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► **To cite this version:**

Quentin Simon, Nicolas Thouveny, Didier Boulès, Laurence Nuttin, Claude Hillaire-Marcel. Authigenic $^{10}\text{Be}/^{9}\text{Be}$ ratios and ^{10}Be -fluxes (^{230}Th -normalized) in central Baffin Bay sediments during the last glacial cycle: Paleoenvironmental implications. *Quaternary Science Reviews*, 2016, 10.1016/j.quascirev.2016.03.0270277-3791 . hal-01456980

HAL Id: hal-01456980

<https://hal.science/hal-01456980>

Submitted on 6 Feb 2017

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1 **Authigenic $^{10}\text{Be}/^9\text{Be}$ ratios and ^{10}Be -fluxes ($^{230}\text{Th}_{\text{xs}}$ -normalized) in central Baffin Bay**
2 **sediments during the last glacial cycle: paleoenvironmental implications**

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10 **Abstract**

11 Authigenic $^{10}\text{Be}/^9\text{Be}$ ratios and ^{10}Be -fluxes reconstructed using the $^{230}\text{Th}_{\text{xs}}$ normalization,
12 proxies of the cosmogenic radionuclide ^{10}Be production rate in the atmosphere, have been
13 measured in a sedimentary core from Baffin Bay (North Atlantic) in order to reconstruct the
14 geomagnetic dipole moment variations during the last ca. 136 ka BP, and for comparison with
15 the relative paleointensity (RPI) record derived from paleomagnetic measurements. Our study
16 revealed that the exchangeable (authigenic) ^{10}Be measured includes a strong climatic
17 component related to the glacial dynamics that characterized the circum Baffin Bay during the
18 last glacial period. Despite normalization applied on the authigenic ^{10}Be concentrations using
19 both the scavenged ^9Be and $^{230}\text{Th}_{\text{xs}}$ approaches, a strong climatic signal still prevails. Both
20 normalization methods yield equivalent results that are both strongly correlated with
21 sedimentological parameters (grain-size and mineralogy). The lower $^{10}\text{Be}/^9\text{Be}$ ratio values are
22 associated with coarse-grained carbonate-rich layers while the higher $^{10}\text{Be}/^9\text{Be}$ ratio values
23 are found with fine-grained feldspar-rich sediments. This variability is due to both i) sediment
24 composition control over beryllium-scavenging rates, ii) the glacial history that contributed to
25 modify the ^{10}Be concentration in the oceanic realm and notably boundary scavenging
26 conditions. No pristine geomagnetic field intensity can thus be derived from ^{10}Be
27 measurements in such a high-variability glacio-marine environment. These results also

28 indicate that the straightforward interpretation of ^{10}Be -concentration variations as a proxy of
29 the Interglacial/Glacial (or major interstadials) cycles in Arctic and sub-Arctic regions must
30 be considered with caution and rather propose to relate ^{10}Be variations to higher-frequency
31 paleoclimatic changes and glacial dynamics.

32 **1. Introduction**

33 The cosmogenic nuclide Beryllium-10 (^{10}Be) is produced in the stratosphere (~65%) and the
34 troposphere (~35%) by spallation reactions when highly energetic galactic cosmic rays
35 interact with nitrogen and oxygen atoms (Lal and Peters, 1967; Dunai and Lifton, 2014). Its
36 production rate is linked to the Sun and Earth magnetic fields variabilities by a non-linear
37 inverse relationship (Elsasser et al., 1956; Lal, 1988; Beer et al., 1990; Masarik and Beer,
38 2009; Kovaltsov and Usoskin, 2010). After its production in the atmosphere, ^{10}Be is quickly
39 scavenged onto aerosol particles themselves precipitated (or thrown) within *ca.* 1-3 years into
40 ocean/continent reservoirs by wet or dry deposition processes (Raisbeck *et al.*, 1981; Beer et
41 al., 1990; Baroni *et al.*, 2011). Previous studies have shown that -while solar activity inflects
42 the production rate on shorter timescales- ^{10}Be flux measured along ice and marine sediment
43 sequences reflect long term signatures of the geomagnetic dipole moment (Raisbeck *et al.*,
44 1981, 1985; Yiou *et al.*, 1985; Wagner *et al.*, 2000; Frank et al. 1997; Carcaillet et al., 2003,
45 2004, Muscheler *et al.* 2004, 2005). The exchangeable- ^{10}Be concentrations (*i.e.*, fraction
46 adsorbed onto settling particles) measured in the sediments, henceforth referred to the
47 authigenic ^{10}Be , are not only reflecting the atmospheric production at the time of their
48 deposition but also depend on the scavenging rates from the overlying water column, to
49 advected ^{10}Be fraction due to oceanic mixing and/or to inherited ^{10}Be fraction scavenged
50 during the terrestrial/glacial/meltwater cycling of the carrier particles. Therefore,
51 normalization of the authigenic ^{10}Be concentration is a first step in order to remove these
52 environmental biases and compare ^{10}Be records with geomagnetic variability.

53 Two methods of normalization have been used in literature: (1) normalization of the
54 authigenic ^{10}Be cosmogenic nuclide by the authigenic stable ^9Be isotope supplied by
55 terrigenous material entering the ocean (*i.e.*, originating from the dissolution of detrital,
56 aeolian, riverine and glacier inputs). This method relies on the similar behavior of both
57 isotopes once homogenized in seawater (Bourlès *et al.*, 1989; Brown *et al.*, 1992). The
58 authigenic $^{10}\text{Be}/^9\text{Be}$ ratio method reliably corrects for ocean/continent secondary
59 contributions and provides robust results clearly demonstrating an inverse relationship with
60 the geomagnetic field (Henken-Mellies *et al.*, 1990; Robinson *et al.*, 1995; Carcaillet *et al.*,
61 2003, 2004a, 2004b; Thouveny *et al.*, 2008; Ménébréaz *et al.*, 2011, 2012, 2014; Valet *et al.*,
62 2014). The second normalization process (2) uses the ^{230}Th -excess method proposed by
63 Bacon (1984) to calculate vertical fluxes of any sedimentary component deposited during the
64 Late Quaternary (see François *et al.*, 2004 for a review) and has been used successfully to
65 calculate ^{10}Be -fluxes in numerous studies (*e.g.*, Franck *et al.*, 1997, 2000; Christl *et al.*, 2007,
66 2010). All these studies notably demonstrate that during periods characterized by low dipole
67 strength (*i.e.*, corresponding to episodes of collapsed field during reversals or excursions), the
68 atmospheric ^{10}Be production rates were significantly larger than during periods with high
69 dipole strength. Accordingly, the reconstruction of atmospheric ^{10}Be production rate signals
70 from marine sedimentary sequences constitutes a complementary approach, independent from
71 paleomagnetism, to decipher past geomagnetic dipole moment variations. Furthermore, the
72 comparison between the ^{10}Be signals from mid-to-low-latitude sites (Ménébréaz *et al.*, 2012,
73 2014) as well as with recent simulations using general circulation models (GCMs; Heikkilä *et al.*
74 *et al.*, 2009, 2013) demonstrate a rapid zonal atmospheric mixing of ^{10}Be before its deposition in
75 geological archives.

76 In the Arctic Ocean where most dating methods encounter serious limitations, the radioactive
77 decay rates of the ^{10}Be measured in marine sediments have been used to establish a Neogene

78 chronostratigraphic framework (assuming a near constant supply of ^{10}Be in first
79 approximation) of the ACEX long sedimentary sequence (Frank et al., 2008). Besides,
80 variations in ^{10}Be concentrations in Arctic and sub-Arctic sediments have also been used in
81 order to constrain the stratigraphy of Late Quaternary sedimentary records assuming that
82 variations of ^{10}Be concentrations roughly represent Glacial/Interglacial cycles or major
83 Interstadial periods (Spielhagen et al., 1997, 2004; Sellén et al., 2009; Alexanderson et al.,
84 2014). In this mechanism, the low ^{10}Be concentrations corresponding to glacial periods are
85 caused by a combination of: (1) low inputs of ^{10}Be due to increased sea ice cover and, (2)
86 ^{10}Be dilution related to higher accumulation rates of ice rafted debris (IRD); and vice versa
87 for interglacial (major interstadial) periods.

88 In this study, we present new authigenic $^{10}\text{Be}/^9\text{Be}$ ratios and $^{230}\text{Th}_{\text{xs}}$ -normalized ^{10}Be -fluxes
89 from a high-latitude marine record (HU2008-029-016PC) from central Baffin Bay and
90 spanning the last 136 ka. This paper includes a revision of the initial chronostratigraphy of the
91 core (Simon et al., 2012) in order to improve the robustness of the chronology for the bottom-
92 half of the core. A primary aim of the study was to examine if ^{10}Be originating from
93 continental inputs or advected through oceanic circulation and glacial processes could be
94 removed from the total authigenic ^{10}Be concentrations by means of normalization procedures
95 before any geomagnetic interpretations. We compare the two existing normalization methods
96 (*i.e.*, authigenic ^9Be and $^{230}\text{Th}_{\text{xs}}$) using a U and Th-series isotope record from the same core
97 (Nuttin and Hillaire-Marcel, 2015) and discuss ^{10}Be -systematics *vs.* sedimentological
98 parameters, geomagnetic dipole moment variations, ^{10}Be -fluxes from models and marine/ice
99 records, and, finally, we discuss its paleoenvironmental implications.

100 **2. Environmental setting**

101 Baffin Bay is a subpolar oceanic basin (1300 km long and 450 km wide, $\sim 690\,000\text{ km}^2$)
102 located in the northwest North Atlantic (Figure 1). The bay is one of the main export routes of

103 freshwater from Greenland, the Canadian Arctic and the Arctic Ocean into the North Atlantic
104 Ocean. Its morphology consists of an elongated abyssal plain (2000–2500 m) surrounded by
105 the continental shelves of Greenland and Baffin Island. During the glacial periods, the
106 northeastern Laurentide Ice Sheet (LIS), the Innuitian Ice Sheet (IIS) and the Greenland Ice
107 Sheet (GIS) formed a nearly continuous and highly dynamic ice belt surrounding Baffin Bay
108 (Figure 1). On the Greenland side, the GIS extended westward over the inner shelf, and as far
109 as the shelf edge off Disko Bugt and the Uummannaq Trough during the Last Glacial
110 Maximum (LGM) (Ó Cofaigh *et al.*, 2012, 2013; Funder *et al.*, 2011). The LIS extended
111 through Baffin Island, probably as far as its fjord mouths and inlets, and possibly over part of
112 the Baffin Island shelf during glacial maxima (Margold *et al.*, 2015). In the northern end of
113 the bay, ice streams in the Smith Sound and Lancaster sounds were particularly large and
114 active (England *et al.*, 2006; Klassen and Fisher, 1988; Li *et al.*, 2011) and probably
115 developed into an ice shelf towards the bay (Alley *et al.*, 2010; Marcott *et al.*, 2011).
116 Numerous studies have demonstrated strong relationships between glacial dynamic, oceanic
117 circulation and sedimentary processes in the bay (Aksu, 1985, 1987; de Vernal *et al.*, 1987;
118 Andrews *et al.*, 1998, 2014; Jennings *et al.*, 2014; Simon *et al.*, 2014; Nuttin and Hillaire-
119 Marcel, 2015). Sedimentation occurs through two main processes: (1) glacial plumes from
120 lateral sources with large volumes of fine-grained feldspar-rich sediments, which were
121 transported to the area by sediment-laden supraglacial and subglacial meltwater or nepheloid
122 layers and (2) Trans-Baffin icebergs drifting from the northern end of the bay with large
123 volumes of coarse-grained carbonated sediments, leading to the deposition of so-called Baffin
124 Bay Detrital Carbonate layers (BBDC; Simon *et al.*, 2014).

125 3. Material and methods

126 3.1. Core description

127 Core HU2008-029-016PC (70°46.14 N/-64°65.77 W; PC16 hereinafter) was raised from a
128 2063 m water depth, on the abyssal plain from central Baffin Bay. The 741-cm long core was
129 retrieved using a piston corer during the 2008-029 CCGS Hudson cruise (Campbell and de
130 Vernal, 2009). The sediment sequence is mainly composed of a succession of homogeneous
131 dark gray to olive-gray silty clayey units and of very poorly sorted grayish-brown gravelly
132 and sandy carbonate-rich layers (dolomite rich, Figure 2). These two lithofacies reflect the
133 origin, transport and mode of deposition of the lithogenic sediments related to the ice sheet
134 dynamics evoked above. Moreover, the top of the core down to 20 cm is characterized by
135 brown to dark brown silty muds while the interval between 120 and 215 cm is constituted by
136 brownish-black to olive-black clayey muds. These two distinct lithofacies represent sediments
137 deposited respectively during the Holocene and during marine isotope stage (MIS) 2 (see
138 Simon *et al.*, 2012, 2014, in prep. and Simon, 2013 for additional information).

139 3.2. Paleomagnetic results and chronology

140 The relative paleointensity (RPI) record previously established by a detailed paleomagnetic
141 analysis (Simon *et al.*, 2012) reinforced and completed preliminary paleomagnetic results
142 obtained on a shorter core from the same site (Thouveny 1988). The RPI proxy (Figure 2) is
143 based on the normalization of the Natural Remanent Magnetization (NRM) with the
144 Anhyseretic Remanent Magnetization (ARM) over the 25-35 mT AF demagnetization
145 interval ($\text{NRM}_{25-35\text{mT}} / \text{ARM}_{25-35\text{mT}}$). With the exception of few problematic layers, the ARM
146 normalizer activates the same magnetic assemblages than the NRM, and the RPI proxy
147 presents no correlation with lithological proxies. Moreover, the derived RPI record was
148 favorably correlated with RPI reference curves and stacks in order to establish the initial
149 chronology of the core (Simon *et al.*, 2012). These reference records included mainly the

150 North Atlantic relative paleointensity stack NAPIS-75 (Laj *et al.*, 2000) and the Labrador Sea
151 core MD95-2024 (Stoner *et al.*, 2000). Two large inclination variations coeval with
152 paleointensity lows allowing the recognition of two major geomagnetic excursions, *i.e.*, the
153 Laschamp and Norwegian-Greenland-Sea excursions, and three-radiocarbon ages further
154 supported the age model (see Simon *et al.*, 2012 for details). This initial chronology is
155 considered robust for the MIS 2-3-4 periods with a high correlation between PC16 and the
156 NAPIS-75 stack ($r=0.82$ between 22-75 ka, Figure 2) and remains unchanged in this study.
157 For the lower part of the core, large chronological uncertainties (correlation coefficients with
158 reference records <0.51) required revision in order to improve its resolution and to compare
159 ^{10}Be production rate variations with references. Therefore, the chronology of core PC16 was
160 updated by tuning the PC16 RPI curve with the ODP 1063 RPI record (Channel *et al.*, 2012)
161 using 13 tie points ($r=0.70$ between 75-136 ka, Figure 2). The age model for ODP Site 1063
162 was constructed by tandem correlation of oxygen isotope and RPI data to calibrated reference
163 templates using the Match protocol (Lisiecki and Lisiecki, 2002), improving its reliability.
164 The revised age model offers a significantly improved statistical robustness and places the
165 PC16 core bottom (741 cm) in the MIS 6 interval representing a 20 ka age shift from the
166 Simon *et al.* (2012) age model. Using this refined age model, the average sedimentation rate
167 for the sedimentary sequence is 5.4 cm/ka and presents large variability mainly related to
168 glacial/deglacial history. The mean sedimentation rates during the Holocene (0-10.6 ka),
169 Deglacial-peak (10.7-12 ka) and last Glacial (12.1-136.7 ka) periods are 1.9 cm/ka, 25.8
170 cm/ka and 5.5 cm/ka, respectively (Figure 8).

171 3.3. Sample preparation

172 Based on the paleomagnetic record and on U and Th-series isotope records, 76 subsamples of
173 ~1 g (dry sediment) were collected along core PC16 and processed for the Be isotope analysis
174 at the CEREGE National Cosmogenic Nuclides Laboratory (France) according to the

175 chemical procedure set-up by Bourlès *et al.* (1989) and summarized in Carcaillet *et al.* (2003,
176 2004a, 2004b) and Ménabréaz *et al.* (2011, 2012, 2014). The method is detailed here since the
177 separation procedure has been modified prior to the AMS measurements. ^{10}Be and its stable
178 isotope ^9Be were co-extracted from the authigenic phase of the sediments using 20 ml.g⁻¹
179 sediment of 0.04 M hydroxylamine (NH₂OH-HCl) in a 25% acetic acid leaching solution at
180 $95 \pm 5^\circ\text{C}$ for 7 hrs. A 2 ml aliquot of the resulting leaching solution was sampled for the
181 measurement of the natural ^9Be concentration. The remaining solution was spiked with 300 μl
182 of a $9.8039 \cdot 10^{-4}$ g.g⁻¹ ^9Be -carrier before the chemical extraction in order to accurately
183 determine ^{10}Be sample concentrations from the measured $^{10}\text{Be}/^9\text{Be}$ ratios. The Be-purification
184 was realized by chromatography in two stages. Before each separation stages, the samples
185 were evaporated and then dissolved in ultra-pure HCl. Be oxy-hydroxides were precipitated at
186 pH 8.5 from the solution by adding NH₃. The precipitate was separated by centrifugation,
187 dissolved in ultra-pure HCl and then loaded onto an exchange column. The iron (Fe) and
188 manganese (Mn) were separated using a Dowex® 1x8 (100–200 mesh) anion-exchange resin.
189 The resin was first rinsed with 20 ml MilliQ water and conditioned with 20 ml 10.2 M HCl.
190 The sample was then loaded onto the column and the Be fraction was collected immediately
191 using 20 ml 10.2 M HCl for elution. The next purification step was carried out on a Dowex®
192 50x8 (100–200 mesh) cation-exchange resin in order to separate the Bore (B) and Aluminum
193 (Al). The resin was rinsed with 30 ml MilliQ water and then conditioned with 30 ml 1 M HCl.
194 After sample loading onto the column, the B and Be were eluted successively within the first
195 40 ml and next 120 ml of 1 M HCl eluent while the Al remained trapped within the column.
196 After the two separation stages, Be oxy-hydroxides were precipitated at pH 8.5 from the final
197 solution by adding NH₃. The precipitate was separated by centrifugation, rinsed by re-
198 suspension using pH 8.5 buffered MilliQ water and centrifugated again. The purified Be oxy-
199 hydroxides were solubilized in HNO₃ and the resulting solution was transferred into a quartz

200 crucible where it was gently evaporated to dryness at 200°C. Finally, the Be oxy-hydroxides
201 deposit was oxidized and converted to BeO by heating at 800°C for 1 hr. The BeO was then
202 mixed with Nb powder and pressed into a cleaned Ti cathode-holder in order to condition the
203 samples for AMS measurements. In addition to sample processing, several routine blanks and
204 2 replicates were measured in order to assess both cleanliness and reproducibility during the
205 chemical extraction.

206 3.4. Measurements

207 The natural authigenic ^9Be concentrations were measured using a graphite-furnace Atomic
208 Absorption Spectrophotometer (AAS) with a double beam correction (Thermo Scientific ICE
209 3400®). The standard-addition method and the addition of a constant volume of MgNO_3
210 solution (matrix modifier) were used to eliminate the matrix effects during the absorption and
211 to allow measurements near the detection limit. ^9Be sample concentrations (Table 1) were
212 determined from repeated absorbance measurements (4 times) performed on each of the four
213 100 μl aliquots of the sample solution, three of them being spiked with increasing amount of a
214 Sharlau ^9Be -carrier diluted to a known concentration ($0.27 - 0.34 \times 10^{-8} \text{ g.g}^{-1}$) using HNO_3
215 0.2%. The standard deviation of repeated absorbance measurements for each sample must be
216 less than 3% to be accepted. After correcting for sample dilution, the authigenic ^9Be sample
217 concentrations along core PC16 vary around $1.99 \pm 0.83 \cdot 10^{-7} \text{ g.g}^{-1}$. The associated
218 uncertainties (2σ) varying from 0.4 to 3.2% (average value: 1.5%) are based on the
219 reproducibility of measurements and the least-square fitting between measured absorbance at
220 each stages of the standard-addition method ($r^2 > 0.9995$).

221 The natural authigenic ^{10}Be concentration measurements were performed at the French AMS
222 national facility ASTER (CEREGE). ^{10}Be sample concentrations are calculated from the
223 measured spiked $^{10}\text{Be}/^9\text{Be}$ ratios normalized to the NIST 4325 Standard Reference Material

224 $(2.79 \pm 0.03 \times 10^{-11}$; Nishiizumi *et al.*, 2007), and are decay-corrected using the ^{10}Be half-life
 225 $(T_{1/2})$ of 1.387 ± 0.012 Ma (Chmeleff *et al.*, 2010; Korschinek *et al.*, 2010):

$$226 \quad \text{Authigenic } \left[\frac{^{10}\text{Be}}{^9\text{Be}} \right]_{at}^{\text{decay-corrected}} = \left(\frac{^{10}\text{Be}}{^9\text{Be}} \right)_M \times \left(\left[^9\text{Be} \right]_{at} + m_{\text{spike}} \times \left[^9\text{Be} \right]_{\text{spike}} \times \frac{N_A}{M^9\text{Be}} \right) \times e^{\left(\frac{\ln(2)}{T_{1/2}} \times t \right)} \quad (1)$$

227 where: $(^{10}\text{Be}/^9\text{Be})_M$ is the measured Be ratio; m_{spike} and $\left[^9\text{Be} \right]_{\text{spike}}$ are respectively the mass
 228 and the concentration of the added spike; N_A is the Avogadro constant ($6.02214 \cdot 10^{23} \text{ mol}^{-1}$);
 229 $M^9\text{Be}$ is the beryllium Molar Mass ($9.0121822 \text{ g} \cdot \text{mol}^{-1}$) and t is the time. The ^9Be
 230 concentrations measured at the AAS are transformed in atoms as follow:

$$231 \quad \text{Authigenic } \left[^9\text{Be} \right]_{at} = \left[^9\text{Be} \right]_{\text{AAS}} \times m'_{\text{sample}} \times \frac{N_A}{M^9\text{Be}} \quad (2)$$

232 where: $\left[^9\text{Be} \right]_{\text{AAS}}$ is the concentration of natural authigenic ^9Be measured at the AAS and

$$233 \quad m'_{\text{sample}} = m_{\text{sample}} \times \left(\frac{m_{\text{leached}} - m_{\text{aliquot}}}{m_{\text{leached}}} \right) \text{ is the weight of sediment remaining after aliquot}$$

234 sampling.

235 The uncertainties (2σ) in the measured $^{10}\text{Be}/^9\text{Be}$ ratios and in the calculated ^{10}Be
 236 concentrations result from statistical and instrumental error propagation (Arnold *et al.*, 2010)
 237 and vary from 1.3 to 3.7% (average value: 2.2%).

238 The authigenic natural $^{10}\text{Be}/^9\text{Be}$ ratios are derived using equations (1) and (2):

$$239 \quad \text{Authigenic } \left(\frac{^{10}\text{Be}}{^9\text{Be}} \right) = \frac{\left[^{10}\text{Be} \right]_{at}^{\text{decay-corrected}}}{\left[^9\text{Be} \right]_{at}} \quad (3)$$

240 The measured and calculated ratios and their uncertainties are presented in Table 1. The
 241 uncertainties (2σ) of the calculated authigenic $^{10}\text{Be}/^9\text{Be}$ ratios are derived from the
 242 propagation of both uncertainties and vary between 2.9 and 9% (average value: 5.5%).
 243 Chemistry blank ratios range from 5.3×10^{-15} to 1.5×10^{-14} , which is at least 3 orders of
 244 magnitude lower than the sample $^{10}\text{Be}/^9\text{Be}$ ratios.

245 3.5. Measurements of $^{230}\text{Th}_{\text{xs}}$ in Baffin Bay sediments

246 In this study, we benefit from recent U and Th-series isotope results from PC16 in order to
247 calculate ^{10}Be -fluxes ($^{230}\text{Th}_{\text{xs}}$ -normalized). The measurements and calculation used for
248 determining $^{230}\text{Th}_{\text{xs}}$ are detailed in Nuttin and Hillaire-Marcel (2015). The estimated initial
249 $^{230}\text{Th}_{\text{xs}}$ activities, recalculated in respect to the revised chronology, are extremely variable
250 ranging from 0.118 ± 0.081 to 5.293 ± 0.212 dpm.g $^{-1}$ (1.145 dpm.g $^{-1}$ on average with a standard
251 deviation of 0.977). Except for few samples, large surplus of $^{230}\text{Th}_{\text{xs}}$ above the production
252 from dissolved-U decay in the overlying water column points toward a sediment-focusing
253 environment related to ice margin dynamics during the last glacial period. The preserved,
254 decay-corrected, $^{230}\text{Th}_{\text{xs}}$ -normalized ^{10}Be deposition rates (referred to ^{10}Be -fluxes hereinafter)
255 are calculated as follow:

$$256 \quad \text{Flux} [^{10}\text{Be}] = \frac{[^{10}\text{Be}]_{\text{at/g}}^{\text{decay.corrected}} \times Z \times \beta_{230}}{^0A_{\text{Th}-230}^{\text{xs}}} \quad (4)$$

257 where: $[^{10}\text{Be}]^{\text{decay.corrected}}$ is the ^{10}Be concentration at the time of deposition in atoms per gram
258 of sediment; Z is the water depth (2063 m); β_{230} is the ^{230}Th production rate from the seawater
259 ^{234}U decays throughout the water column ($2.67 \cdot 10^{-2}$ dpm.m $^{-3}$.y $^{-1}$; François *et al.*, 2004) and

260 $^0A_{\text{Th}-230}^{\text{xs}} = A_{\text{Th}-230}^{\text{xs}} \times e^{\left(\frac{\ln(2)}{T_{1/2}} \times t\right)}$ is the scavenged $^{230}\text{Th}_{\text{xs}}$ concentration at the time of deposition.

261 The uncertainties (2σ) of the calculated ^{10}Be -fluxes are derived from the propagation of both
262 uncertainties and vary between 1.6 and 25.1% (average value: 5.0%). Note that solely 2
263 samples have uncertainties above 10% representing probable measurement outliers. The water
264 depth was considered constant in first approximation for the calculation. The Th analyses
265 have been performed after total digestion of 1 g of sediment, while Be isotopes were extracted
266 from 1 g of sediment after partial leaching (in order to avoid the extraction of matricial Be;
267 Bourlès *et al.*, 1989). Given the difference of these two approaches, the ^{10}Be -fluxes calculated
268 here thus represent minimal values. This should be considered when comparing our results

269 against reference values, but it does not prevent any qualitative interpretations. A constant
270 leaching efficiency of ~60 % (Bourlès *et al.*, 1989) has been verified based on the results from
271 Ménabréaz *et al.* (2011) and Ménabréaz (2012) and can be use for calibration.

272 **4. Results and discussion**

273 Sample concentrations, ratios and fluxes are listed in Table 1 and presented vs. depth along
274 the core photos, CT-Scan images and description in Figures 3 and 5. Sample ratios and fluxes
275 are presented vs. age in Figures 7 and 8. Note that all results and statistical averages are
276 reported hereafter with a $\pm 2\sigma$ uncertainty.

277 4.1. Authigenic ^{10}Be and ^9Be concentrations

278 The authigenic ^{10}Be (decay-corrected) concentrations vary from 0.051 ± 0.002 to 6.403 ± 0.085
279 $\times 10^8 \text{ at.g}^{-1}$ (mean: $1.64 \times 10^8 \text{ at.g}^{-1}$; $\sigma=1.74$). Such broad variability (> 2 orders of magnitude)
280 is unusual in marine records where ^{10}Be concentrations usually vary by factors of 2 or 3.
281 Comparable large amplitude variation was only found in few polar cores from the Lomonosov
282 Ridge (central Arctic), Fram Strait and the Ross Sea (Antarctica) where ^{10}Be concentration
283 ranging from 0.2 to $19.5 \times 10^8 \text{ at.g}^{-1}$ (Eisenhauer *et al.*, 1994; Spielhagen *et al.*, 1997, 2004;
284 Aldahan *et al.*, 1997; Sjunneskog *et al.*, 2007) were interpreted as paleoclimatic signals with
285 high (resp. low) ^{10}Be concentrations during Interglacial or major Interstadial (resp. Glacial)
286 periods. Given the temporal resolution of core PC16, we can argue that in Baffin Bay the ^{10}Be
287 concentration varies on shorter time scales and is not directly related to Glacial/Interglacial
288 cycles contrary to the mechanism proposed for the Arctic Ocean. The mean ^{10}Be
289 concentrations in Baffin Bay are slightly lower than values from the central Arctic and
290 Arctic/Nordic Seas of $3\text{-}4 \times 10^8 \text{ at.g}^{-1}$ and $6.7 \times 10^8 \text{ at.g}^{-1}$, respectively (Frank *et al.*, 2008;
291 Eisenhauer *et al.*, 1994; Aldahan *et al.*, 1997), as well as with values from lower latitude sites
292 such as the Portuguese margin and the Gulf of Papua: $\sim 4.4 \times 10^8 \text{ at.g}^{-1}$ and $\sim 5.6 \times 10^8 \text{ at.g}^{-1}$,

293 respectively (Carcaillet *et al.*, 2004b; Ménabréaz *et al.*, 2011, 2014). A strict interpretation of
294 the ^{10}Be concentration in term of atmospheric fluxes between distinct sites must be avoided
295 because of environmental parameters. However, the overall low ^{10}Be concentration from
296 Baffin Bay compared to low-latitude and Arctic sites suggests relatively reduced ^{10}Be inputs
297 to the coring site (especially within the very low ^{10}Be concentration interval at 232-275 cm
298 depth).

299 The authigenic ^9Be concentrations vary from 0.621 ± 0.004 to $3.269\pm 0.065 \times 10^{16}$ at.g $^{-1}$ (mean:
300 1.33×10^{16} at.g $^{-1}$; $\sigma=0.56$). The variability of ^9Be concentration is about 25 times lower than the
301 ^{10}Be variation. Still, this range of variation (~5 times) is larger than the variability observed in
302 lower-latitude marine cores (~2 times), advocating for important ^9Be transport and deposition
303 changes along core PC16. The mean ^9Be concentration values from PC16 are lower than
304 those from the Portuguese margin and the Golf of Papua, $2-4 \times 10^{16}$ at.g $^{-1}$, but slightly higher
305 than values from the central Arctic ($0.6-1 \times 10^{16}$ at.g $^{-1}$) and from ODP983 site ($\sim 0.25 \times 10^{16}$
306 at.g $^{-1}$; Knudsen *et al.*, 2008). Notwithstanding specific interpretations related to each sites,
307 this comparison reveals the strong association between ^9Be concentration and terrigenous
308 inputs (*e.g.*, Bourlès *et al.*, 1989; Brown *et al.*, 1992). We can argue that the central and
309 deepest part of Baffin Bay is a sediment-focusing area that received slightly more ^9Be inputs
310 than deep open ocean basins because of its proximity to continental margins and the reduced
311 size and shape of the bay. This is in accordance with the U-Th results from PC16 (Nuttin and
312 Hillaire-Marcel, 2015).

313 The fluctuations of ^{10}Be and ^9Be concentrations present a high correlation coefficient ($r=0.88$,
314 Table 2), suggesting that sources of ^{10}Be and ^9Be into the bay (Figure 3) are both located on
315 the boarding continents and that the atmospheric ^{10}Be component directly precipitated over
316 the bay likely represents minor amounts compared to these large inputs of continental origin.

317 4.2. Authigenic $^{10}\text{Be}/^9\text{Be}$ ratios

318 The authigenic $^{10}\text{Be}/^9\text{Be}$ ratios ranging from 0.043 ± 0.003 to $2.624\pm 0.123 \times 10^{-8}$ (mean: $1.01 \times$
319 10^{-8} ; $\sigma=0.86$) disclose a variability over almost 2 orders of magnitude very similar to those of
320 ^{10}Be and ^9Be concentrations. The fact that authigenic $^{10}\text{Be}/^9\text{Be}$ ratio values are broadly
321 coherent with $^{10}\text{Be}/^9\text{Be}$ ratio values ranging from 0.36 to 1.54×10^{-8} measured in Arctic rivers
322 (Franck *et al.*, 2009) supports the continental origin of the dissolved beryllium in Baffin Bay.
323 Measurements from the Baffin Bay watersheds are however required to further discuss this
324 issue. The authigenic $^{10}\text{Be}/^9\text{Be}$ ratio population can be divided in two groups: values (1) lower
325 than 10^{-8} and (2) higher than 10^{-8} (Table 1, Figures 3, 5 and 6). The first group has ^{10}Be
326 concentrations and $^{10}\text{Be}/^9\text{Be}$ ratios similar to those found within poorly sorted diamicton
327 layers of the Ross Ice Shelf (Sjunneskig *et al.*, 2007) and of Arctic sediments of the last 350
328 kyr (*e.g.*, Aldahan *et al.*, 1997). They are thus coherent with other records from Polar
329 Regions. The second group presents higher values in range with the lower limits of authigenic
330 $^{10}\text{Be}/^9\text{Be}$ ratios from mid-to-low-latitude cores with similar sedimentation rates (Carcaillet *et*
331 *al.*, 2004b; Ménabréaz *et al.*, 2014). This nearly bi-modal distribution of the authigenic
332 beryllium concentrations and ratios presents a strong affinity with lithofacies changes
333 (Figures 3 and 5).

334 4.3. Beryllium and sedimentological features

335 From the authigenic Be result variations presented above, questions arise about the nature of
336 the forcing parameters. In order to understand the links with lithofacies changes, core PC16
337 offers a unique opportunity because of the numerous multi-proxy results available. The
338 detailed sedimentological features of core PC16 have been largely presented and discussed
339 elsewhere (see Simon *et al.*, 2012, 2014, in prep.; Simon, 2013; Nuttin and Hillaire-Marcel,
340 2015 for details) and are beyond the scope of this paper. Sediment compositional variability
341 in core PC16 is a combination of changes in sediment delivery, transport and provenance

342 directly related to ice margin dynamics (Simon *et al.*, 2014). This variability is illustrated here
343 by grain-size (bulk and magnetic), density (CT number), XRF Ca/Fe and XRD Carbonate
344 percents along with the images and CT-Scans of core PC16 (Figures 3 and 8) and summarized
345 by a principal component analysis (PCA; Table S1, Figure 4) that disentangle the main
346 compositional (mineralogy and grain-size) changes. The first principal component (PC1)
347 accounts for 61% of the total variance and has positive loadings with proxies of coarse detrital
348 carbonate layers such as dolomite and calcite (XRD), XRF Ca, density and >63 μm %; and
349 negative loadings with proxies related to finer sediments, feldspars (XRD) and XRF Fe and Ti
350 (Figures 3 and 4; Table 2). It highlights the two main sedimentation features from Baffin Bay:
351 *i.e.*, coarse-grained carbonate-rich sediments transported by icebergs and sea ice originating
352 from the northern end of the bay against fine-feldspar rich sediments originating from
353 Greenland and Baffin Island (Simon *et al.*, 2014).

354 Table 2 presents the correlation coefficients between Be isotopes and the sedimentological
355 parameters. As suggested from Figures 3 and 7, the high correlation coefficients between PC1
356 and Be isotopes and ratios (>-0.8) express a strong association with sedimentological
357 parameters. The authigenic beryllium values (^{10}Be , ^9Be and $^{10}\text{Be}/^9\text{Be}$ ratios) generally present
358 significant increases at levels corresponding to fine grained feldspar-rich intervals associated
359 with increases of clay minerals (30-50%; Simon *et al.*, 2014) while coarse carbonate-rich
360 sediments (30-40% dolomite and 10-15% calcite) are characterized by lower Be values
361 (Figure 3). This pattern is consistent with the scavenging efficiency of dissolved Be that
362 depends on the composition and size of the particles available in the water column. The
363 authigenic beryllium (^{10}Be , ^9Be and $^{10}\text{Be}/^9\text{Be}$ ratios) and thorium-excess ($^{230}\text{Th}_{\text{xs}}$) distribution
364 are strongly grain size dependent, with significant positive correlations with very fine to fine
365 silts (2-8 μm) while clays (0-2 μm) and medium to very coarse silts (8-63 μm) does not
366 present any significant correlations (Table 2). The high correlation of authigenic Be isotopes

367 with the 2-8 μm fraction rather than with the clay-sized material is intriguing because we
368 expected larger association with the smaller particles (because of their higher specific surface
369 area available). We tentatively explain these results by the cohesive behavior of the clay-sized
370 material that tends to form aggregate. These aggregates would have faster sinking rates and
371 lower specific surface area available for the adsorption of dissolve Be explaining higher
372 scavenging rates associated with very fine to fine silts. To the opposite, the authigenic Be
373 concentrations and ratios are significantly anti-correlated with the coarser intervals ($>63 \mu\text{m}$,
374 Figure 3) related to iceberg transported sediments (*i.e.*, BBDC). Despite an overall similar
375 behavior of both isotopes in respect to sedimentary parameters, the authigenic ^{10}Be
376 concentrations and authigenic $^{10}\text{Be}/^9\text{Be}$ ratios are usually slightly more correlated with
377 sedimentological parameters than ^9Be isotopes. We interpret this difference by distinct
378 sources and transport processes between both isotopes. The continental inherited- ^{10}Be (*i.e.*,
379 atmospheric ^{10}Be deposited onto ice-sheets) are derived from the melting and calving of ice
380 sheets and are released in the water column during the ice melting, while the ^9Be isotopes are
381 mainly coming from the mechanical erosion of bedrocks at the basal interface of ice streams.
382 It is likely that a fraction of the dissolved ^9Be isotopes was scavenged rapidly on the
383 continental margins within the buoyant turbid meltwater plumes or nepheloid layers (*i.e.*, re-
384 suspension of fine sediment particles by bottom currents) due to higher particles
385 concentration (Bacon and Rutgers van der Loeff, 1989) while the ^{10}Be transported within
386 icebergs or sea ice was exported farther from the sources. A fraction of ^9Be would thus be
387 immobilized in glacial/river estuaries (Kusakabe *et al.*, 1991) and removed from the water
388 column within the less saline meltwater plumes on the inner-shelf (Frank *et al.*, 2009; Ó
389 Cofaigh *et al.*, 2013) rather than being transported to the centre of the bay. Therefore, despite
390 complex relationships between granulometry and mineralogy of these sediments, we can claim

391 that the authigenic ^{10}Be signatures in Baffin Bay reveal local marine and glacial influences
392 from the surrounding continents instead of reflecting a global ^{10}Be -production signal.

393 4.4. ^{10}Be -fluxes ($^{230}\text{Th}_{\text{xs}}$ -normalized)

394 The preserved vertical ^{10}Be deposition rates (^{10}Be -fluxes) varies from 0.717 ± 0.046 to
395 $25.388 \pm 0.973 \times 10^8 \text{ at.cm}^{-2}.\text{ka}^{-1}$ with a mean value of $6.94 \times 10^8 \text{ at.cm}^{-2}.\text{ka}^{-1}$ and a standard
396 deviation of 7.037 expressing a large variability (Table 1, Figure 5). When increased by a
397 factor of 40%, to encompass the methodological bias induced by the partial leaching of Be
398 isotopes vs. the total digestion of Th (see section 3.5), the ^{10}Be -fluxes are oscillating between
399 1.004 ± 0.065 to $35.544 \pm 1.362 \times 10^8 \text{ at.cm}^{-2}.\text{ka}^{-1}$ with an average of $9.72 \times 10^8 \text{ at.cm}^{-2}.\text{ka}^{-1}$
400 (Table 3). The Holocene average ^{10}Be -flux is estimated at $7.6 \pm 0.2 \times 10^8 \text{ at.cm}^{-2}.\text{ka}^{-1}$ (10.7 ± 0.2
401 $\times 10^8 \text{ at.cm}^{-2}.\text{ka}^{-1}$ value corrected), while during the glacial period the fluxes oscillated around
402 14.3 ± 0.5 (20.1 ± 0.7) and 2.0 ± 0.2 (2.8 ± 0.2) $\times 10^8 \text{ at.cm}^{-2}.\text{ka}^{-1}$ within the carbonate-free and
403 carbonate-rich layers, respectively (carbonate layers are highlighted by white banding in
404 Figures 3 and 5). The Holocene ^{10}Be -fluxes are similar with the actual ^{10}Be production of
405 $12.1 \pm 0.3 \times 10^8 \text{ at.cm}^{-2}.\text{ka}^{-1}$ (Monaghan *et al.*, 1986). During the glacial period, results of ^{10}Be -
406 fluxes ranging from 0.7 to $25.4 \times 10^8 \text{ at.cm}^{-2}.\text{ka}^{-1}$ are coherent with the bulk ^{10}Be -fluxes
407 calculated in the Arctic and Fram Strait that range from 2×10^8 to $33 \times 10^8 \text{ at.cm}^{-2}.\text{ka}^{-1}$
408 (Eisenhauer *et al.*; 1994; Aldahan *et al.*, 1997). Despite an obvious sedimentological
409 relationship in core PC16, and at the exception of few intervals, the ^{10}Be -fluxes are
410 compatible with the modeled ^{10}Be -production range: ~ 3 to $27 \times 10^8 \text{ at.cm}^{-2}.\text{ka}^{-1}$; with the
411 ^{10}Be -fluxes from ice and marine records which range from ~ 1.2 to $70 \times 10^8 \text{ at.cm}^{-2}.\text{ka}^{-1}$, and
412 with the globally integrated and long-term averaged ^{10}Be -fluxes into marine sediments
413 varying between 9 to $28 \times 10^8 \text{ at.cm}^{-2}.\text{ka}^{-1}$ (see Table 3 for details and references). The lowest
414 ^{10}Be -fluxes that characterized the BBDC layers are also rather similar to those found within
415 the Ice Summit record of GISP2: 2 to $6 \times 10^8 \text{ at.cm}^{-2}.\text{ka}^{-1}$ (Muscheler *et al.*, 2004). It is

416 puzzling to obtain ^{10}Be -fluxes similar to ice core records in core PC16 given the strong
417 lithogenic imprint. This similarity might be explained by a smoothing of the ^{10}Be cosmogenic
418 nuclide signal related to the long-lasting deposition/drift onto the regional ice-sheets and by
419 the thick sea ice and/or ice-shelf cover during glacial periods. The relatively low ^{10}Be -fluxes
420 calculated in Baffin Bay would thus likely represent a ^{10}Be production signal buffered by the
421 glacial factors controlling the ^{10}Be inputs and transport within the water column. Furthermore,
422 these low ^{10}Be -fluxes from Baffin Bay are also coherent with the low depositions of ^{10}Be in
423 Greenland and in the Arctic Ocean (Eisenhauer *et al.*, 1994; Spielhagen *et al.*, 1997) likely
424 due to atmospheric circulation patterns and the distribution of precipitation in the northern
425 hemisphere (Heikkila *et al.*, 2013; Frank *et al.*, 2009).

426 4.5. Testing the $^{10}\text{Be}/^9\text{Be}$ ratios and $^{230}\text{Th}_{\text{xs}}$ -normalized ^{10}Be fluxes methods

427 The two existing normalization methods were independently used in former studies, but only
428 two studies applied the two methods on the same sedimentary core. The first study by
429 Knudsen *et al.* (2008) obtained large uncertainties about $^{230}\text{Th}_{\text{xs}}$ values precluding their use
430 for reliable normalization. The second study by Ménabréaz *et al.* (2011) demonstrated close
431 agreement of the results obtained by the two methods based on a small number of samples. In
432 PC16, the ^{10}Be -flux variability calculated using the $^{230}\text{Th}_{\text{xs}}$ method is similar to that of the
433 authigenic $^{10}\text{Be}/^9\text{Be}$ ratios measured at the same depths (Figure 5). The high correlation
434 coefficient ($r=0.81$) and coherence between the two signals provide strong evidence that the
435 two normalization processes yield equivalent results in a very contrasted environment where
436 large compositional variations prevailed (Figures 5 and 6, Table 2). It somehow demonstrates
437 that normalizing authigenic ^{10}Be concentrations by authigenic ^9Be concentrations permits an
438 accurate correction for the total particle flux variation. Both normalization processes present
439 large variations directly related to lithofacies changes (Figure 3). The normalizers (*i.e.*, $^{230}\text{Th}_{\text{xs}}$
440 and ^9Be) are highly correlated (Table 2) which might seem surprising given the distinct

441 affinities (depending on particle composition) and the different scavenging residence times of
442 both elements in the open ocean (10-50 years for Th against 500-1000 years for Be, Chase *et*
443 *al.*, 2002). The large domination of glaciomarine lithogenic particles presenting strong surface
444 reactivity with both nuclides probably explained the first assertion (Roy-Barman *et al.*, 2005,
445 2009). Our results also suggest quicker adsorption rates and thus shorter scavenging residence
446 time for dissolved Be within the Baffin Bay water column probably related to the high
447 concentration of lithogenic particles. This is coherent with previous results from the Arctic,
448 North Atlantic basin and circum-Antarctica where lower scavenging residence time for Be
449 isotopes of about 80-200 years associated with high particles concentration and oceanic
450 mixing have been proposed (von Blanckenburg *et al.*, 1996, 1999; von Blanckenburg and
451 Bouchez, 2014; Frank *et al.*, 2009). An oceanic circulation/mixing influence over the
452 authigenic Be signal into Baffin Bay during the glacial period is also supported by higher
453 correlation coefficients between both normalization methods when considering the rare
454 episodes of Arctic Waters overflow through the Davis Strait (illustrated by blue dots in
455 Figures 5 and 6). These episodes have been demonstrated within core PC16 by the occurrence
456 of short periods of $^{230}\text{Th}_{\text{xs}}$ export ($^{230}\text{Th}_{\text{xs}}$ -losses, Nuttin and Hillaire-Marcel, 2015).

457 4.6. Comparison with relative paleointensity

458 In core PC16, the comparison of ^{10}Be -proxies with the RPI record does not demonstrate any
459 clear relationship (Figure 7). No systematic increases of authigenic $^{10}\text{Be}/^9\text{Be}$ ratios and ^{10}Be -
460 fluxes are observed during the large RPI lows corresponding to the Laschamp, Norwegian-
461 Greenland Sea and post-Blake/Blake events (Figure 7). The authigenic $^{10}\text{Be}/^9\text{Be}$ ratios and
462 ^{10}Be -fluxes do not present any significant correlation coefficients with the RPI record (Figure
463 7, Table 2). This absence of correlation with the RPI record clearly supports the dominance of
464 the environmental imprints –as discussed above- on the ^{10}Be deposition in Baffin Bay. The
465 high correlation ($r=0.83$) between the RPI normalizers (such as the ARM) and the authigenic

466 $^{10}\text{Be}/^9\text{Be}$ ratios also indicates a strong environmental imprint related to specific grain-size
467 ranges that insure maximum Be adsorption rates (namely the very fine to fine silt range is
468 highly correlated with the ARM: $r=0.7$). The strong correlation of the ^{10}Be -proxies with
469 environmental parameters suggests a local signature related with the succession of events
470 driven by paleoclimatic controls precluding any interpretation of the ^{10}Be -proxies in term of
471 geomagnetic variability. On the opposite, our findings provide strong evidence supporting the
472 use of the cosmogenic nuclide ^{10}Be as a stratigraphic marker in the Arctic and sub-Arctic
473 regions as suggested by several authors (Eisenhauer *et al.*, 1994; Spielhagen *et al.*, 1997;
474 Aldahan *et al.*, 1997; Frank *et al.*, 2008).

475 **5. Paleoenvironmental implications**

476 The quasi bi-modal distribution of Be results in core PC16 corresponding with the main
477 Baffin Bay sedimentary features (Figure 3) and the temporal resolution of Be variations
478 suggest glacial dynamics as the main forcing parameter to explain Be inputs and transport
479 changes. The most striking features of this glacial dynamic around Baffin Bay is the
480 variability of the GIS limits over the Greenland continental shelf and the ice streaming events
481 from the LIS and IIS together with sea ice/ice-shelf cover variability (Figure 9). The origin,
482 timing and limits of these large glacial variations are still very poorly constrained. However,
483 recent studies have demonstrated that ice-sheets advanced well onto the continental margins
484 all over the bay and as far as the shelf edges during the LGM (Li *et al.*, 2011; Ó Cofaigh *et*
485 *al.*, 2012, 2013; Hogan *et al.*, 2012; Dowdeswell *et al.*, 2013; Simon *et al.*, 2014; Margold *et*
486 *al.*, 2015).

487 The cumulative inventory of measured $^{230}\text{Th}_{\text{xs}}$ in core PC16 (see Nuttin and Hillaire-Marcel,
488 2015 for details), recalculated in respect to the revised age model, clearly exhibits maximal
489 sediment fluxes corresponding to minimum relative sea level (RSL) during the LGM (Figure
490 8). High Be concentrations (Figures 7 and 8) and *ca.* 100% sediments originating from

491 Greenland (Simon *et al.*, 2014) also characterized this period. The other periods of high
492 authigenic Be concentrations are also associated with finer sediments originating mainly from
493 Greenland (Figure 8). This facies is related to the resuspension of fine lithogenic sediments
494 (glacial flour) associated to nepheloid layers during period of intense ice margin advances, to
495 outflow of dense winter water from the continental shelves, or to meltwater sediment-laden
496 plumes during period of ice margin retreats. Both scenarios imply ice margins/ice streams
497 extending over the Greenland continental shelf. Together with the inherent higher scavenging
498 efficiency of smaller particles, the increase in particle concentration due to the proximal ice
499 margin likely contributed to the increase of the Be-scavenging rates in central Baffin Bay
500 (Figure 9). We can also assume that a fraction of the Be transported onto ice floes sediments
501 might be transferred into the water column by wave wash-off or turning of floes during these
502 glacial maxima. Two ^{10}Be concentration values of 2.2 ± 0.1 and $2.8\pm 0.1 \times 10^8 \text{ at.g}^{-1}$ measured
503 in clay samples from Arctic ice floes (Eisenhauer *et al.*, 1994) and corresponding to the
504 higher ^{10}Be concentration values from PC16 support this assumption, although samples from
505 Baffin Bay are requested to further discuss this assertion.

506 On the other hand, the low authigenic Be concentrations and ratios within the BBDC layers
507 support a Be dilution in these IRD layers (Figure 2). Such a dilution is possibly associated to
508 (1) increases of terrigenous sediment inputs, (2) changes of sediment composition/grain-size
509 affecting the scavenging efficiency of dissolved beryllium, (3) reduced net Be inputs into the
510 centre of the bay and/or (4) higher exports of Be by extensive sea ice/iceberg drift episodes.
511 The absence of significant increases in sedimentation rates (Figure 8) and/or $^{230}\text{Th}_{\text{xs}}$ changes
512 (Nuttin and Hillaire-Marcel, 2015) during these intervals favor the last three hypotheses.
513 However one cannot totally exclude that short episodes of increased inputs of terrigenous
514 sediments could remain uncaptured by the age model resolution. During the BBDC intervals,
515 the limited amount of sediment originating from Greenland supports a distant Greenland ice

516 margin limit while coarse sediments originating from Baffin Island are explained by high
517 calving rates from Laurentide ice streams (Simon *et al.*, 2014). According to this model, the
518 BBDC intervals occur during periods of retreated ice limits over the inner continental shelves
519 (Figure 9). The cause of ice margin instabilities around Baffin Bay is still not well understood
520 but may be related to higher summer insolation and/or to the advection of warm Atlantic
521 Water. Indeed the phasing between magnetic grain size in core PC16 (SIRM/ k_{LF} ratios in
522 Figure 8, Simon *et al.*, 2012, 2014) and the insolation variation supports a climatic control on
523 ice margin dynamics and iceberg drifting in the bay. The advection of an intermediate warm
524 water mass in Baffin Bay probably contributed also to destabilize the ice margins (Holland *et*
525 *al.*, 2008; Jennings *et al.*, 2014) increasing meltwater delivery along the Baffin Bay shelves
526 and slopes. It resulted in increased stratification of the water column and therefore to longer
527 residence times of Be within the deep water masses (von Blanckenburg and O'Nions, 1999).
528 Such an oceanic pattern would reduce the Be adsorption and deposition rates explaining
529 possibly part of the low ^{10}Be concentration within the BBDC intervals. Moreover, and despite
530 complex interactions of eustatic and isostatic parameters over the relative sea level (RSL)
531 around Baffin Bay, we can reasonably assess that the sea level rose over continental shelves
532 during ice margin retreat periods (Long *et al.*, 2008; Simpson *et al.*, 2009; Funder *et al.*,
533 2011). Throughout these periods of marine transgression, oceanic circulation and boundary
534 scavenging (changes in nature and intensity, Lao *et al.* 1992) may have become significant
535 processes involving large transfer of dissolved Be from the centre of the bay toward the
536 margins (Roy-Barman *et al.*, 2009). The low authigenic $^{10}\text{Be}/^9\text{Be}$ ratios and ^{10}Be -fluxes
537 during the BBDC intervals within core PC16 might thus be partly explained by high boundary
538 scavenging rates over the large Greenland continental shelf together with important
539 iceberg/sea ice drifts due to very active Laurentide ice streams. For instance, the very low
540 authigenic $^{10}\text{Be}/^9\text{Be}$ ratios found during the late MIS-3 (*ca.* 40-28.5 ka BP) is explained by

541 retreated GIS limits and numerous ice streaming episodes from the LIS and IIS possibly
542 related to the Dansgaard-Oeschger events between the large glacial surges of Heinrich-events
543 3 and 4 (Figure 8; Simon et al., 2014).
544 Even though interpretations resulting from the authigenic Be signature in Baffin Bay can be
545 discussed and are likely the results of several processes, we can reasonably proposed that the
546 glacial dynamic of regional ice sheets (*i.e.*, GIS, IIS and LIS) is the main internal driving
547 mechanism in term of ^{10}Be inputs and delivery into the bay. Moreover, ice-sheets topography
548 changes related to such glacial dynamic also imply a reorganization of the atmospheric
549 circulation (*e.g.*, Steffensen *et al.*, 2008) modifying the stratosphere/troposphere exchanges,
550 the wet/dry deposition ratio and the dust inputs into the ocean (Werner *et al.*, 2002). Periods
551 of ice-sheet growth (resp. decay) characterized by higher (resp. lower) wet deposition would
552 then favor higher (resp. lower) ^{10}Be deposition rates within the bay. Although our results are
553 coherent with these views, atmospheric modeling of ^{10}Be deposition considering regional ice-
554 sheet topography changes are needed to verify and quantify these assumptions.

555 **6. Conclusions**

556 The authigenic ^{10}Be cosmogenic nuclide and ^9Be stable isotope data were measured along a
557 7.41 m sedimentary core of the sub-arctic basin Baffin Bay in order to reconstruct the
558 geomagnetic dipole moment variations using the $^{10}\text{Be}/^9\text{Be}$ ratios and ^{10}Be -fluxes ($^{230}\text{Th}_{\text{xs}}$ -
559 normalized). The results of the two normalizations are coherent and reveal that environmental
560 processes such as glacial dynamics and oceanic variability directly control the variations of
561 the authigenic ^{10}Be and ^9Be concentrations in Baffin Bay sediments. The contribution of the
562 atmospheric ^{10}Be cosmogenic nuclide production modulated by the geomagnetic field
563 intensity remains hidden behind this environmental signal. Accordingly, in such conditions
564 ^{10}Be production proxies do not allow to characterize the geomagnetic features such as dipole
565 lows linked to excursions or reversals. Beryllium isotopes are preferentially adsorbed on fine

566 silicate particles associated with sediment plumes originating from lateral ice margin
567 advances. On the contrary they have little affinity with coarse-grained and carbonate-rich
568 sediments (*i.e.*, BBDC) associated with iceberg and sea-ice transport originated from the
569 North-Eastern Laurentide Ice Sheet and Innuitian Ice Sheet ice streaming events. During
570 these BBDC episodes, glacial margin retreats together with marine transgression over the
571 Greenland continental shelf likely contributed to increase boundary scavenging involving
572 large transfers of dissolved beryllium from the centre of the bay toward the margins. Our
573 findings provide strong evidences that support the use of cosmogenic nuclide ^{10}Be as a
574 stratigraphic marker in the Arctic and sub-Arctic regions. Yet, our results caution a
575 straightforward use of ^{10}Be -concentrations as a proxy of Interglacial/Glacial cycles or major
576 Interstadial periods (as suggested by several authors), and rather propose to relate the ^{10}Be
577 variations to higher-frequency paleoclimatic changes and glacial dynamics. Therefore,
578 studying the cosmogenic ^{10}Be and stable ^9Be isotopes in combination with ancillary
579 sedimentological parameters in arctic and sub-arctic marine sediments provide valuable
580 climatic information, but it must be acknowledged that the cost/information ratio is probably
581 too large to systematically use atmospheric ^{10}Be cosmogenic nuclides and ^9Be isotopes as
582 paleoclimatic proxies.

583 **Acknowledgements**

584 We acknowledge Valery Guillou (CEREGE) for his help during chemical sample preparation
585 and ^9Be measurements at the AAS. M. Arnold, G. Aumaître and K. Keddadouche are thanked
586 for their valuable assistance during the measurements performed at the ASTER AMS national
587 facility (CEREGE, Aix en Provence). This equipment is supported by the INSU/CNRS, the
588 IRD and the CEAEA and by the ANR through the program “EQUIPEX Investissement
589 d’Avenir”. This study is a contribution of the project MAGORB ANR 09 BLAN 0053
590 (CEREGE, IGP, LSCE). G. St-Onge and C. Hillaire-Marcel acknowledge financial support

591 from the Natural Science and Engineering Council of Canada (NSERC). CEREGE belongs to
592 Observatoire des Sciences de l'Univers OSU-Institut Pythéas.

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998

999 **Table 1.** AMS measurements, authigenic ^{10}Be and ^9Be concentrations, authigenic $^{10}\text{Be}/^9\text{Be}$
1000 ratios and calculated ^{10}Be -fluxes of core HU2008-029-016PC samples.

1001 **Table 2.** Correlation coefficients of Be isotopes (concentration, ratio and fluxes) and
1002 sedimentological parameters.

1003 **Table 3.** ^{10}Be -production/flux in literature (in $\text{atoms}\cdot\text{cm}^{-2}\cdot\text{kyr}^{-1}$)
1004

1005 **Figure 1.** General bathymetry, simplified oceanic circulation, sketch of the Paleozoic
1006 outcrops (*MacLean, 1985*) and paleo-Ice-Sheets (including LGM unknown maximum ice
1007 margin extents) of the Baffin Bay region. The location of HU2008-029-016PC sampling site
1008 is indicate by a red dots. Red arrows illustrate Atlantic “warm” waters, whereas the blue
1009 arrows represent colder Arctic waters. The simplified representations of the Greenland
1010 (green), Innuitian (blue) and Laurentide ice sheet (red) limits and major ice stream locations
1011 during the LGM (colored areas) are adapted from *Funder et al. (2011)*, *Dyke (2004)* and
1012 *England et al. (2006)*.

1013 **Figure 2.** Relative paleointensity (NRM/ARM_{25-35mT}) tuned on RPI record of ODP Site 1063
1014 (*Channell et al., 2012*) and chronostratigraphy of PC16. The RPI tie-points used in this study
1015 are represented by green dots while the red diamonds and yellow triangle are radiocarbon
1016 ages and calcite tie-points respectively (*Simon et al., 2012*). The incertitude in the age model
1017 is represented by shaded area.

1018 **Figure 3.** Beryllium isotopes results vs. high-resolution physical, geochemical and
1019 mineralogical results from core PC16. Log: general simplified stratigraphy of the core; CT:
1020 CAT-Scan image of the core; HRI: high-resolution digital image; CT Number (density
1021 proxy); XRD carbonates (dolomite and calcite) cumulative percents; log(Ca/Fe): μXRF
1022 element ratio for calcium and iron measured with the ITRAX© core scanner. Grain size (%)
1023 for clay, silt fractions and coarse fractions measured at 4 cm intervals by laser diffraction.
1024 Authigenic ^{10}Be , ^9Be and $^{10}\text{Be}/^9\text{Be}$ ratios are display in log scale. Distinct lithological facies

1025 are highlighted with color banding. Red: uppermost brownish gray silty mud unit (Uppermost
1026 Brownish, “UB”); light green: olive-black silty to clayey mud unit (Olive Clay, “OC”); white:
1027 carbonate-rich yellowish-brown to dark- brown very poorly sorted gravelly sandy mud
1028 detrital layers (DC); dark green: olive gray to dark gray poorly sorted silty to sandy mud low
1029 carbonate detrital layers (LDC) (see text for details).

1030 **Figure 4.** Principal component analysis (PCA) of the mineralogical and grain-size dataset.
1031 The loading scores for PC1 vs. PC2 explain, respectively, 61 and 11.5% of the total variance.
1032 PCA analysis illustrates the two sedimentary modes in Baffin Bay.

1033 **Figure 5.** Authigenic $^{10}\text{Be}/^9\text{Be}$ ratios and ^{10}Be -fluxes on depth. The ^{10}Be -fluxes are compared
1034 to references: (1) global-average ^{10}Be -production values from models, see Table 3 for values
1035 and references (the blue star is the mean value from Masarik and Beer, 2009); (2) global
1036 values measured and averaged by Monaghan *et al.*, 1986; (3) ^{10}Be -flux in GRIP/GISP2 ice
1037 core (Muscheler *et al.*, 2005); (4) ^{10}Be -flux in marine records representing the so-called
1038 “allowed range” from Christl *et al.*, 2007, 2010; (5) ^{10}Be -flux range in deep sea sedimentary
1039 cores from the Arctic Ocean and the Norwegian Sea (Eisenhauer *et al.*, 1994). Red (resp.
1040 blue) dots represent measured $^{230}\text{Th}_{\text{xs}}$ vertical flux higher (resp. lower) than theoretical $^{230}\text{Th}_{\text{xs}}$
1041 vertical flux.

1042 **Figure 6.** Authigenic $^{10}\text{Be}/^9\text{Be}$ ratios vs. ^{10}Be -fluxes ($^{230}\text{Th}_{\text{xs}}$ -normalized). Blue dots represent
1043 measured $^{230}\text{Th}_{\text{xs}}$ -flux samples lower than theoretical flux and illustrate episodes of Arctic
1044 Waters outflow through Davis Strait (Nuttin and Hillaire-Marcel, 2015).

1045 **Figure 7.** Authigenic $^{10}\text{Be}/^9\text{Be}$ ratios, ^{10}Be -fluxes ($^{230}\text{Th}_{\text{xs}}$ -normalized), PC 1 and RPI vs. age.
1046 RPI (Channell *et al.*, 2012) and ^{10}Be -fluxes (Christl *et al.*, 2010) from ODP Site 1063 are
1047 presented along with ^{10}Be -fluxes reference ranges from distinct archives and models (see
1048 Figure 5 and Table 3 for references). Marine isotopic stages 1 to 6 are represented by color
1049 boxes along the age axis. BBDC-layers are represented by a vertical grey banding and
1050 numbered according to Simon *et al.* (in prep.).

1051 **Figure 8.** Paleoclimatic interpretations of authigenic $^{10}\text{Be}/^9\text{Be}$ ratio variations. The NGRIP
1052 $\delta^{18}\text{O}$ ice core record (red curves) and synthetic Greenland record (GLT-syn, blue curve) are
1053 from www.icecores.dk and Barker *et al.* (2011). Marine isotope stages 1 to 6 are represented
1054 with color boxes. BBDC-layers (defined by high XRD carbonates cumulative percents) are
1055 represented by the grey vertical bars and numbered according to Simon *et al.*, (in prep.). The
1056 North Atlantic Heinrich events (Hemming, 2004) are indicated for comparison. The July
1057 insolation is calculated at 70°N (from Laskar *et al.*, 2004). Eustatic relative sea level (RSL) is
1058 from Grant *et al.* (2012). SIRM/ k_{LF} is a magnetic grain size proxy (Simon *et al.*, 2012, in

1059 prep.). The cumulative inventory ratio compares the $^{230}\text{Th}_{\text{xs}}$ accumulation fluxes with the
1060 theoretical vertical production in the overlying water column. Values below 1 during the
1061 Holocene indicate $^{230}\text{Th}_{\text{xs}}$ -losses while values above 1 during the glacial period suggest a
1062 sediment-focusing environment (Nuttin and Hillaire-Marcel, 2015).

1063 **Figure 9.** Simplified Baffin Bay paleogeography during the last glacial cycle. (a) Trans-
1064 Baffin drift characterized by IRD sediments originating from the northern ice streams, and by
1065 meltwater sediment-laden plumes from Baffin Island ice streams. (b) Extended ice margins
1066 characterized by Greenland and Baffin Island glacial flour sediments corresponding to a
1067 permanently ice-covered bay, and by extended ice margin limits over continental shelves.

Table2

Table 2. Correlation coefficients of Be isotopes (concentration, ratio and fluxes) and sedimentological parameters.

Parameters	⁹ Be (at./g)	¹⁰ Be (at./g)	Rapport ¹⁰ Be/ ⁹ Be	²³⁰ (Th _{xs}) ⁰ (dpm.g ⁻¹)	Flux- ¹⁰ Be Th_norm. (atoms.cm ⁻² .kyr ⁻¹)	PC1	PC2
SAR 170714B (cm/kyr)	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.
XRD- Quartz (%)	0.51	0.50	0.49	n.s.	0.49	-0.68	n.s.
XRD- K-Feldspar (%)	0.55	0.69	0.75	n.s.	0.66	-0.77	n.s.
XRD- Plagioclase (%)	0.54	0.72	0.80	n.s.	0.71	-0.84	n.s.
XRD- Calcite (%)	-0.62	-0.72	-0.76	n.s.	-0.66	0.80	n.s.
XRD- Dolomite (%)	-0.51	-0.68	-0.74	n.s.	-0.71	0.89	n.s.
XRF- Ca/Fe	-0.73	-0.80	-0.81	0.55	-0.68	0.95	n.s.
XRF- K/Ti	-0.66	-0.74	-0.77	0.42	-0.65	0.85	n.s.
XRF- Ti/Ca	0.71	0.78	0.79	0.51	0.67	-0.94	n.s.
XRF- Ca	-0.71	-0.78	-0.77	0.54	-0.66	0.92	n.s.
XRF- Fe	0.69	0.77	0.81	0.53	0.66	-0.91	n.s.
GS- 0-2 μm (%)	n.s.	n.s.	0.40	n.s.	0.39	-0.58	0.74
GS- 2-4 μm (%)	0.73	0.77	0.72	0.51	0.61	-0.86	0.47
GS- 4-8 μm (%)	0.79	0.78	0.67	0.62	0.55	-0.82	0.48
GS- 8-63 μm (%)	n.s.	n.s.	n.s.	0.00	n.s.	n.s.	n.s.
GS- >63 μm (%)	-0.70	-0.73	-0.63	0.61	-0.51	0.76	-0.50
Density (g.cm ⁻³)	-0.77	-0.76	-0.73	0.58	-0.66	0.77	n.s.
⁹ Be (at)	1.00	0.88	0.71	0.72	0.60	-0.81	n.s.
¹⁰ Be (at)	0.88	1.00	0.94	0.68	0.73	-0.89	n.s.
Ratio ¹⁰ Be/ ⁹ Be	0.71	0.94	1.00	0.50	0.81	-0.88	n.s.
²³⁰ (Th _{xs}) ⁰ (dpm.g ⁻¹)	0.72	0.68	0.50	1.00	n.s.	-0.54	n.s.
Flux- ¹⁰ Be sed. (atoms.cm ⁻² .kyr ⁻¹)	0.60	0.73	0.81	n.s.	1.00	-0.75	n.s.
PC1 Sedimentological parameter	-0.81	-0.89	-0.88	-0.54	-0.71		
PC2 Sedimentological parameter	n.s.	n.s.	n.s.	n.s.	n.s.		






n.s. are not significant (P < 0.001)




Table 3. ^{10}Be -production/flux in literature (in $\text{atoms.cm}^{-2}.\text{kyr}^{-1}$)

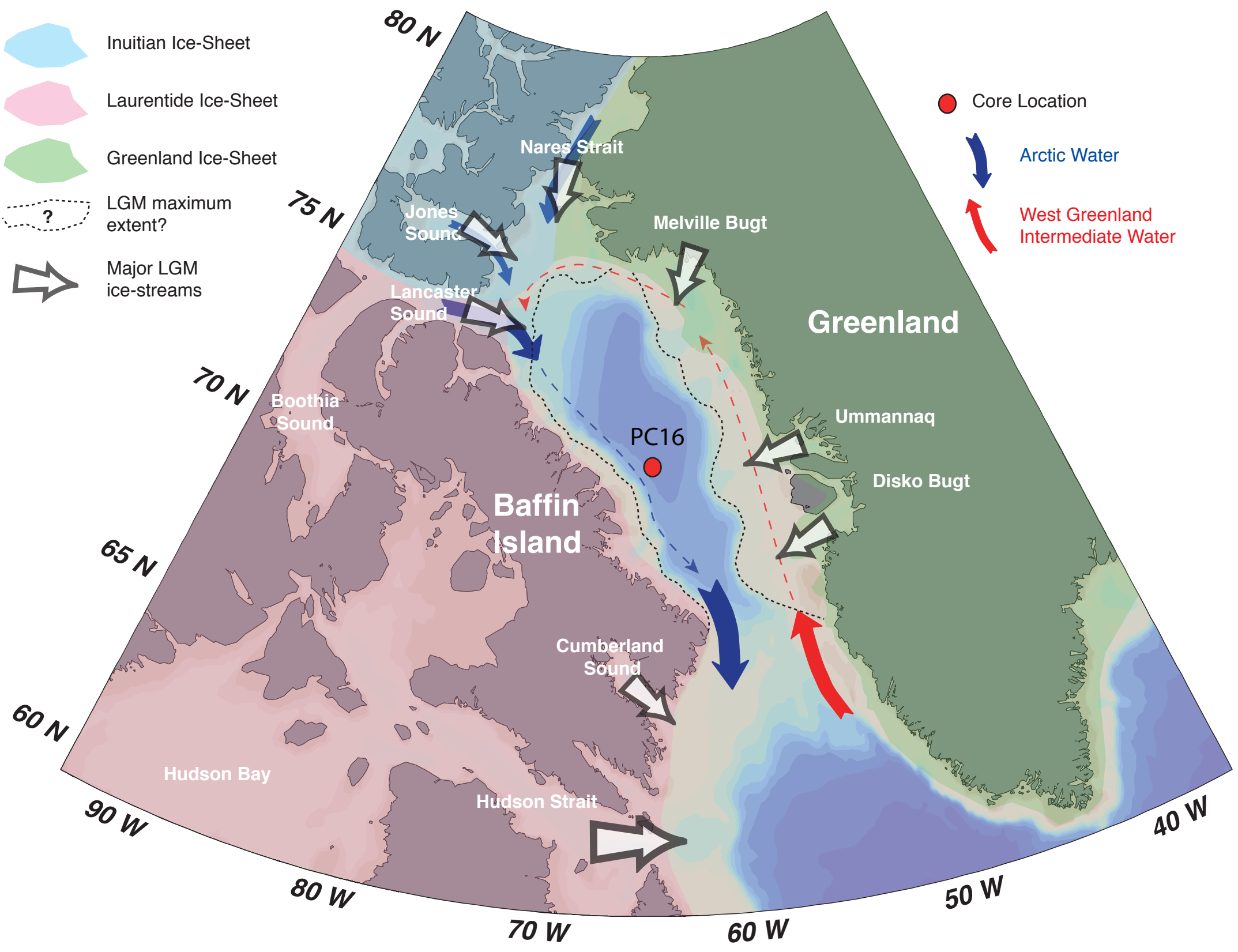
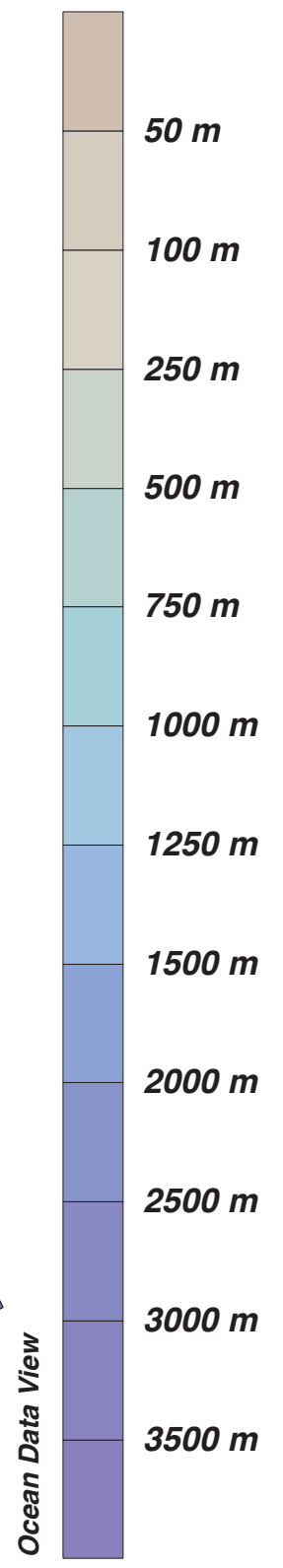
Range (x 10^8)	Mean values (x 10^8)	Notes	References
^{10}Be-production			
5.7 - 10		Global average ^{10}Be -production (model)	<i>Webber et al., 2007; Masarik and Beer, 1999,</i>
12 - 19		60-90°N ^{10}Be -production (model)	<i>2009; Kovaltsov and Usoskin, 2010</i>
	6.6	Global average ^{10}Be -production (model)	<i>Masarik and Beer, 2009</i>
5.2 - 26.4	12.1	Global average ^{10}Be -production (analytical)	<i>Monaghan et al., 1986 (from precipitation</i>
11 - 17	14.0	Long term averaged glob. av. ^{10}Be -prod. (analytical)	<i>measurements during the year 1980)</i>
			<i>Webber and Higbie, 2003; Webber et al., 2007</i>
	6.0	Global average ^{10}Be -production (model)	
	14.2	Global average ^{10}Be -production (empiric)	<i>Lal and Peters, 1967</i>
	9.5	Global average ^{10}Be -production (analytical)	<i>O'Brien, 1979; O'Brien et al., 1991</i>
	12.6	Global average ^{10}Be -production (empiric)	<i>Lal, 1988</i>
^{10}Be-flux in ice-core records			
1.8 - 8.6	3.4	Greenland Summit (GRIP/GISP2)	<i>Muscheler et al., 2005; Finkel and Nishiizumi, 199.</i>
	3.3	Dye 3 (Greenland)	<i>Beer et al., 1991</i>
1.5 - 12.8		Renland Ice Core over 1931-1987 (Greenland)	<i>Aldahan et al., 1998</i>
1.4 - 10		Dronning Maud Land ice core over 1932-1988 (Antarctic)	
	4.0±0.3	EPICA DML (Neumayer surface snow, Antarctic)	<i>Auer et al., 2009</i>
	1.9±0.1	EPICA DML (Kohnen deep ice core, Antarctic)	
	1.4±0.2	Dome C (surface firn and firn core 0-12 m; Antarctica)	
1.2 - 2.5		Epica DC between 320-340 kyr	<i>Cauquoin et al., 2014</i>
0.6 - 3.3	1.6	Epica DC between 200-800 kyr	<i>Cauquoin, 2013</i>
	2.5	Vostok, South Pole (Antarctic)	<i>Raisbeck and Yiou, 1985</i>
	1.4	Taylor Dome (Antarctic)	<i>Steig et al., 1996</i>
	1.7 and 1.8	Concordia and Vostok over the last 60 years (Antarctic)	<i>Baroni et al., 2011</i>
	4.8±1.6	Law Dome snow pit over the year 2001 (Antarctic)	<i>Pedro et al., 2006</i>
2.0 - 4.5		Dome Fuji ice core over 700-1900 yr CE (Antarctica)	<i>Horiuchi et al., 2008</i>
	2.2 (2.3)	GRIP snow pit (Greenland) ^{10}Be deposition flux (1986-1998)	<i>Heikkilä et al., 2008</i>
	2.8	ECHAM5-HAM general circulation model ^{10}Be deposition	
^{10}Be-flux in marine records			
8.2 - 31.1	16.3	ODP1063 (Bermuda Rise): 0-250 kyr BP	<i>Christl et al., 2007, 2010</i>
7.8 - 37.1	16.1	ODP983 (North Atlantic): 0-250 kyr BP	
9 - 28		Globally intergrated / long-term averaged ^{10}Be -fluxes	
15 - 35		ODP 1063A (Bermuda Rise): 170-200 kyr BP, IB exc.	<i>Knudsen et al., 2008</i>
10 - 70		ODP 983B (North Atlantic): 170-200 kyr BP, IB exc.	
5 - 27		ODP 925 (Ceara Rise): 0-7 Ma	<i>Murayama et al., 1997</i>
5 - 60		Globally stacked deep-sea sediments: 0-200 kyr BP	<i>Frank et al., 1997</i>
6 - 7		ACEX average ^{10}Be -flux for the past 12.3 Ma (Actic)	<i>Frank et al., 2008</i>
31.9 - 65.9	40.9	Portuguese Margin (20-45 kyr BP)	<i>Ménabréaz et al., 2011</i>
2 - 23		Four Arctic Ocean cores (F=C x S x D)	<i>Eisenhauer et al., 1994; Aldahan et al., 1997</i>
^{10}Be-flux in PC16 (Baffin Bay marine sediments)			
0.7 - 25.4 (1 - 35.5)	6.9 (9.7)	(values corrected)*	<i>This study</i>
fluxes by facies:			
6.5 - 8.6 (9.1 - 12)	7.6 (10.7)	Facies UB (Holocene)	
0.7 - 6.3 (1 - 8.8)	2 (2.8)	Facies BBDC (coarse carbonate-rich sediments)	
4.2 - 25.4 (5.8 - 35.5)	14.5 (20.2)	Facies LDC (fine feldspar-rich sediments)	
9.2 - 24.7 (12.8 - 34.6)	14.1 (19.8)	Facies OC (very fine feldspar-rich sediments)	

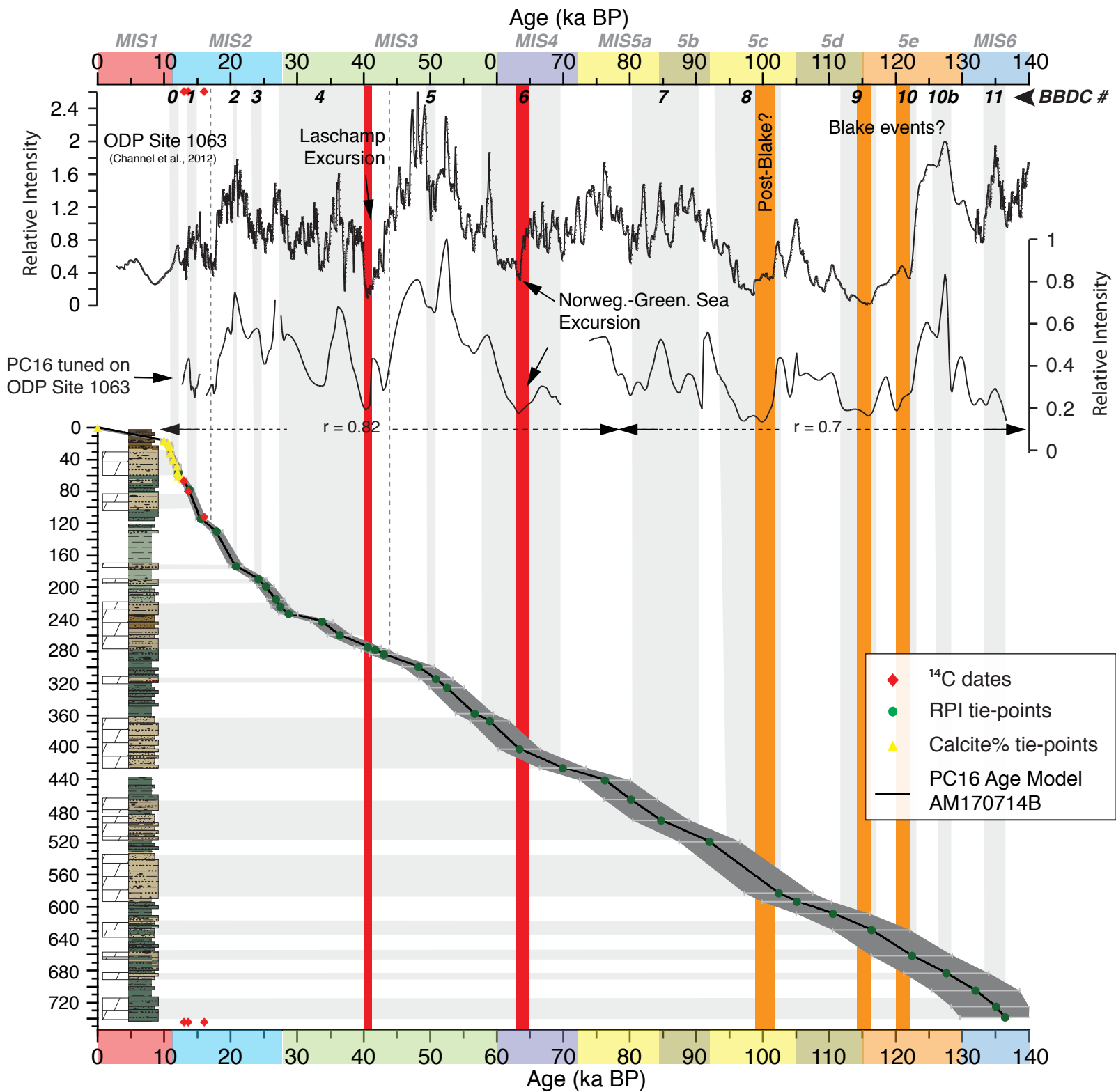
*correction to encompass the underestimation of the flux due to methodological biases, see text for details

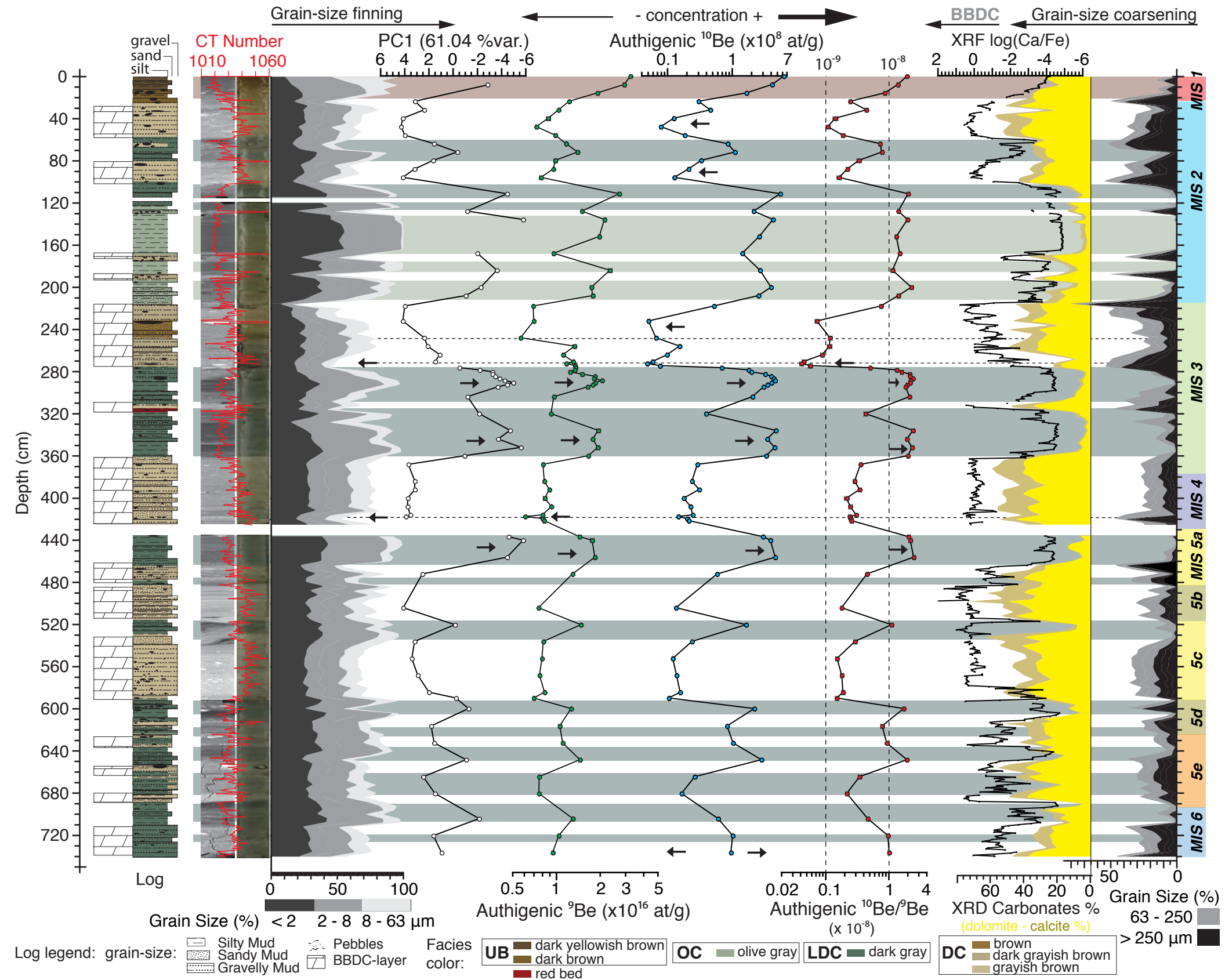
Figure

-  Inuitian Ice-Sheet
-  Laurentide Ice-Sheet
-  Greenland Ice-Sheet
-  LGM maximum extent?
-  Major LGM ice-streams

-  Core Location
-  Arctic Water
-  West Greenland Intermediate Water

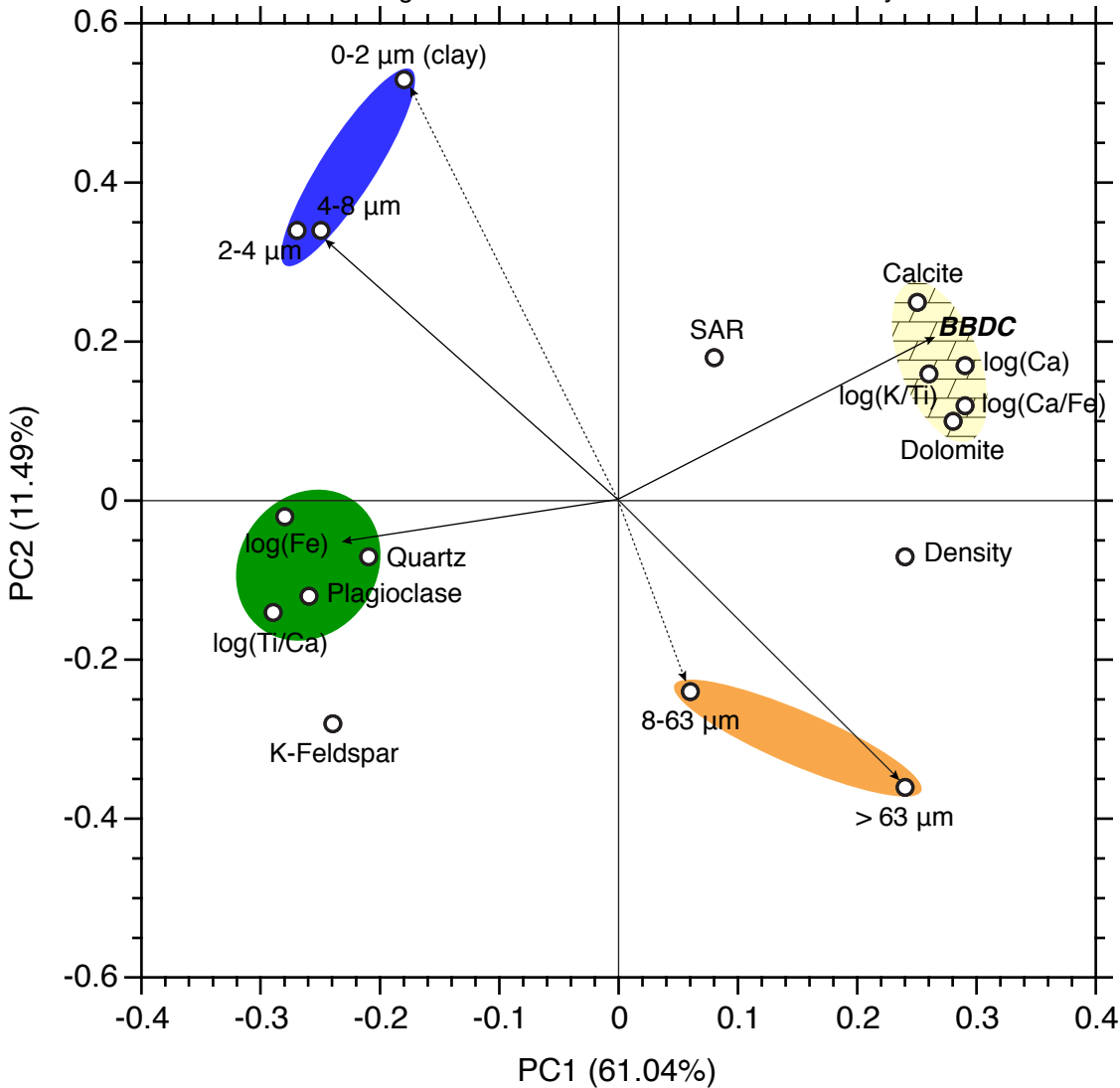


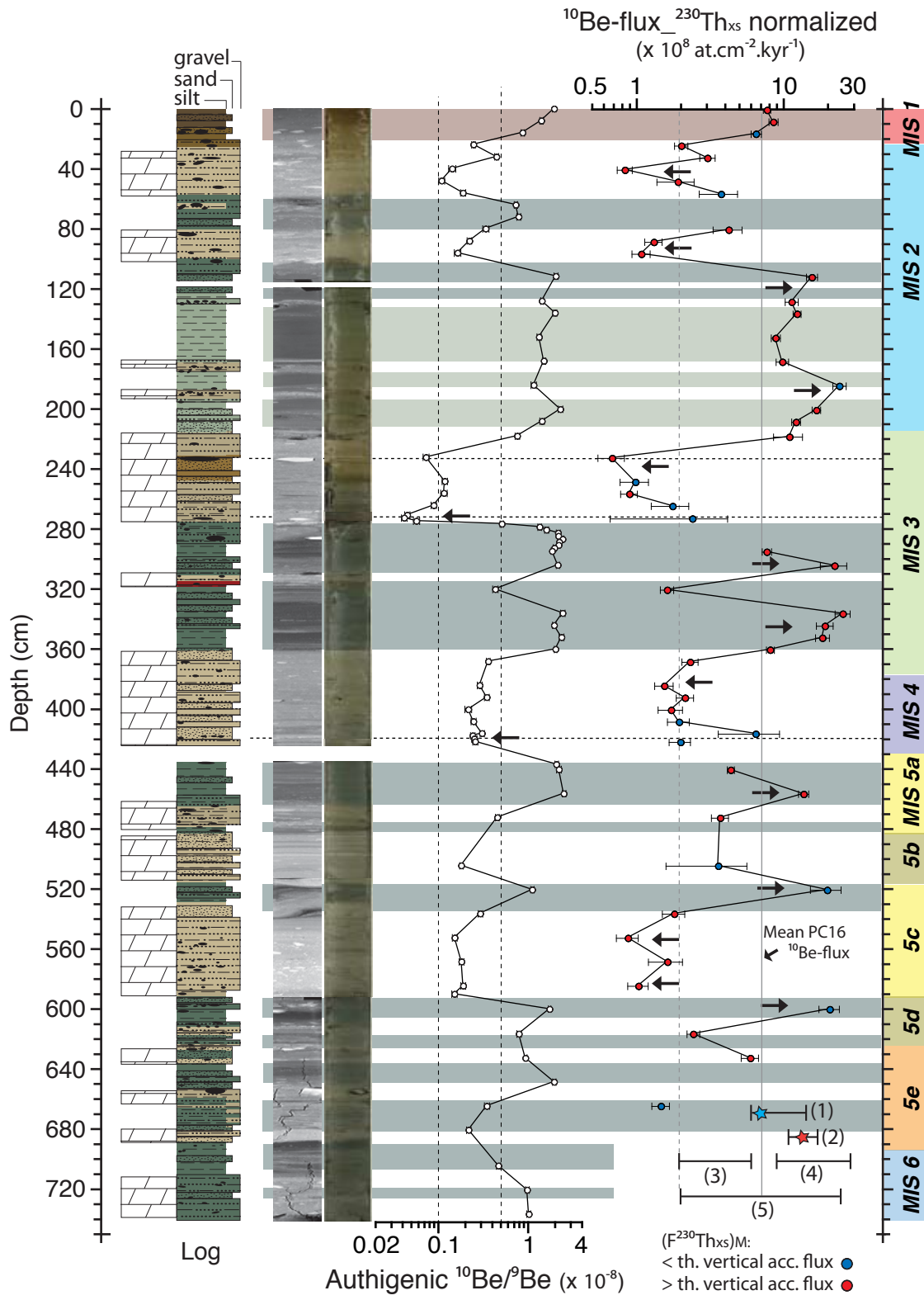


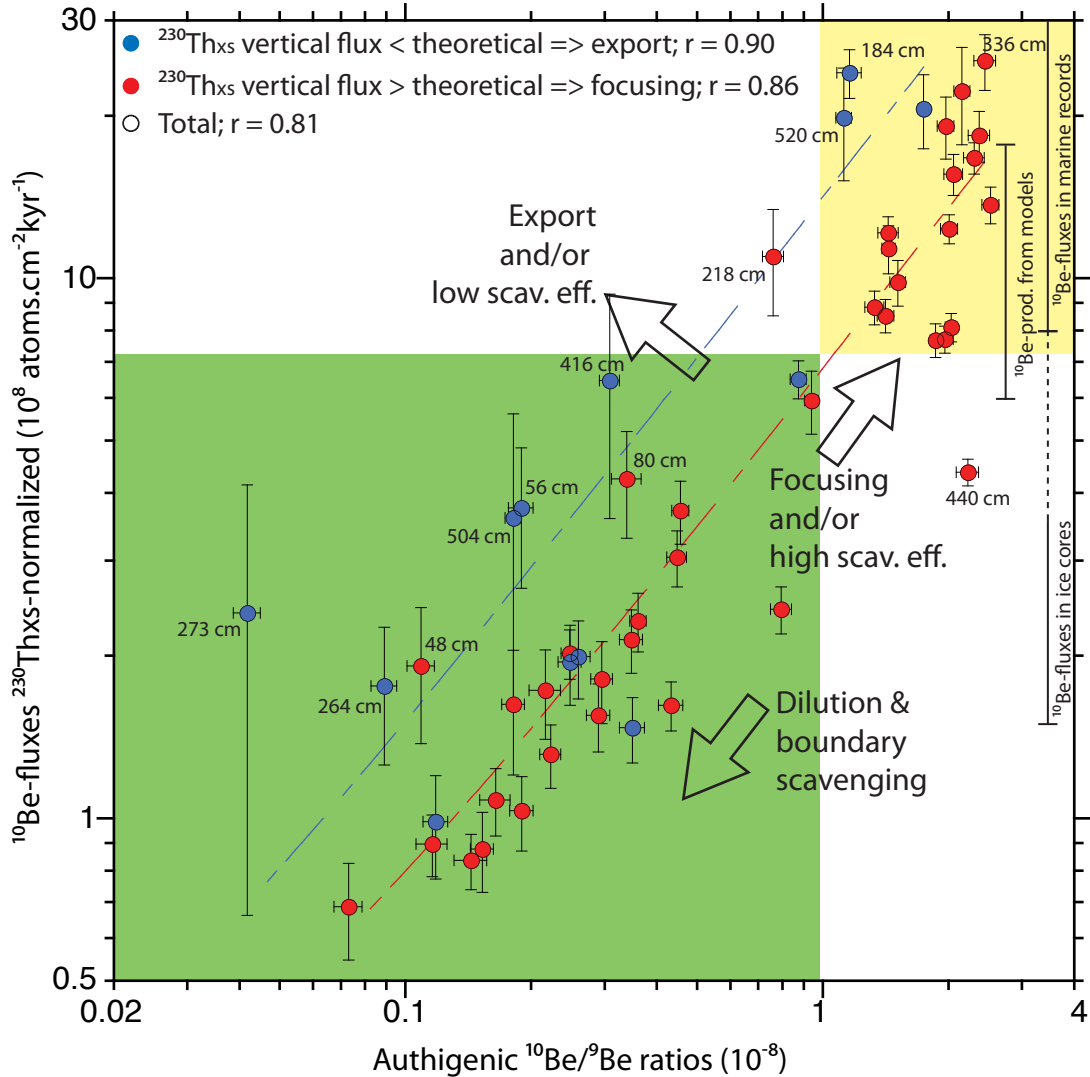


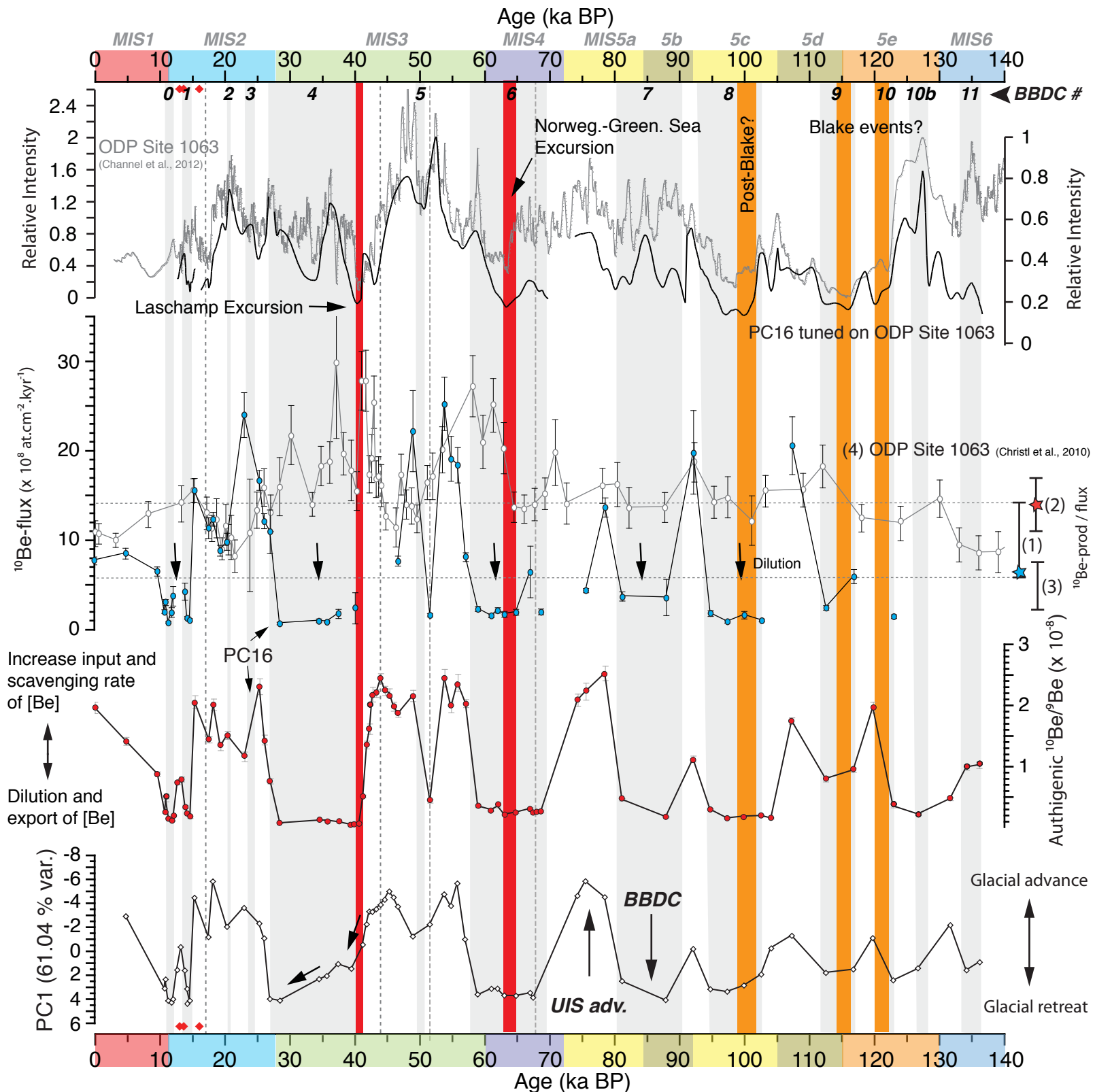
Fine sediments associated
with extended ice margins

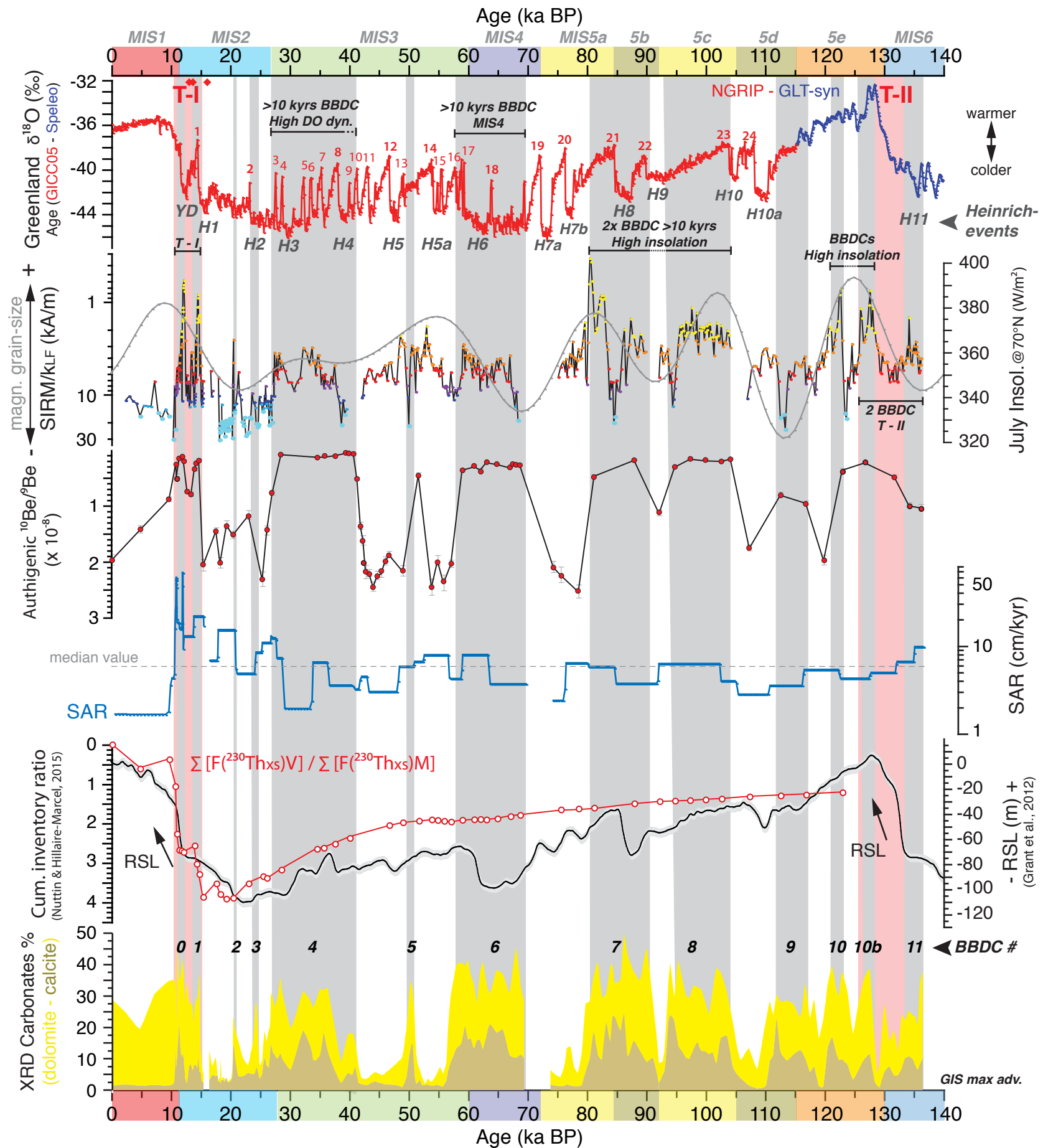
Coarse particles transported
by IRD and sea-ice











(a) Trans-Baffin drift**(b) Extended ice margins**