{26}Al), and S4 (for ^{14}C). Blanks for ^{10}Be and ^{26}Al are detailed in Table S3.

Figure 1. Map of Greenland showing the three locations from which cobble-sized rocks were collected from the present-day ice sheet margin. “Icebound” cobbles were embedded directly in the ice, whereas “outwash” cobbles are from large outwash tunnels proximal to the ice sheet margin; both are “subglacial”. Conversely, “surficial” cobbles are from the proglacial landscape and have presumably been exposed since deglaciation. Also shown are other cosmogenic isotope records of detrital sediments as discussed in the text.

Figure 2. Theoretical models of ^{10}Be production (dark gray lines) and ^{14}C production (light gray lines) by both spallation (thick lines) and muons (thin lines) as a function of depth. Shown also is the resulting $^{14}\text{C}/^{10}\text{Be}$ ratio (heavy black line). Depth is expressed in terms of mass depth. All curves assume sea level production in central western Greenland. ^{14}C spallation production rates are derived from measurements of CRONUS-A material extracted in the Tulane cosmogenic nuclide lab (B.M. Goehring, unpublished data, $n = 20$), and ^{14}C muon production rates are from (Balco, 2017).

Figure 3. Probability density function of ^{10}Be concentrations of subglacial cobbles ($n = 73$ above detection limit). Thin gray lines represent the measured isotopic concentrations and internal uncertainties for each sample; thick black line represents the summed probability. Sample names are shown for the cobbles with the highest ^{10}Be concentrations (see Table 1 for detail). Inset: Histogram of ^{10}Be concentrations of the 86 subglacial cobbles we analyzed for ^{10}Be (including 13 that were below detection limit); note logarithmic scale on the x-axis.

Figure 4. Box plots based on four different metrics (location, type, angularity, and lithology) for describing the ^{10}Be concentrations of the subglacial cobbles. Each population includes 73 total subglacial cobbles that were above detection limit for ^{10}Be . The heavy black line shows the mean, while the dashed black line shows the median; the top and bottom of the box show the mean $\pm 1\text{SD}$.

Figure 5. ^{26}Al - ^{10}Be paired nuclide plot for 13 subglacial cobbles. The thick and thin black curves show the continuous exposure pathway and steady-states with respect to steady erosion endpoints respectively for the Greenland $^{26}\text{Al}/^{10}\text{Be}$ production ratio of 7.3 (based on Corbett et al. (2017)). The thick and thin gray curves show the constant production pathway and erosion

endpoints for the commonly-assumed $^{26}\text{Al}/^{10}\text{Be}$ production ratio of 6.75. Error bars show $\pm 1\text{SD}$.

Figure 6. Paired $^{14}\text{C}/^{10}\text{Be}$ plot for 10 subglacial cobbles with detectable ^{14}C presented in terms of production rate normalized $^{14}\text{C}/^{10}\text{Be}$ ratio and ^{10}Be concentrations. Normalization was made assuming the ^{14}C and ^{10}Be production rates for sea level and high latitude. Error ellipses are shown at the 68% confidence level. All but two of the samples plot above the field of continuous exposure, one sample is consistent with continuous exposure, and another sample is consistent with at least one period of exposure and burial.

Figure 7. Linear regressions of ^{14}C concentration versus ^{10}Be concentration (top panel, $n = 10$) and ^{26}Al concentration (bottom panel, $n = 9$) for subglacial cobbles. Regressions are for samples symbolized with gray dots; those with white dots (samples with ^{10}Be below detection limit) and black dots (samples enriched in the long-lived isotopes) are not included in the regression.

Figure 8. Simulated ^{10}Be concentrations and $^{26}\text{Al}/^{10}\text{Be}$ ratios for Pleistocene exposure scenarios #1 (dark gray) and #2 (light gray) from Schaefer et al. (2016). The two bars at the top show exposure scenarios, with burial during gray intervals and exposure during white intervals. The upper panel shows the simulated ^{10}Be concentrations and $^{26}\text{Al}/^{10}\text{Be}$ ratios at the surface of a bedrock column subjected to glacial erosion rates ranging from 5 to 50 m Myr^{-1} under thick ice cover (i.e., no production during burial intervals). Nuclide concentrations in the bedrock column were in steady state with 20 m Myr^{-1} erosion at the start of the simulations; different pre-glacial erosion rates would shift the curves up or down only modestly by the end of each simulation. The kernel density estimates along the left side of the figure show our measured glacial cobble data (as in Fig. 3). The lower panel is the same as the upper panel, but simulates low-level nuclide production by muons through thin, 50-m ice cover during burial intervals. For individual views of any of the above simulations, see Figs. S1 and S2.

Figure 9. ^{10}Be concentrations (top panel) and inferred exposure ages (bottom panel) of surficial cobbles from well outside the modern-day ice margin ($n = 3$ per site, detail in Table S5). Cobbles at each site were collected from the same location, all within several meters of one another. Error bars show 1σ analytic uncertainties (not visible in all cases). Gray lines show the average concentration/age at each site, and the gray box shows $\pm 1\text{SD}$. Dashed lines denote comparisons. Photographs of the sites at which cobbles, the bedrock surface, and a boulder were all sampled are shown in Fig. 2E (Ilulissat) and 2F (Upernavik).

Sample Name	¹⁰ Be	1σ ¹⁰ Be	²⁶ Al	1σ ²⁶ Al	²⁶ Al/ ¹⁰ Be	1σ ²⁶ Al/ ¹⁰ Be	¹⁴ C	1σ ¹⁴ C	¹⁴ C/ ¹⁰ Be	1σ ¹⁴ C/ ¹⁰ Be
	Concentration (atoms g ⁻¹) ^a	Uncertainty (atoms g ⁻¹) ^a	Concentration (atoms g ⁻¹) ^b	Uncertainty (atoms g ⁻¹) ^b	Ratio	Uncertainty	Concentration (atoms g ⁻¹) ^c	Uncertainty (atoms g ⁻¹) ^c	Ratio	Uncertainty
GK015	3.26E+04	5.67E+02	2.16E+05	1.30E+04	6.64	0.42	8.32E+04	8.60E+03	2.56	0.27
GK022	1.10E+04	3.09E+02	7.28E+04	4.44E+03	6.64	0.45	1.24E+05	9.50E+03	11.31	0.92
GK040	4.85E+03	4.25E+02	2.41E+04	4.95E+03	4.98	1.11	6.48E+04	1.74E+04	13.37	3.78
GK051	8.32E+03	2.76E+02	ND	ND	ND	ND	9.09E+04	2.52E+04	10.92	3.05
GK070	4.29E+03	3.04E+02	3.44E+04	3.06E+03	8.03	0.91	BDL	BDL	ND	ND
GK071	4.06E+03	2.25E+02	3.08E+04	4.31E+03	7.59	1.14	7.90E+04	8.70E+03	19.45	2.40
GK072	3.53E+03	1.68E+02	2.34E+04	2.96E+03	6.61	0.89	BDL	BDL	ND	ND
GK097	1.82E+04	4.12E+02	1.37E+05	1.18E+04	7.52	0.67	ND	ND	ND	ND
GK099	1.18E+04	3.09E+02	8.67E+04	6.85E+03	7.37	0.61	BDL	BDL	ND	ND
GL028	5.00E+03	2.08E+02	2.97E+04	2.61E+03	5.93	0.58	6.91E+04	9.10E+03	13.81	1.91
GL036	3.99E+03	2.15E+02	3.34E+04	4.40E+03	8.38	1.19	6.60E+04	8.50E+03	16.55	2.31
GU010	1.12E+05	2.23E+03	7.71E+05	2.51E+04	6.87	0.26	5.10E+04	1.37E+04	0.45	0.12
GU034	7.38E+03	3.94E+02	3.92E+04	3.21E+03	5.31	0.52	7.81E+04	9.00E+03	10.58	1.34
GU126	1.89E+04	4.18E+02	1.42E+05	1.18E+04	7.52	0.65	1.51E+05	1.50E+04	8.00	0.81

^aThe ¹⁰Be/⁹Be measurements were made at Lawrence Livermore National Laboratory and were normalized to standard 07KNSTD3110 with an assumed ratio of 2.85 x 10⁻¹¹ (Nishiizumi et al., 2007).

^bThe ²⁶Al/²⁷Al measurements were made at Purdue Rare Isotope Measurement Laboratory and were normalized to standard KNSTD with an assumed ratio of 1.818 x 10⁻¹² (Nishiizumi et al., 2004).

^dThe ¹⁴C measurements were made at University of Cologne.

ND = No data (sample failed during measurement yielding no usable data)

BDL = Below detection limit (see text for details)

Table 1.

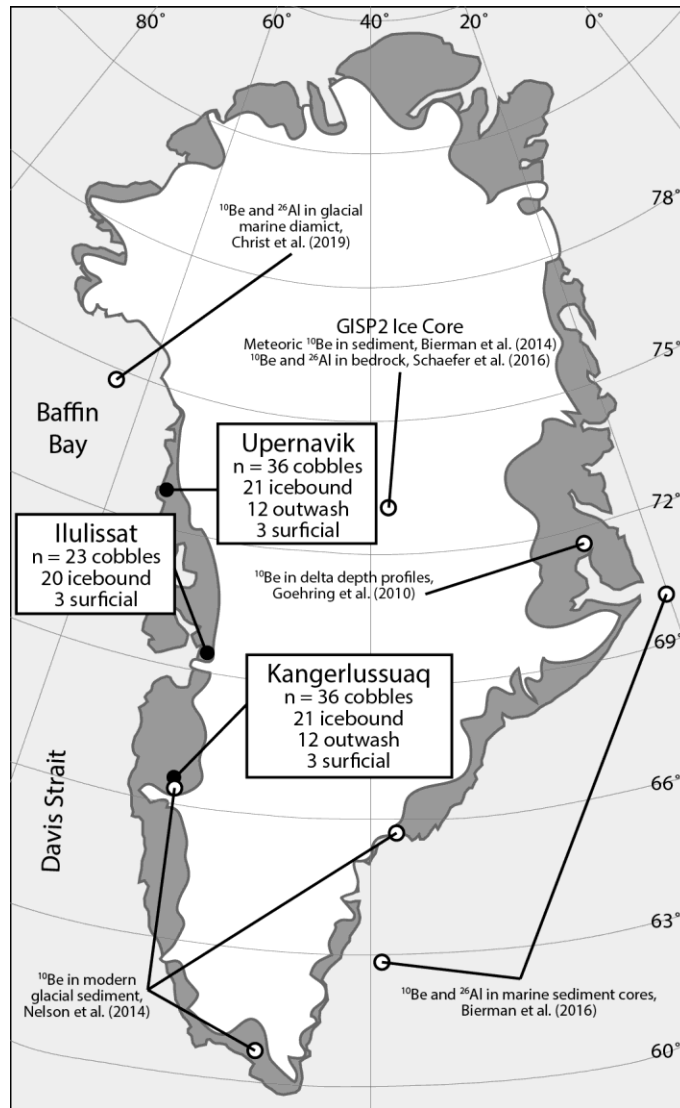


Figure 1.
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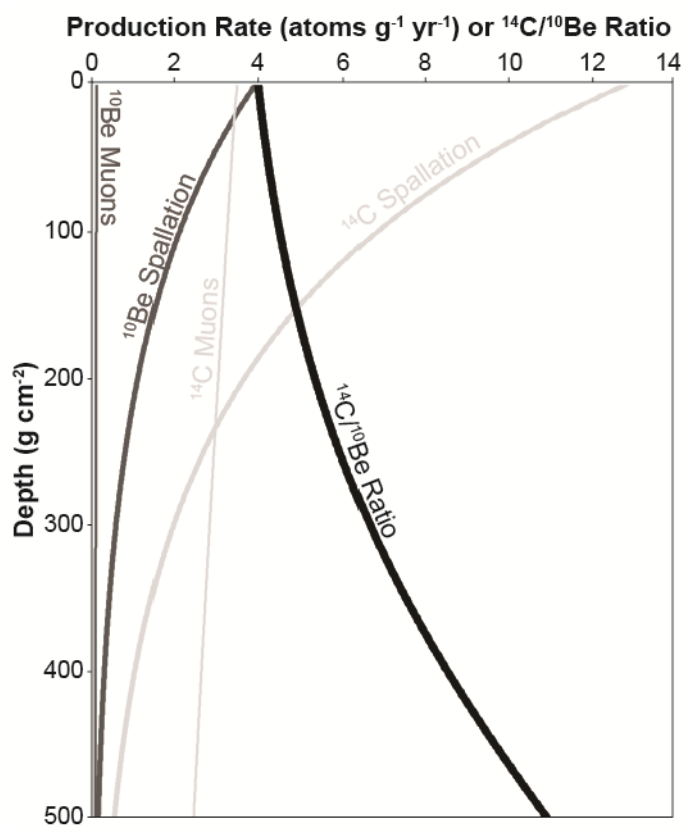


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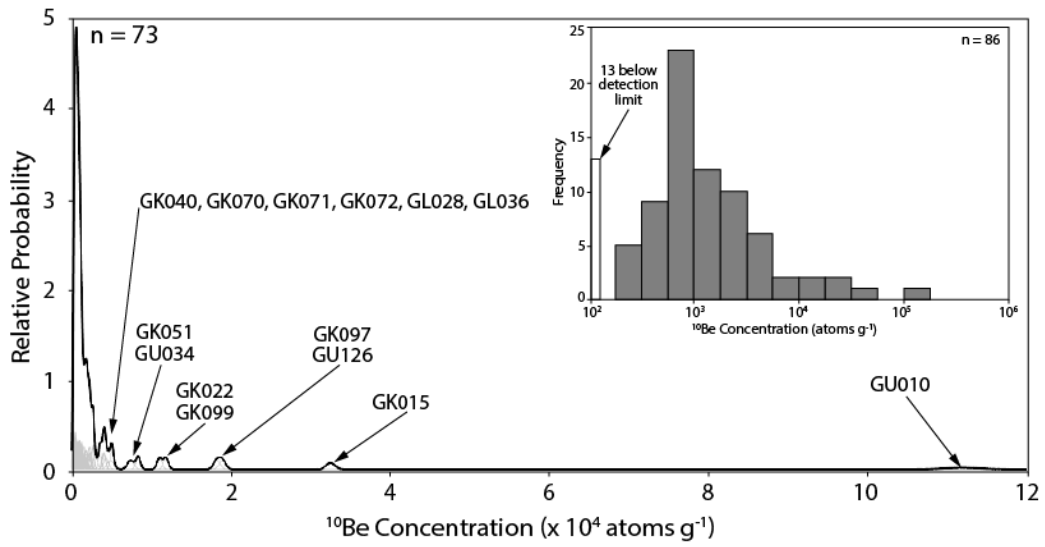


Figure 3.
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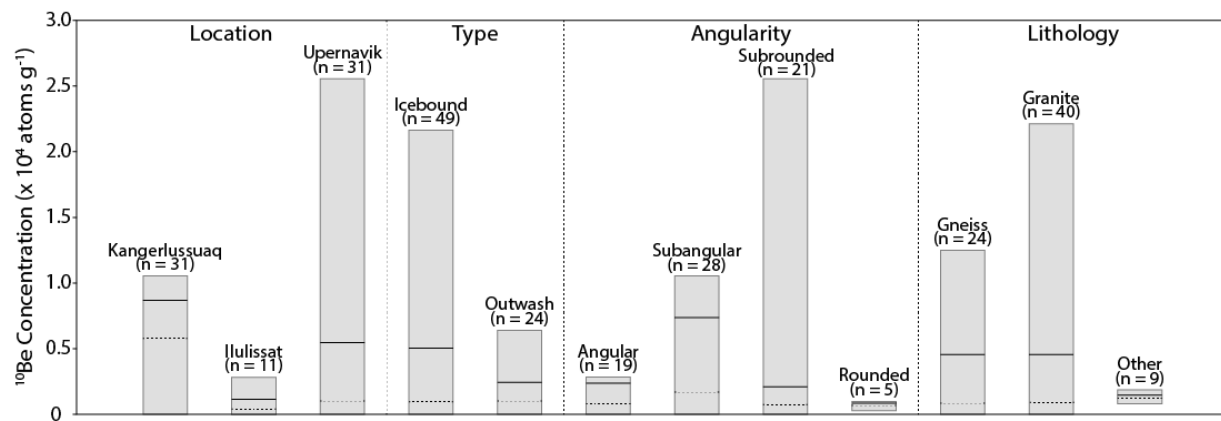


Figure 4.
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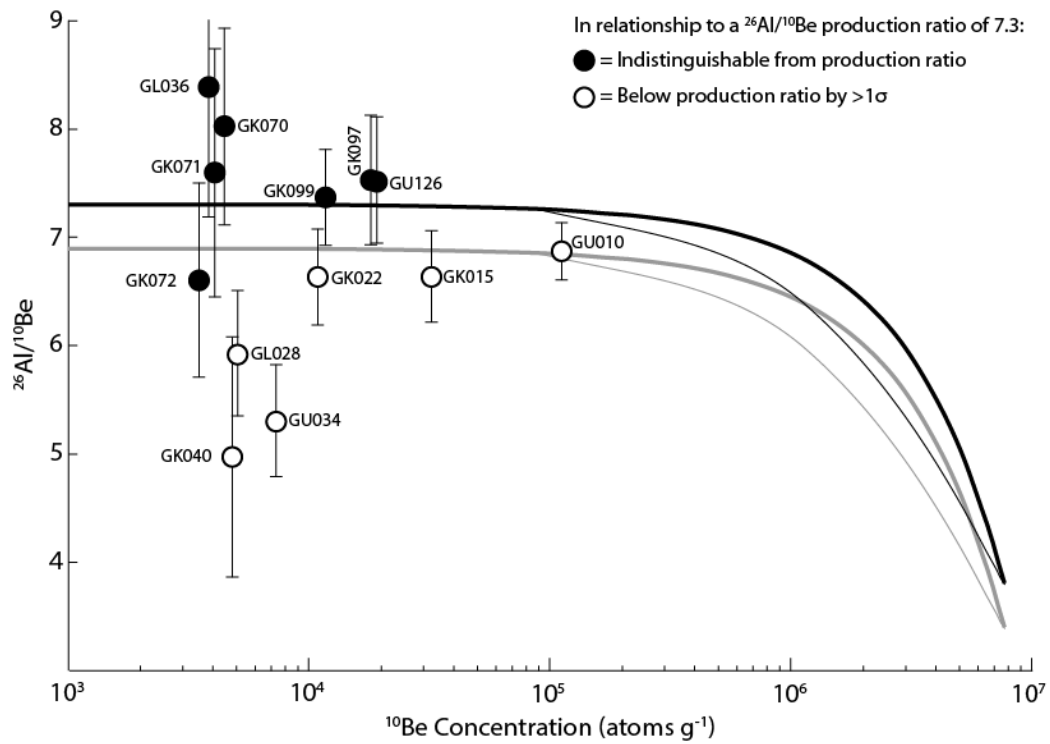


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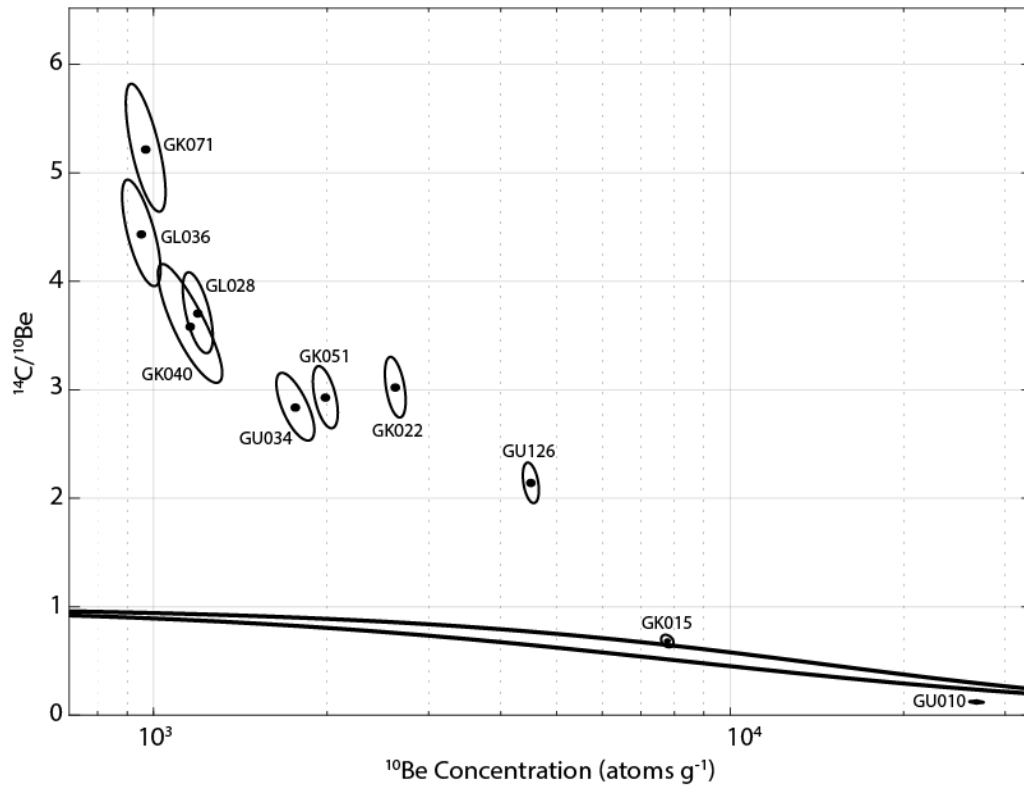


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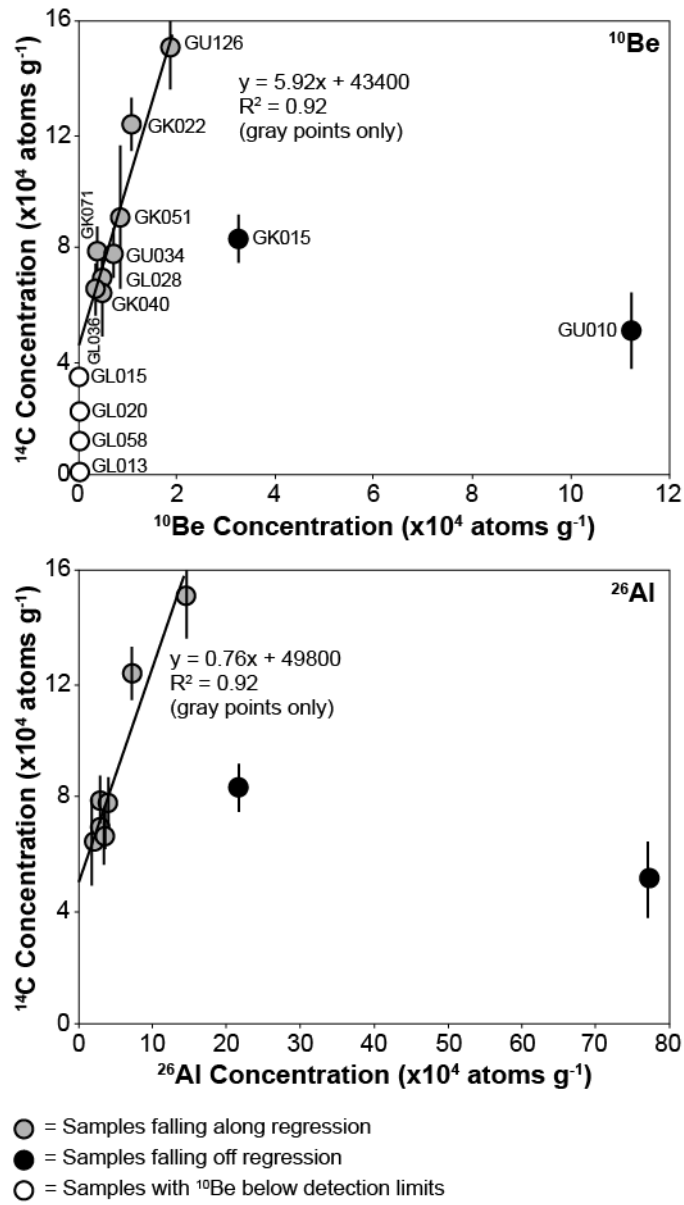


Figure 7.
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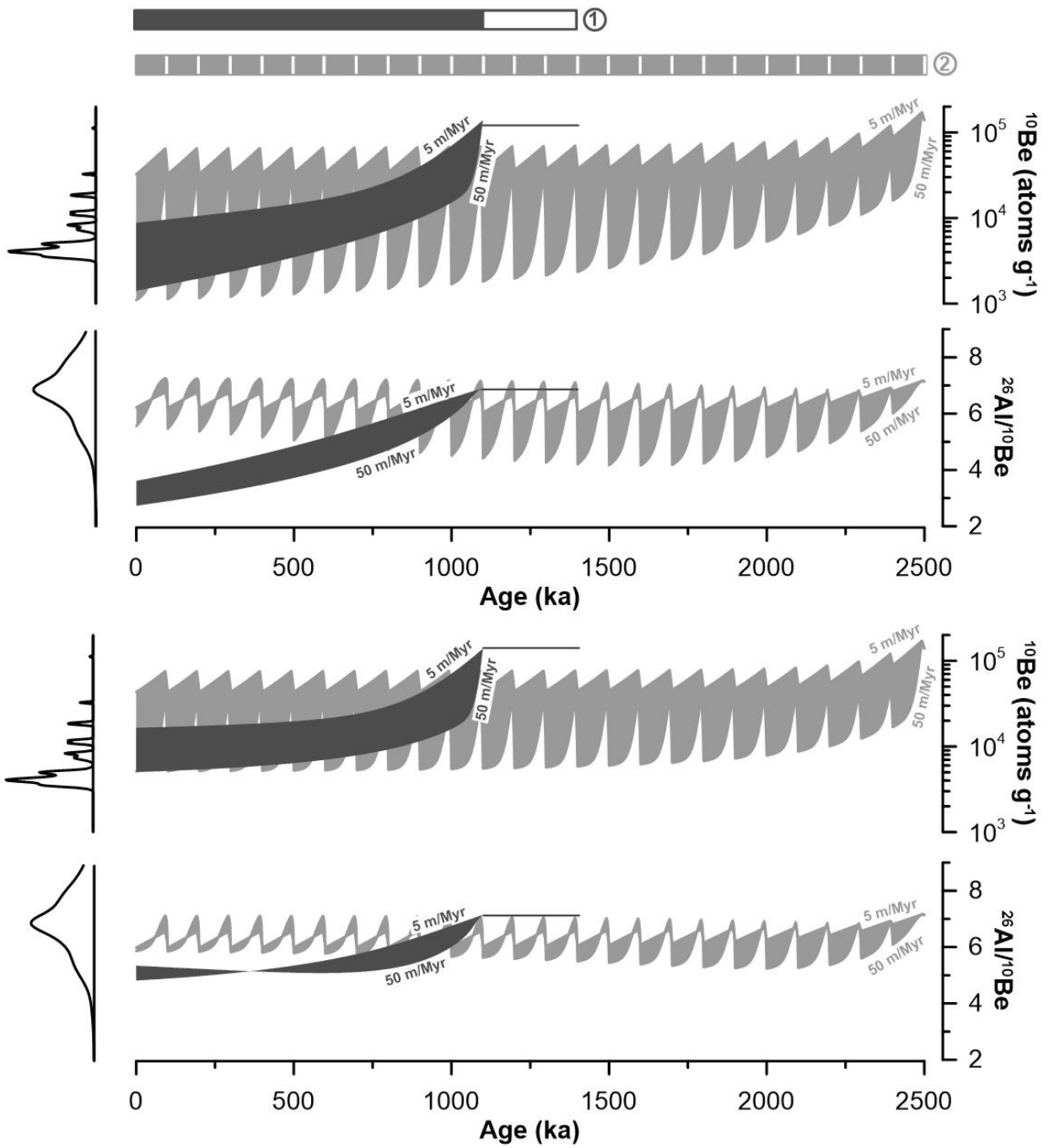


Figure 8.
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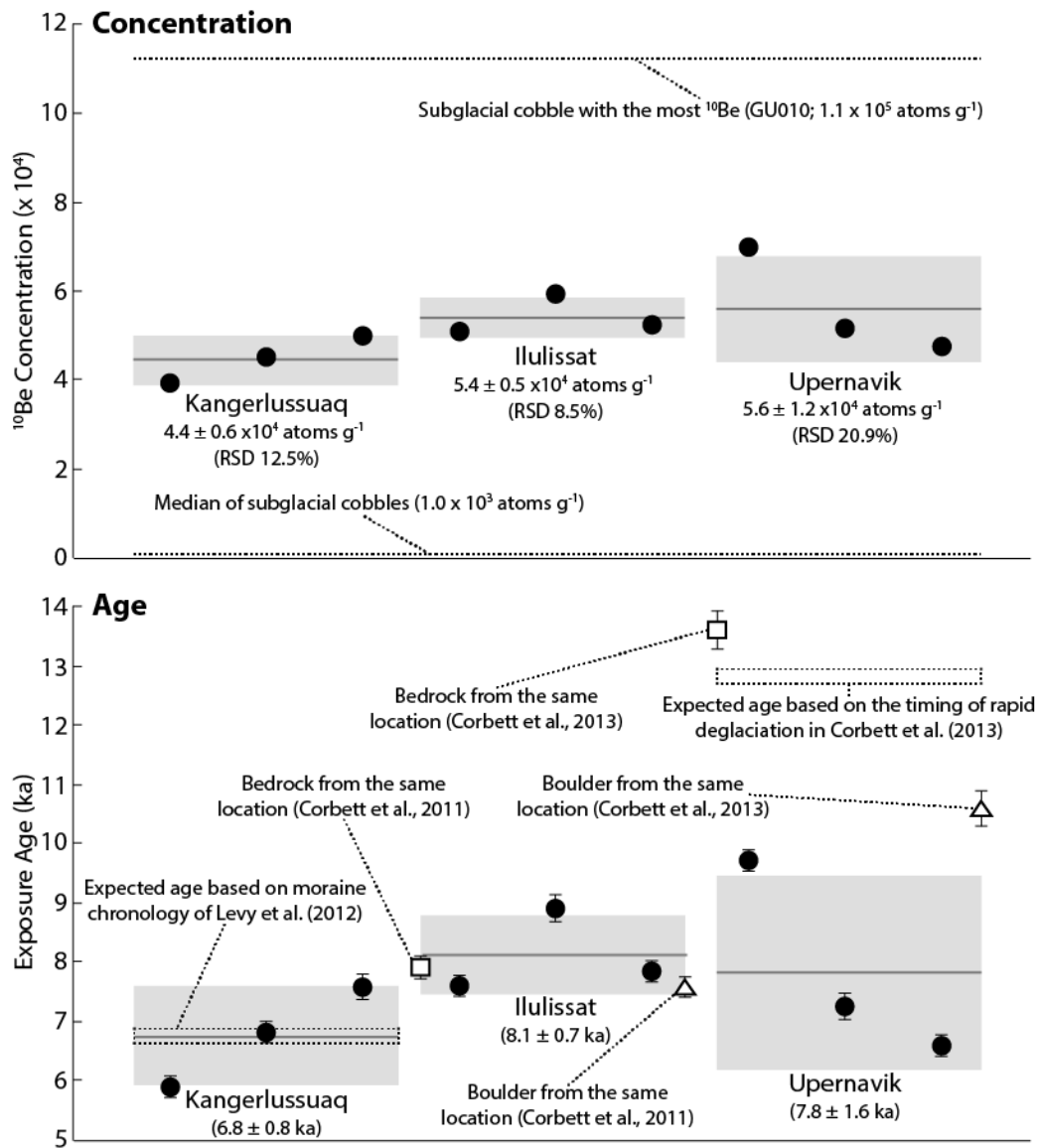


Figure 9.
(Width = 140 mm, 1.5 columns)