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The Laschamp geomagnetic dipole low expressed as a cosmogenic ^{10}Be atmospheric overproduction at 41 ka

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32 between the ^{14}C and U-Th ages is due to the Laschamp geomagnetic dipole low. Such new
33 high resolution datasets from records obtained from different latitudes will be required to
34 make significant advances in understanding the causes of atmospheric $\Delta^{14}\text{C}$ variations.

35

36 **Key words:** Cosmogenic nuclides; atmospheric ^{10}Be production rate; geomagnetic dipole
37 moment; Laschamp Excursion; marine sediments.

38

39 **1. Introduction**

40 The cosmogenic nuclide Beryllium-10 (^{10}Be , half-life: 1.387 ± 0.012 Ma (Chmeleff et
41 al., 2010; Korschinek et al., 2010)) is produced in the Earth's upper atmosphere by nuclear
42 interactions between energetic primary and secondary cosmic ray particles with target
43 elements O and N. As shown in the early work of Elsasser et al. (1956), and later refined by
44 Lal (1988), ^{10}Be atmospheric production is directly modulated by the geomagnetic field
45 strength over millennial time scales, following a negative power law. This relationship
46 between the dipole moment and the atmospheric production rate of cosmogenic nuclides was
47 initially inferred from archeomagnetic absolute paleointensities and neutron monitor data
48 comparison (Elsasser et al., 1956). More recently, a physical model of cosmic ray particle
49 interactions with atmospheric targets yielded compatible simulations (Masarik and Beer,
50 1999; Wagner et al., 2000). Regarding the classically-used paleomagnetic reconstructions,
51 ^{10}Be -derived paleointensity records can therefore constitute an alternative global and
52 independent reading of the dipole moment variations, and more particularly those that
53 accompany geomagnetic excursions and polarity events. During the last few years, efforts
54 have been made to extract a geomagnetic signal from both single and stacked ^{10}Be records in
55 natural archives such as ice sheets (e.g. Muscheler et al., 2005) and marine sediments (e.g.
56 Frank et al., 1997; Carcaillet et al., 2004a; Christl et al., 2007). For these latter archives, two
57 correction techniques were used to account for oceanic transport: the authigenic (adsorbed
58 from the water column) $^{10}\text{Be}/^9\text{Be}$ ratio (Bourlès et al., 1989; Henken-Mellies et al., 1990;
59 Robinson et al., 1995; Carcaillet et al., 2003; Carcaillet et al., 2004a; Carcaillet et al., 2004b;
60 Leduc et al., 2006) and $^{230}\text{Th}_{\text{xs}}$ -normalisation (e.g. Frank et al., 1997; Christl et al., 2003;
61 Christl et al., 2007; Christl et al., 2010). The sole attempt to compare both correction
62 techniques was presented by Knudsen et al. (2008), but large uncertainties and environmental

63 complications prevented them from isolating the ^{10}Be production component from the
64 $^{10}\text{Be}/^{230}\text{Th}_{\text{xs}}$ signal.

65 The last 50 ka is a key period for the study and calibration of geomagnetic excursions
66 and the associated cosmogenic responses, because this period contains the best-documented
67 and best-dated geomagnetic instability, the Laschamp Excursion (LE). This excursion was
68 first identified in the Laschamp and Olby lava flows of the Chaîne des Puys, France
69 (Bonhommet and Babkine, 1967), and was later described in lava flows of Iceland (e.g.
70 Kristjansson and Gudmundsson, 1980; Levi et al., 1990) and New Zealand (Cassata et al.,
71 2008) notably. Other records of the Laschamp Excursion and/or associated dipole low were
72 obtained from lacustrine and marine sediments (e.g. Thouveny and Creer, 1992; Vlag et al.,
73 1996; Lund et al., 2005; Channell, 2006). The most recent chronological constraints are
74 provided by Ar-Ar, K-Ar, and ^{230}Th -U dating of lava flows from France (Laschamp, Olby and
75 Louchadière), and from New Zealand (McLennans Hill) (Guillou et al., 2004; Plenier et al.,
76 2007; Singer et al., 2009). The statistical analyses of these age populations have led to the
77 following average ages: 40.4 ± 2.0 ka (Guillou et al., 2004) 40.7 ± 1.0 ka (Singer et al., 2009),
78 and 37.0 ± 0.7 ka (Plenier et al., 2007), the third thus introducing a significantly younger age
79 estimation.

80 Another younger, but more controversial, excursion has also been documented: the
81 Mono Lake (ML) Excursion that was first detected in the Wilson Creek formation, California,
82 by Denham and Cox (1971) as they were seeking for a record of the Laschamp Excursion. It
83 was later described in other northwestern American lacustrine (e.g. Negrini et al., 1984;
84 Liddicoat, 1992; Coe and Liddicoat, 1994) as well as in marine sediments (e.g. Nowaczyk
85 and Knies, 2000; Channell, 2006). The radiometric age of the ML Excursion has been revised
86 from ~ 24 ka (Denham and Cox, 1971) to ~ 32 ka (Negrini et al., 2000; Benson et al., 2003).
87 Surprisingly, other ^{39}Ar - ^{40}Ar dating of ash layers surrounding the excursion at Wilson Creek
88 (Kent et al., 2002) has yielded ages between 38 and 41 ka, which suggests that the excursion
89 recorded at Mono Lake could be the Laschamp Excursion. The correlation of the Mono Lake
90 RPI stack (Zimmerman et al., 2006) to the GLOPIS record of Laj et al. (2004) supported the
91 assignment of this excursion to the Laschamp. It remains true that most RPI records exhibit
92 two successive RPI lows at 32-34 ka and at 41 ka (e.g. Channel et al., 2006; Lund et al.,
93 2005), which is also supported by recent reports of excursions directions and low absolute
94 paleointensities in lava flows at ages near 32 ka, i.e. distinct from the Laschamp age (Cassata
95 et al., 2008; Singer et al., 2009; Kissel et al., 2011).

96 The dipole field reduction linked to the Laschamp Excursion has been recorded as
97 enhancements of ^{10}Be deposition rate in ice cores (e.g. Yiou et al., 1997; Muscheler et al.,
98 2004; Raisbeck et al., 2007) and marine sediments (e.g. Frank et al., 1997; Christl et al., 2003;
99 Carcaillet et al., 2004b; Leduc et al., 2006). Regarding the ML Excursion, a Chlorine-36
100 atmospheric production enhancement detected at around 32 ka in the GRIP ice core (Wagner
101 et al., 2000) constitutes the only documented evidence for a significant increase of the
102 cosmogenic production. Atmospheric cosmogenic nuclide production variations induced by
103 these geomagnetic instabilities need to be well constrained, notably because they impact ^{14}C
104 age calibration methods.

105 Here we present a highly resolved reconstruction of the ^{10}Be atmospheric production
106 variations over the 20-50 ka interval, obtained from a well-dated 10 m sediment sequence
107 deposited off the coast of Portugal. Authigenic $^{10}\text{Be}/^9\text{Be}$ ratios were measured, and compared
108 to $^{230}\text{Th}_{\text{xs}}$ -derived ^{10}Be fluxes which were investigated on the same samples. The ^{10}Be
109 production record was then transformed into a ^{10}Be -derived geomagnetic dipole moment
110 record.

111

112 **2. Material and methods**

113 ***2.1 Coring site and sediment description***

114 Climate and oceanographic settings as well as sedimentation processes within the area
115 have been described notably in Baas et al. (1997) and Paillet and Bard (2002). The sediment
116 deposition regime is related to glacial-interglacial sea-level variations and to the Tagus River
117 inputs. The contribution of large and variable continental material gave rise to a sediment
118 carbonate content of ~10% during glacials and ~50% during interglacials (Thomson et al.,
119 1999; Hall and McCave, 2000), and with less than 1% organic matter (Paillet and Bard,
120 2002). Hemipelagic sediments along the Portuguese Margin also contain scarce and thin Ice
121 Rafted Debris (IRD) material diluted in the clay matrix (feldspars, quartz, magnetite and
122 hematite-coated grains). These layers detected by their IRD abundance and Magnetic
123 Susceptibility (MS) signatures have been shown to be contemporaneous with Heinrich Events
124 (Bard et al., 2000; Thouveny et al., 2000; Moreno et al., 2002).

125 Using different methodologies, several studies have demonstrated that the upper part
126 of the giant cores collected with the Calypso corer of the R/V *Marion Dufresne* are affected

127 by significant oversampling (e.g. Thouveny et al., 2000; Skinner and McCave, 2003 and
128 Széréméta et al., 2004). This artefact has led to overestimations of apparent sedimentation
129 rates (e.g. Thompson et al., 1999). Despite their perfect stratigraphic preservation, the upper
130 12 m of core MD95-2042 are affected by this artefact, which hampered the evaluation of
131 accurate sedimentation rates and the interpretation of paleomagnetic and cosmogenic nuclides
132 results.

133 During the MD-140 PRIVILEGE campaign of the R/V *Marion Dufresne* (2004) on
134 the Portuguese Margin, a giant box corer CASQ (0.25m²x12m) allowed to collect a high
135 resolution, undisturbed sediment sequence (Core MD04-2811; Lat.: 37°48 N; Long.: 10°09
136 W; 3162m water-depth) at the site of the previously studied SU81-18 and MD95-2042 cores
137 (Fig. 1; see references in caption). These cores were extensively studied for radiocarbon
138 calibration (section 3.2) and therefore benefits from an excellent chronological control. Core
139 MD04-2811 was thus selected in order to document a lack of the MD95-2042 record in the
140 critical time interval (20-35 ka) and to assess accurate sediment fluxes using the ²³⁰Th_{xs}
141 normalization.

142 MD04-2811 sediments are composed of oxidized carbonate-rich silty clays in the
143 upper Holocene section (1.5m), and homogeneous hemipelagic silty clays deposited during
144 the last glacial period.

145 ***2.2 Paleomagnetic record and sampling strategy***

146 Paleomagnetic investigations were performed on 1.5 m long U-channels collected
147 along the MD04-2811 core. Low field Magnetic Susceptibility (MS) was measured using a
148 Bartington MS2C probe. As described in former studies (e.g. Thouveny et al., 2000), the
149 major structures of the MS profile consist in a succession of peaks corresponding to the
150 ferrimagnetic responses of IRD layers deposited during Heinrich Events 1 to 5 superimposed
151 on the paramagnetic contribution of the clayey fraction and a diamagnetic contribution of
152 carbonates (Fig. 2).

153 Natural Remanent Magnetization (NRM) and artificially induced magnetizations were
154 measured using a 2G cryogenic magnetometer 760-SRM model and were demagnetized by
155 Alternating Fields (AF). The Relative Paleointensity (RPI) curve was established by
156 normalizing the NRM intensity to the Anhyseretic Remanent Magnetization (ARM)
157 intensity. The NRM/ARM ratio obtained for the 30mT AF demagnetization step (NRM_{30mT}

158 /ARM_{30mT}) represents the best estimate of the RPI. Three main RPI lows are recorded in the
159 435-520 cm, 575-630 cm and 740-810 cm intervals, the latter displaying the lowermost values
160 (Fig. 2). This structure is classically attributed to the dipole low related to the Laschamp
161 Excursion (e.g. Laj et al., 2004; Thouveny et al., 2004). The mean inclination of the
162 Characteristic Remanent Magnetizations equals $\sim 50^\circ$, i.e. it is weaker than the inclination
163 (56.8°) of the field created by a geomagnetic axial dipole field (GAD) at the site latitude.
164 Unlike other records of the area (Thouveny et al., 2004), there is no anomalous direction
165 related to the RPI low.

166 Based on the interpretation of the RPI profile, 52 sediment samples were taken at a 30
167 cm resolution, increased to 10 cm within paleointensity lows intervals, for ^{10}Be and ^9Be
168 measurements.

169 **2.3 Analytical techniques**

170 *Leaching procedure: dissolution of the authigenic phase.*

171 In this study, authigenic $^{10}\text{Be}/^9\text{Be}$ ratios, interpreted as a proxy of ^{10}Be fluxes, are
172 presented. Following the procedure thoroughly described in Bourlès et al. (1989), the
173 authigenic phase is extracted by leaching $\sim 1\text{g}$ of dried and homogenized sediment using a
174 0.04M hydroxylamine in 25% acetic acid solution. A 2ml aliquot of the resulting leaching
175 solution is separated for natural ^9Be measurements. The remaining solution, spiked with
176 $300\mu\text{l}$ of a 10^{-3} g/g ^9Be -carrier, is used for chemical extraction and Accelerator Mass
177 Spectrometry (AMS) measurements of the resulting $^{10}\text{Be}/^9\text{Be}$ ratio. This procedure permits to
178 overcome changes in extraction efficiency.

179 *Total sample digestion and bulk sediment ^{10}Be content.*

180 $^{230}\text{Th}_{\text{xs}}$ -normalized ^{10}Be flux quantification requires a total dissolution of sediment
181 samples, assuming that lattice-bound ^{10}Be (from in-situ production) concentrations are
182 negligible with respect to authigenic content. Sample dissolution was achieved at 70°C in a
183 mixture of HF, HNO_3 and HCl (Bourlès et al., 1989), and spiked with $300\mu\text{l}$ of a 10^{-3} g/g ^9Be -
184 carrier to perform chemical extraction and AMS measurements.

185 *Beryllium isotopes chemical extraction and measurements.*

186 Once the sample is in solution, whatever the method used, Beryllium is then extracted
187 in the form of Be acetylacetonate using organic solvent. After decomposition of these

188 complexes, Be oxy-hydroxides are precipitated at pH 8 and oxidized to perform AMS
 189 measurements at ASTER, the French AMS national facility (installed at the CEREGE) which
 190 operates at 5MV. ^{10}Be concentrations are deduced from spiked $^{10}\text{Be}/^9\text{Be}$ measured ratios
 191 calibrated against the NIST 4325 Standard Reference Material with an assigned $^{10}\text{Be}/^9\text{Be}$
 192 value of 2.79×10^{-11} (Nishiizumi et al., 2007), and are decay-corrected using the ^{10}Be half-life
 193 of 1.387 ± 0.012 Ma (Chmeleff et al., 2010; Korschinek et al., 2010):

$$194 \quad [^{10}\text{Be}]_{g/g}^{sample} = \left(\frac{^{10}\text{Be}}{^9\text{Be}} \right)_M \cdot \frac{m_{spike} \cdot [^9\text{Be}]_{spike}}{m_{sample}} \cdot e^{\lambda_{10} \cdot t}$$

195 where $(^{10}\text{Be}/^9\text{Be})_M$ is the measured Be ratio, λ_{10} is the decay constant of ^{10}Be and m_{sample} is the
 196 weight of the leached sediment powder sample.

197 Measured ratios as well as the resulting ^{10}Be concentrations are listed in Table 1.
 198 Chemistry blanks cluster around 9×10^{-15} , up to 1000 times lower than the $^{10}\text{Be}/^9\text{Be}$ ratio of the
 199 samples.

200 The external reproducibility obtained on the NIST SRM 4325 standard (estimated
 201 from 1σ standard deviation of long-term repeated measurements and integrating all effects
 202 contributing to the variability of ASTER) is limited to 0.6% (Arnold et al., 2010).
 203 Uncertainties in the measured $^{10}\text{Be}/^9\text{Be}$ ratios and ^{10}Be concentrations were calculated from
 204 counting statistics and instrumental error propagation, according to the standard equation (e.g.
 205 Taylor, 1997).

206 Natural ^9Be concentrations were measured on the assigned 2ml aliquot using a
 207 graphite-furnace Atomic Absorption Spectrophotometer with a Zeeman effect background
 208 correction (Thermo Scientific ICE 3400) and the standard additions method (Table 1).
 209 Uncertainties in ^9Be concentrations depend on: (1) the reproducibility of the measured
 210 absorbances at each added ^9Be concentrations (standard deviation less than 1%), and (2) the
 211 least-square fitting between measured absorbances and added ^9Be concentrations ($R^2 > 0.999$).
 212 Uncertainties (1σ) in the ^9Be concentrations are generally significantly lower than 10%
 213 (average value: 4.3%).

214 The final $^{10}\text{Be}/^9\text{Be}$ uncertainty results from standard propagation of the above-cited
 215 ^{10}Be and ^9Be concentrations uncertainties.

216 Replicate authigenic ^{10}Be and ^9Be analyses were performed in order to check the
217 reproducibility of the Be isotopes chemical extraction (Table 1).

218 *Uranium and Thorium isotopic analyses: chemical procedure and TIMS measurements.*

219 U-series measurements were performed by Thermo-ionisation Mass Spectrometry
220 (TIMS) at the CEREGE) by using a ^{236}U - ^{233}U - ^{229}Th mixed spike calibrated according to the
221 procedure described in Deschamps et al. (2003).

222 The chemical procedure used to separate the uranium and thorium fractions for
223 determination of $^{230}\text{Th}_{\text{xs}}$ is similar to that described in Deschamps et al. (2004). Before total
224 digestion, samples were spiked. Total digestion was achieved by sequential treatments with
225 aqua regia and concentrated nitric and hydrofluoric acids. Following co-precipitation on iron
226 oxyhydroxides, U and Th fractions were chemically separated and purified using standard ion
227 exchange resins (AG-1X8 and U-Teva) and separation protocols modified from Deschamps et
228 al. (2004). Typical chemical blanks attained over the course of this study were about 0.25 ppb
229 for uranium and 0.28 ppb for thorium.

230 U and Th analyses were performed using a VG-Sector 54-30 mass spectrometer
231 equipped with a 30-cm electrostatic analyzer and an ion-counting Daly detector. The
232 instrumental abundance sensitivity is greater than 0.15 ppm at 1 amu (proportion of the ^{238}U
233 ion beam measured at mass 237). The detailed procedure is described in Bard et al. (1990)
234 with further modifications described in Deschamps et al. (2011). U and Th fraction was
235 loaded on the side filament of a triple zone-refined Re filament assembly. Reported errors
236 (2σ) for the $^{234}\text{U}/^{238}\text{U}$ and $^{230}\text{Th}/^{232}\text{Th}$ ratios are about 0.14 ‰ and 0.9 ‰, respectively. The
237 internal analytical reproducibility achieved in the course of this study on replicate
238 measurements of the NBS-960 international standard yielded a mean $\delta^{234}\text{U}$ value of
239 $36.5 \pm 0.8\text{‰}$ (2σ , $n=23$) which is in excellent agreement with values reported in the literature
240 (Deschamps et al., 2003; Andersen et al., 2004).

241 To calculate excess ^{230}Th activity ($^{230}\text{Th}_{\text{xs}}$), measured ^{230}Th concentration is corrected
242 for detrital and authigenic ^{230}Th components. The fraction of ^{230}Th supported by the decay of
243 detrital ^{238}U is estimated from the ^{232}Th sample concentration and an assumed average
244 lithogenic $^{238}\text{U}/^{232}\text{Th}$ activity ratio of $R = 0.6 \pm 0.1$ for the Atlantic Basin (Henderson and
245 Anderson, 2003). Ingrowth of ^{230}Th from decay of authigenic uranium is estimated by

246 assuming that authigenic uranium has an initial $^{234}\text{U}/^{238}\text{U}$ activity ratio equivalent to that of
 247 sea-water (i.e. 1.147, Andersen et al., 2010). $^{230}\text{Th}_{\text{xs}}$ activities were calculated as:

$$\begin{aligned}
 &^{230}\text{Th}_{\text{xs}} = ^{230}\text{Th}_{\text{Total}} \\
 &+ R_{\text{Crust}} \times ^{232}\text{Th}_{\text{Total}} \left(\left[\left(\frac{^{234}\text{U}}{^{238}\text{U}} \right)_0 - 1 \right] \times \frac{\lambda_{230}}{\lambda_{230} - \lambda_{234}} \left[e^{-\lambda_{234} \times t} - e^{-\lambda_{230} \times t} \right] - e^{-\lambda_{230} \times t} \right) \\
 &- ^{238}\text{U}_{\text{Total}} \left(1 - e^{-\lambda_{230} \times t} + \left[\left(\frac{^{234}\text{U}}{^{238}\text{U}} \right)_0 - 1 \right] \times \frac{\lambda_{230}}{\lambda_{230} - \lambda_{234}} \left[e^{-\lambda_{234} \times t} - e^{-\lambda_{230} \times t} \right] \right)
 \end{aligned}$$

252 where $^{230}\text{Th}_{\text{Total}}$, $^{232}\text{Th}_{\text{Total}}$ and $^{238}\text{U}_{\text{Total}}$ are the measured ^{230}Th , ^{232}Th and ^{238}U activities,
 253 respectively, and $^{238}\text{U}_{\text{Total}}$ is the sedimentary ^{238}U activity. $(^{234}\text{U}/^{238}\text{U})_0$ stands for the initial
 254 (decay-corrected) authigenic activity ratio and the decay constants of radioisotopes i are
 255 expressed by λ_i .

256 The preserved, decay-corrected, $^{230}\text{Th}_{\text{xs}}$ -normalized ^{10}Be deposition rates are then
 257 calculated following equations (11) and (12) as given in François et al. (2004), using a water
 258 depth of 3162 m.

259 Analytical errors and uncertainty related to assumption concerning the lithogenic
 260 $^{238}\text{U}/^{232}\text{Th}$ activity ratio, R , were properly propagated through $^{230}\text{Th}_{\text{xs}}$ and flux calculations.
 261 Note that owing to small analytical errors achieved by the TIMS method, most of the final
 262 errors on $^{230}\text{Th}_{\text{xs}}$ and flux calculations are related to the arbitrary fixed uncertainty on R .

263

264 **2.4 Using authigenic $^{10}\text{Be}/\text{Be}$ and $^{230}\text{Th}_{\text{xs}}$ -normalized ^{10}Be fluxes as tracers of geomagnetic** 265 **moment variations**

266 The particle-reactive ^{10}Be is mainly produced in the atmosphere through nuclear
 267 reactions (spallation reactions) on oxygen (O) and nitrogen (N), and is transferred to the
 268 Earth's surface in soluble form by precipitation (Raisbeck et al., 1981) within ~3 yrs (Baroni
 269 et al., 2011). Atmospheric ^{10}Be is ultimately removed from water via settling particles and is
 270 either deposited in marine and lacustrine sediments, or is efficiently retained by continental
 271 sediment components, and finally enters the oceans through various paths and deposits in
 272 deep sea sediments. Due to its short atmospheric residence time and chemical reactivity, ^{10}Be
 273 records of production enhancements should be synchronous in the geophysical reservoirs. In
 274 marine sediments, the ^{10}Be concentration results from a complex interplay between several

275 processes: e.g. cosmogenic production, redistribution by atmospheric, riverine and oceanic
276 transport, adsorption and deposition processes. Considering the adsorption processes, for
277 example, the absolute ^{10}Be concentration is controlled by the scavenging efficiency and the
278 specific surface of the settling sedimentary particles and thus depends on environmental
279 conditions affecting their chemical and grain size composition. Consequently, ^{10}Be
280 concentrations are meaningless and in order to account for these dependencies, a correction
281 procedure is required. Soluble forms of ^{10}Be and ^9Be have different sources: while ^{10}Be is
282 cosmogenic, the stable isotope ^9Be originates from partial dissolution of detrital, aeolian and
283 riverine inputs (Brown et al., 1992). Once homogenized in seawater, both isotopes are
284 scavenged with the same efficiency. Authigenic $^{10}\text{Be}/^9\text{Be}$ ratio of marine surface sediments
285 reflects the Be isotope composition of the overlying deep waters, and its spatial variability is
286 mainly controlled by the proximity to continental ^9Be inputs (Bourlès et al., 1989). Although
287 the authigenic ^9Be normalization method has provided promising results in specific
288 environments (Bourlès et al., 1989; Carcaillet et al., 2003; Carcaillet et al., 2004a; Carcaillet
289 et al., 2004b; Lebatard et al., 2010), quantitative reconstructions of the ^{10}Be fluxes to the
290 sediment requires taking into account syndepositional lateral transport of adsorbed Be.
291 Because residence time of Be in the water column is about 500-1000 yrs, boundary
292 scavenging and deepwater circulation must also contribute to the removal of dissolved Be and
293 thus influence the $^{10}\text{Be}/^9\text{Be}$ ratio. Moreover, lateral transport of adsorbed Be can be
294 significant in sediments heavily affected by focusing, such as drift deposits. However, at
295 ocean margins where particle flux is high, the short Be residence time allows to record the
296 ^{10}Be atmospheric flux variations with almost no signal attenuation.

297 Assuming a scavenged flux equivalent to its known production rate in the overlying
298 water column (Bacon, 1984), ^{230}Th can be used to quantify rates of particle rain to the
299 seafloor and to correct for syndepositional sediment redistribution by bottom currents. The
300 principles and limitations of this method are fully reviewed in Henderson and Anderson
301 (2003) and François et al. (2004). Like authigenic $^{10}\text{Be}/^9\text{Be}$, the $^{10}\text{Be}/^{230}\text{Th}_{\text{xs}}$ ratio may also be
302 potentially influenced by advection of dissolved Be. Moreover, biases resulting from nuclide
303 characteristics (different chemical affinities) are introduced: one illustration of this problem is
304 reported by Chase et al. (2002), who showed a composition-dependant differential scavenging
305 of Be and Th, suggesting that opal/carbonate ratio of settling particles may partly explain the
306 sedimentary $^{10}\text{Be}/^{230}\text{Th}_{\text{xs}}$ variability.

307 Here, authigenic $^{10}\text{Be}/^9\text{Be}$ ratios were measured for 51 sampled layers, and an
308 additional $^{230}\text{Th}_{\text{xs}}$ -normalization was performed for 9 of them, which were selected using the
309 Be isotopes ratio profile. This allows direct comparison and cross-calibration on the same
310 homogenized material.

311 **3. Results**

312 ***3.1 The authigenic $^{10}\text{Be}/^9\text{Be}$ ratios and RPI on the same sequence***

313 Authigenic ^9Be concentrations vary from 4.35×10^{-7} g/g to 7.22×10^{-7} g/g (Table 1),
314 refining and confirming the previous range obtained on adjacent core MD95 2042 by
315 Carcaillet et al. (2004). Decay-corrected authigenic ^{10}Be concentrations vary around a mean
316 value of 7.32×10^{-15} g/g with a prominent increase to 13.6×10^{-15} g/g at 712 cm. The range of
317 variation (5.33 to 13.6×10^{-15} g/g) and the average value are identical to those of the adjacent
318 core MD95 2042 (Carcaillet et al., 2004b). The resulting authigenic $^{10}\text{Be}/^9\text{Be}$ evolution is
319 presented in Figure 2, and compared to the RPI and MS profiles. The main feature of the
320 $^{10}\text{Be}/^9\text{Be}$ record is a unique significant enhancement at 712 cm, with ^{10}Be concentrations
321 leading to a near doubling of the long-term average $^{10}\text{Be}/^9\text{Be}$ value of 1.44×10^{-8} to 2.69×10^{-8} .

322 Along this sequence, both records of the ^{10}Be atmospheric production and its
323 geomagnetic modulation display inverse long-term trends. While the RPI record exhibits
324 three minima, only one significant enhancement appears in the $^{10}\text{Be}/^9\text{Be}$ record, with values
325 above the “mean+1 σ ” occurring between 662 and 722 cm depth. This interval overlaps the
326 depth interval recording the lowest RPI values located between ~690 and ~820 cm. The depth
327 shift thus evidenced between these two expressions of the dipole low is further discussed in
328 section 4.1

329 ***3.2 Chronostratigraphy***

330 The first step in establishing the chronostratigraphy of core MD04-2811 is to construct
331 a correlation matrix based on the MS records (Fig. 3) with the dated neighboring core MD95
332 2042 (Lat.: 37°45'N; Long.: 10°10'W; 3146 m water depth) in order to translate the available
333 chronological data onto the MD04-2811 depth scale (Fig. 4).

334 Two major studies have established radiocarbon dates series on MD95-2042 (Bard et
335 al., 2004; Shackleton et al., 2004, Electronic Appendix, Table A.1) that were compared with
336 ages obtained from correlations with annually laminated Greenland ice cores (GRIP and
337 GISP2). Among the paleoclimatic proxies measured along the MD95-2042 sequence, sea-

338 surface temperature proxies (alkenones and planktonic $\delta^{18}\text{O}$) show marked variations which
339 are in robust mutual stratigraphic agreement, and which correlate to Dansgaard-Oeschger
340 cycles and Heinrich Events. For this study, the age control on the core MD95-2042 was
341 updated by tuning the alkenone-derived sea-surface temperature record with the NGRIP
342 oxygen isotopes record using a minimum number of tie points ($R= 0.876$, Fig. 3 and 4).

343 Uncertainties affecting the age model are derived from the two-sigma GICC05 time-
344 scale uncertainties associated with multi parameter counting of annual layers (Andersen et al.,
345 2006; Rasmussen et al., 2006): the glacial part of this time scale has an estimated associated
346 error of 2% back to 40 ka and of 5–10% back to 57 ka. Although multiple correlation
347 procedures - limited by precision and resolution of the individual time series - and the
348 accuracy of the cross-correlation procedure represent a source of uncertainty which is difficult
349 to quantify using simple statistical parameters, the computed age model is validated by the
350 fact that the ages assigned to the Heinrich Events (Fig. 4) are indistinguishable from the
351 calibrated ^{14}C ages obtained on North Atlantic basin sediment sequences (Thouveny et al.,
352 2000). Moreover, the ^{14}C ages obtained on the studied sequence calibrated using the
353 INTCAL09 dataset (Reimer et al., 2009) relying mainly on $^{234}\text{U}/^{230}\text{Th}$ absolute dating of Hulu
354 Cave stalagmites (China, Wang et al., 2001) over the 35-45 ka period (Electronic Appendix,
355 Table A.1) are fully compatible with this ice core-derived chronology.

356 **3.3 ^{10}Be deposition rates (excess ^{230}Th -normalized ^{10}Be fluxes)**

357 Bulk ^{10}Be concentrations obtained after total sample dissolution are comprised
358 between 1.1×10^{-14} g/g and 1.8×10^{-14} g/g (see Table 2). Total mass flux ranges between 4.79
359 and $6.09 \text{ g/cm}^2/\text{ka}$, and is fully compatible with results obtained further northward on the
360 Portuguese Margin by Thomson et al. (1999), who determine a total sediment accumulation
361 flux of $\sim 5 \text{ g/cm}^2/\text{ka}$ during glacial times over the last 140 ka. Preserved vertical ^{10}Be
362 deposition rates vary from 3.18×10^9 to 6.59×10^9 atoms/cm²/ka (Fig. 5a). The ^{10}Be deposition
363 rate downcore evolution is strictly similar to that of the authigenic $^{10}\text{Be}/^9\text{Be}$ ratios measured
364 on the same samples: the characteristic increase with a maximum at 712 cm depth is indeed
365 reflected in the ^{10}Be deposition rates record. The comparison between authigenic $^{10}\text{Be}/^9\text{Be}$
366 ratios and $^{230}\text{Th}_{\text{xs}}$ -normalized ^{10}Be fluxes is illustrated in Figure 5b. The linear relationship
367 between these two variables is highly significant ($R = 0.912$), albeit mainly carried by the
368 maximum variability recorded during the main enhancement.

369

370 **4. Discussion**

371 ***4.1 Identification of the Laschamp dipole low and chronological implications***

372 During low dipole field intervals, the magnetic torques are insufficient to align
373 magnetic grains and so ensure acquisition of a well-defined magnetization. Nevertheless, after
374 recovery of the field strength, re-alignments of magnetic grains and acquisition of viscous
375 magnetization (VRM) result in a biased record of directions and intensities (e.g. Coe and
376 Liddicoat, 1994). In core MD04-2811, as in other neighboring cores (see Figures 4 and 9 in
377 Thouveny et al., 2004), the structure of paleomagnetic records (directions and RPI) does not
378 allow an accurate and precise stratigraphic positioning of the dipole field minimum.
379 Comparison of the positions of the respective boundaries of the main RPI low (at ~700 and
380 ~820 cm) with the main $^{10}\text{Be}/^9\text{Be}$ enhancement (at ~660 cm and ~725 cm) calls attention to a
381 ~50 cm lag between the two signals (Fig. 2). This lag integrates a maximum 10 cm shift due
382 to the residence time with respect to scavenging of Be, given a ~1 cm/50 yrs average
383 sedimentation rate. The remaining 40 cm lag is greater than the 15-30 cm lags revealed from
384 neighboring cores by Carcaillet et al. (2004); however given the sources of distortions cited
385 above, it is not possible to assign this lag to the magnetization lock-in depth only.

386 In core MD95-2042 core, the declination anomaly and the RPI low related to the
387 Laschamp excursion are recorded 50 cm beneath the MS peak associated with the Heinrich
388 Event 4 (Thouveny et al., 2004). In North Atlantic Ocean cores (Kissel et al., 1999),
389 excursions peaks related to the Laschamp are recorded at the same stratigraphic
390 position as rock-magnetic signatures of the Dansgaard-Oeschger (D-O) Interstadial 10, i.e.
391 prior to H4. Nevertheless, the best stratigraphic indicator of the occurrence of the dipole low
392 related to the Laschamp is the cosmogenic nuclide signature which is not affected by delays
393 and distortions imparted by post-depositional processes.

394 Along core MD04-2811 the highest $^{10}\text{Be}/^9\text{Be}$ ratio peak - recorded ~40 cm beneath the
395 MS peak corresponding to H4 - can thus undoubtedly be interpreted as the ^{10}Be atmospheric
396 overproduction due to the geomagnetic dipole reduction associated with the Laschamp
397 Excursion (LE). This 40 cm depth interval suggests that the LE occurred ~2 ka before
398 Heinrich Event 4, which is dated at ~39 ka in North Atlantic sedimentary sequences (e.g.
399 Thouveny et al., 2000). Indeed, the chronological reconstruction positions the $^{10}\text{Be}/^9\text{Be}$
400 Laschamp signature at 41.2 ± 1.6 ka (minimal uncertainty derived from GICC05 chronology).
401 This age is coherent with the most recent independent radiometric age determinations

402 obtained on volcanic material (40.4 ± 2.0 ka, Guillou et al., 2004; 40.7 ± 0.9 ka, Singer et al.,
403 2009). It also coincides tightly with the age of the maximum ^{10}Be flux in Summit ice cores
404 (Muscheler et al., 2005; Svensson et al., 2008). This suggests that the ^{10}Be residence time in
405 the water column in such settings is short enough to preserve the simultaneity of the record in
406 both geological reservoirs, considering age uncertainties and resolution.

407 The ^{10}Be production and recording processes in sedimentary sequences that are totally
408 independent from: 1) the local geomagnetic vector, and 2) magnetization processes in weak
409 and highly variable local fields, enable to evaluate the duration of the dipole field anomaly.
410 Both the high resolution and the accurate dating of this record highlight the persistence of a
411 weak dipole field defined by authigenic $^{10}\text{Be}/^9\text{Be}$ values above “mean+1 σ ” during ~ 2.5 ka.
412 The minimum field intensity phase defined by authigenic $^{10}\text{Be}/^9\text{Be}$ values above “mean+2 σ ”
413 lasted about 1 ka. This duration estimate is similar to that provided by the ice core record of
414 Muscheler et al. (2005).

415 **4.2 Significance of the $^{10}\text{Be}/^9\text{Be}$ record**

416 *$^{10}\text{Be}/^9\text{Be}$ ratios and $^{230}\text{Th}_{\text{xs}}$ -normalized ^{10}Be fluxes: a cross evaluation*

417 In this study, ^{10}Be deposition rates reinforce the authigenic $^{10}\text{Be}/^9\text{Be}$ signal validity.
418 Comparison between the amplitude of $^{10}\text{Be}/^{230}\text{Th}_{\text{xs}}$ and authigenic $^{10}\text{Be}/^9\text{Be}$ variations shows
419 that both records agree within uncertainties (Fig. 6a) and document an increase by a factor of
420 ~ 2 coincidental with the Laschamp geomagnetic dipole low. This is the first time that such a
421 match has been reported, demonstrating that both normalization methods provide equivalent
422 results in this case, and that normalizing authigenic ^{10}Be concentrations by authigenic ^9Be
423 concentrations allows a reliable correction for total particle flux variations. Moreover,
424 quantification of the vertical ^{10}Be fluxes allows to underline the effects of boundary
425 scavenging: the ^{10}Be deposition rates determined in this study (mean value: $4.09 \pm 0.31 \times 10^9$
426 $\text{at./cm}^2/\text{ka}$) are (1) both slightly higher than an estimate of the present global production rate
427 of $1.21 \pm 0.26 \times 10^9 \text{at./cm}^2/\text{ka}$ (Monaghan et al., 1986) and higher than Greenland ^{10}Be fluxes
428 (0.25 to $0.6 \times 10^9 \text{at./cm}^2/\text{ka}$ (Muscheler et al., 2004)); and (2) within the range reported for
429 Atlantic drift deposits of the last 75 ka (~ 1 to $7 \times 10^9 \text{at./cm}^2/\text{ka}$, Christl et al., 2007; Knudsen et
430 al., 2008; Christl et al., 2010). Although the resolution of the $^{10}\text{Be}/^{230}\text{Th}_{\text{xs}}$ record must be
431 improved, the observed close agreement suggests that the effects of rapid millennial-scale
432 changes in deep water circulation (advection of dissolved Be and Th) could be either similar
433 in both approaches, or insignificant.

434 *Comparison with ^{10}Be production rate ice core records and with relative geomagnetic field*
435 *intensity reconstructed from paleomagnetic records*

436 Over the 20-50 ka interval, there is a remarkable match between the MD04-2811
437 authigenic $^{10}\text{Be}/^9\text{Be}$ record and the ^{10}Be flux variations recorded along the GRIP and GISP2
438 ice cores (Fig. 6b; Finkel and Nishiizumi, 1997; Yiou et al., 1997; Muscheler et al., 2005). It
439 must be emphasized that even small variations of the two records match which strongly
440 supports the high quality of the authigenic $^{10}\text{Be}/^9\text{Be}$ record presented. Observed for such
441 different geophysical reservoirs as marine sediments and polar ice, this agreement
442 demonstrates that this ratio most likely mainly reflects the changes in global atmospheric ^{10}Be
443 production rate. This confirms that the normalization of authigenic ^{10}Be concentrations to
444 authigenic ^9Be concentrations minimizes the impact of secondary mechanisms, among which
445 ocean system effects (see section 2.4). In addition, the fact that these variations are recorded
446 with the same amplitude in both archives, indicates that the ^{10}Be residence time in the
447 atmosphere-ocean reservoirs should be short enough to minimize the attenuation of the
448 production signal. The Laschamp signature corresponds to the near-doubling of the global
449 atmospheric ^{10}Be production expected from the physically constrained algorithm (Elsasser et
450 al., 1956; Lal, 1988) which expresses the control of global atmospheric ^{10}Be production rate
451 exerted by the magnetospheric shielding as modulated by geomagnetic moment variations.,
452 This also suggests that a significant component of the Greenland ^{10}Be flux signal results from
453 an atmospheric transport from lower latitudes. Another important result is that these marine
454 and ice core records of the atmospheric ^{10}Be production rates both fail to display any other
455 significant ^{10}Be deposition enhancement. Wagner et al. (2000a) identified in the GRIP ice
456 core an increased ^{36}Cl flux between D-O events 6 and 7, which they attributed to the RPI
457 minimum observed at ~34 ka in the NAPIS-75 stack (based on the GISP-2 chronology; Laj et
458 al., 2000), and referred to as the Mono Lake Excursion. After smoothing of the dataset with a
459 $1/2000 \text{ yrs}^{-1}$ low-pass filter, the ^{36}Cl peak amplitude appears equivalent to that of the
460 Laschamp. Surprisingly, in Wagner et al. (2000b), smoothing of the dataset with a $1/3000 \text{ yrs}^{-1}$
461 low-pass filter reduces the ^{36}Cl Mono Lake peak previously noted in Wagner et al. (2000a)
462 to the level of an insignificant variation. Together with this observation, the ^{10}Be fluxes in
463 Summit ice cores and the presented high-resolution $^{10}\text{Be}/^9\text{Be}$ record question the validity of a
464 ^{36}Cl signature linked to the Mono Lake dipole field low.

465 Lal's algorithm (1988) has been used to convert the authigenic $^{10}\text{Be}/^9\text{Be}$ record into a
466 relative Virtual Dipole Moment (VDM) record that is compared to the GLOPIS-75 stacked

467 paleomagnetic record (Fig. 6c). This reconstruction and the GLOPIS-75 curve agree relatively
468 well over the 30-48 ka time interval, although the cosmogenic record points out possible
469 higher intensities for the 20-30 ka period and beyond 48 ka than indicated in the
470 paleomagnetic records of the GLOPIS-75 stack.

471 ***4.3 Reconstruction of the geomagnetic dipole moment variations from the $^{10}\text{Be}/^9\text{Be}$ ratios***

472 *A stack of the Portuguese Margin $^{10}\text{Be}/^9\text{Be}$ records*

473 In order to achieve a statistically significant record of the authigenic $^{10}\text{Be}/^9\text{Be}$ ratio,
474 the MD04-2811 record was combined with the record previously obtained by Carcaillet et al.
475 (2004) on core MD95-2042, recovered at the same site. These data were relocated on the
476 depth scale of core MD04-2811 using the MS correlation matrix (reciprocal to the one used to
477 transfer chronological data (Fig. 7a)). Be ratios measured on both cores vary in a similar
478 range and present the same structure characterized both by a sharp maximum linked to the LE
479 dipole low and by the absence of any other significant peak. Differences between MD95-2042
480 and MD04-2811 data sets result from local sedimentary processes, differential post-
481 depositional processes (i.e. compaction), sampling and analytical uncertainties. After stacking
482 the two $^{10}\text{Be}/^9\text{Be}$ ratios records, a smoothed composite record was produced by computing
483 weighted moving averages over a 1000 years sliding window offset by 500 years (except for
484 the 20-25 ka interval, for which MD04-2811 data are directly reported). Associated
485 uncertainties combine both the sample analytical uncertainty and standard deviation around
486 each weighted average, and are calculated using the unbiased weighted estimator of the
487 sample variance (see Electronic Appendix Table A.2). Fig. 7b shows this composite record
488 and the associated standard error (1σ). The 1000 yrs-smoothed authigenic $^{10}\text{Be}/^9\text{Be}$ composite
489 record displays a ~ 1.45 time increase during the LE time interval compared to the long-term
490 average (1.32×10^{-8}). Although the studied time interval covers the age range of the Mono
491 Lake Excursion, the slight increase of the $^{10}\text{Be}/^9\text{Be}$ composite curve at ~ 33 -34 ka is not
492 significant regarding the uncertainties associated with the adjacent data points. This reveals
493 the absence of any ^{10}Be overproduction linked to a dipole reduction associated with the Mono
494 lake Excursion. This observation -if confirmed at other sites- would have strong implications
495 on the interpretation of the magnitude and duration of the “geomagnetic dipole low”
496 associated with the Mono lake Excursion.

497

499 The previously described ^{10}Be -stack was converted into Virtual Dipole Moment
500 (VDM) variations using absolute paleointensities reconstructed from lava flows. In order to
501 reproduce the full range of the ^{10}Be production variation, the $^{10}\text{Be}/^9\text{Be}$ ratios measured along
502 core MD04-2811 were associated with VDM values selected from the GEOMAGIA-50
503 (Korhonen et al., 2008) absolute paleointensity database (see Electronic Appendix Fig. A.3
504 and Table A.3). ^{10}Be production rates are inversely proportional to VDM values, thus (1) the
505 $^{10}\text{Be}/^9\text{Be}$ maximum value was assigned to the minimum VDM value linked to the LE, (2) the
506 intermediate $^{10}\text{Be}/^9\text{Be}$ ratios were assigned to the intermediate VDM values, and (3) the
507 minimum $^{10}\text{Be}/^9\text{Be}$ values were assigned to the maximum VDM values.

508 The relationship thus obtained between authigenic $^{10}\text{Be}/^9\text{Be}$ ratios and VDMs is, at
509 best, fitted by a polynomial of order 2 (Fig. A.3). When applied to the $^{10}\text{Be}/^9\text{Be}$ stack this
510 polynomial produces a “ ^{10}Be -derived VDM” record for the 20-50 ka interval, which can be
511 compared to GLOPIS-75 and SINT-800 paleointensity stacks (Fig. 8). This ^{10}Be based VDM
512 record provides a reconstruction in better agreement with GLOPIS-75 than does Lal’s
513 algorithm conversion (Fig. 6c). It documents a stronger geomagnetic moment before
514 (maximum $\sim 10.5 \times 10^{22} \text{ A.m}^2$) than after (maximum $\sim 7.0 \times 10^{22} \text{ A.m}^2$) the LE, and allows to
515 estimate a decreasing rate of $\sim 1 \times 10^{22} \text{ A.m}^2/\text{ka}$ for the dipole collapse which initiates the
516 Laschamp (minimum $\sim 2.6 \times 10^{22} \text{ A.m}^2$ reached at $\sim 41 \text{ ka}$). Although this record exhibits the
517 same main variations as GLOPIS-75, it fails to reproduce a sharp VDM reduction at the age
518 of the Mono Lake Excursion: at 34-32 ka relatively low VDM values ($\sim 4.7 \times 10^{22} \text{ A.m}^2$) are in
519 opposition with the sharp increase recorded at 32-33 ka in the GLOPIS-75.

520

521 **5. Conclusion**

522 Authigenic $^{10}\text{Be}/^9\text{Be}$ ratios measured along a rapidly accumulating sedimentary
523 sequence from the North-East Atlantic (core MD04-2811) provide a new record of dipole
524 geomagnetic moment variations over the 20-50 ka time interval, independent from
525 paleomagnetic methodological constraints.

526 This first successful geomagnetic reconstruction using both authigenic $^{10}\text{Be}/^9\text{Be}$ and
527 $^{10}\text{Be}/^{230}\text{Th}_{\text{xs}}$ normalization techniques of ^{10}Be concentrations in marine sediments evidence an
528 almost doubling of the atmospheric ^{10}Be production which documents the occurrence of the

529 Laschamp geomagnetic dipole low prior to the Heinrich Event 4, at an age of ~ 41 ka. These
530 ratios are fully compatible with the Summit ice core ^{10}Be record, which further supports their
531 significance in terms of ^{10}Be atmospheric production. Identical ^{10}Be flux responses recorded
532 in a polar ice core and in a mid-latitude marine core strongly suggest that the cosmogenic
533 nuclide overproduction signal in Greenland ice results from the global geomagnetic moment
534 modulation.

535 The authigenic $^{10}\text{Be}/^9\text{Be}$ ratio records of neighboring cores MD04-2811 and MD95-
536 2042 cores were compiled and averaged over a 1 ka sliding window. The obtained stack was
537 calibrated using absolute virtual dipole moment (VDM) values, determined from lava flow
538 paleointensities, to provide a ^{10}Be derived VDM record. This allows to characterize the
539 Laschamp dipole low by a decrease from $\sim 10.5 \times 10^{22} \text{ A.m}^2$ at ~49 ka to $\sim 2.6 \times 10^{22} \text{ A.m}^2$ at ~41
540 ka. The latter estimate is concordant with radiometric ages obtained on excursions lava
541 flows. This record fails to document a significant VDM reduction at the age of the Mono
542 Lake Excursion (32-34 ka).

543 By providing a new high-resolution ^{10}Be dataset at mid-latitude, the results presented
544 here are of great interest for better constraining the influence of geomagnetic field intensity
545 variations on cosmogenic nuclide production rates. Among these cosmogenic nuclides,
546 atmospheric ^{14}C is of particular importance considering that the ^{14}C calibration curve lacks
547 accurate data during the time period covered by this study. Such new high-resolution datasets
548 from different latitudes are required to make significant advances in quantifying all processes
549 involved in atmospheric $\Delta^{14}\text{C}$ variations. Similarly, reconstructing past VDM variations is of
550 fundamental importance in quantifying in-situ cosmogenic nuclide production rate variations
551 and thus in accurately measuring Earth's surface processes. Further investigations, notably at
552 low-latitude sites where the shielding effect from the geomagnetic dipole field is maximized,
553 are currently underway.

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Figure Captions :

919 **“The Laschamp geomagnetic dipole low expressed as a cosmogenic ^{10}Be atmospheric**
920 **overproduction at ~41 ka.”**

921 By L. Ménabréaz *, N. Thouveny, D.L. Bourlès, P. Deschamps, B. Hamelin and F. Demory.

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923 *Revised version submitted to Earth and Planetary Science Letters*

924 **Fig. 1.** Location map of the studied MD04-2811 core. Other cores cited in the text
925 (MD95-2042 and SU81-18 were recovered at the same site.

926 **Fig. 2.** Magnetic properties and cosmogenic nuclide ^{10}Be record as a function of depth in core
927 MD04-2811. (a) Magnetic susceptibility; susceptibility peaks identify distal ferrimagnetic
928 IRD layers deposited during Heinrich Events 1 to 5 (labeled H1 to H5). (b) Relative
929 paleointensity (NRM/ARM ratios at 30mT alternating field step). (c) Authigenic $^{10}\text{Be}/^9\text{Be}$
930 ratios (10^{-8}) with 1σ uncertainty. Solid and dotted lines correspond to the mean value over the
931 20-50 ka time period and the associated standard deviation (1σ), respectively (see Table 1).

932 **Fig. 3.** (a-b) Paleoclimatic variations over Greenland and the Portuguese margin: (a)
933 variations of the air paleotemperature proxy over Greenland: $\delta^{18}\text{O}$ of the NGRIP ice core
934 (NGRIP-Members, 2004 and references therein) plotted on the GICC05 time-scale (NGRIP
935 dating group, 2006 and references therein); (b) variations of the Portuguese margin sea-
936 surface paleotemperature proxy: UK'37 alkenone index in core MD95-2042 (raw data of
937 Pailler and Bard, 2002) on a time scale tuned to NGRIP $\delta^{18}\text{O}$ using the Analyserie software
938 (Paillard et al., 1996). Correlation tie-points are marked by vertical dashes. Triangles indicate
939 the position of the 23 samples used for AMS dating: raw data (Electronic Appendix, Table
940 A.1), are from Bard et al., 2004 and Shackleton et al., 2004); (c-d) Magnetic Susceptibility
941 (MS) records of Portuguese margin cores as a function of depth: (c) core MD95-2042
942 (specific MS in $10^{-9} \text{ m}^3/\text{kg}$) (data from Thouveny et al., 2000 transferred on the NGRIP
943 chronology; (d) core MD04-2811 (volume susceptibility 10^{-5} SI). Susceptibility peaks (S0 to
944 S5) mark IRD layers deposited during the Younger Dryas (YD) and Heinrich events (H1 to
945 H5) events.

946 **Fig. 4.** Age-depth relationship in core MD04-2811. Open circles stand for the age model
947 derived from the GICC05 chronology: the MD95-2042 age control obtained tuning UK'37
948 alkenone index and NGRIP $\delta^{18}\text{O}$ (see Fig.3) was applied to core MD04-2811 by correlation
949 of MS profiles using the Analyserie software (Paillard et al., 1996). Ages of the MD04-2811
950 susceptibility peaks (grey bands, ages reported on the left) are undistinguishable from ages of
951 Heinrich layers computed by Thouveny et al. (2000): H2: 24778 ± 297 yr BP; H3: 31163 ± 509
952 yr BP; H4: 39379 ± 948 yr BP.

953 **Fig. 5.** (a) Downcore variation of the $^{230}\text{Th}_{\text{xs}}$ -normalized ^{10}Be fluxes and (b) relation with
954 authigenic $^{10}\text{Be}/^9\text{Be}$ ratios measured on the same homogenized sediment samples.

955 **Fig. 6.** (a and b) Relative temporal variations of ^{10}Be atmospheric production rate proxies
956 over the 20-50 ka interval. All records are normalized to their respective mean values over the
957 studied time interval. (a) Black dots represent the authigenic $^{10}\text{Be}/^9\text{Be}$ record and the
958 associated 1σ uncertainty. Grey dots represent the $^{230}\text{Th}_{\text{xs}}$ -normalized ^{10}Be fluxes. (b) Black
959 dots represent the authigenic $^{10}\text{Be}/^9\text{Be}$ record and the associated 1σ uncertainty, and grey line
960 represents the Greenland ^{10}Be flux in Summit ice cores on GICC05 time scale (Finkel et al.,
961 1997; Yiou et al., 1997; composite GRIP-GISP2: Muscheler et al., 2005). (c) Relative
962 temporal variations of geomagnetic field intensity over the 20-50 ka interval. Solid lines
963 represents the 1σ uncertainty envelop of the GLOPIS-75 stack: raw data from Laj et al.,
964 (2004) are normalized to 8×10^{22} A.m² (most recent GLOPIS VADM value) and plotted on
965 the GICC05 time scale using the NGRIP, GRIP and GISP2 synchronization match points
966 dataset (Rasmussen et al., 2006). Dots are the authigenic $^{10}\text{Be}/^9\text{Be}$ -derived VDM record
967 (M/M_0) calculated using the Lal's algorithm (1988): $P/P_0 = (M/M_0)^{-0.5}$, where P/P_0 is the
968 authigenic $^{10}\text{Be}/^9\text{Be}$ ratio divided by the surface sample value of $(1.36 \pm 0.35) \times 10^{-8}$,
969 undistinguishable within uncertainties from the long-term average $(1.44 \pm 0.043) \times 10^{-8}$.

970 **Fig. 7.** (a) Authigenic $^{10}\text{Be}/^9\text{Be}$ ratios measured along MD04-2811 (solid circles) and MD95-
971 2042 cores (open circle, Carcaillet et al., 2004) on the MD04-2811 depth scale. (b) Calculated
972 authigenic $^{10}\text{Be}/^9\text{Be}$ composite record over the 20-50 ka interval. This record is low-pass
973 filtered ($1/1000$ yr⁻¹). Associated uncertainties are 1σ .

974 **Fig. 8.** Geomagnetic dipole moment variation during the 20-50 ka interval. The black curve is
975 the Virtual Axial Dipole Moment (VADM) variations reconstructed from SINT800 (Guyodo
976 and Valet, 1999). The grey zone is the 1σ envelop of the GLOPIS-75 based VADM

977 reconstruction (Laj et al., 2004). Bold black curve is the authigenic $^{10}\text{Be}/^9\text{Be}$ -based VDM
978 (Virtual Dipole Moment) record and 1σ associated uncertainties.

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980

981 **Highlights to the manuscript:** “The Laschamp geomagnetic dipole low expressed as a
982 cosmogenic ^{10}Be atmospheric overproduction at ~ 41 ka.”

983

984 by L. Menabreaz *, N. Thouveny, D.L. Bourles, P. Deschamps, B. Hamelin and F. Demory.

985

Earth and Planetary Science Letters

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987 Cosmogenic ^{10}Be concentration was measured along a marine sediment core (20 -50ka)

988

Authigenic ^9Be and excess ^{230}Th were used to normalize ^{10}Be concentrations

989

A doubling of the cosmogenic production is revealed at ~ 41 ka (NGRIP age)

990

Geomagnetic field intensity variations are reconstructed from ^{10}Be production rates

991

The dipole low of Laschamp excursion is confirmed at 41 ka.

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