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**Abstract**

Authigenic $^{10}$Be/$^{9}$Be ratio measurements were performed at high resolution along a Portuguese Margin deep-sea core (37°48 N; 10°09 W) spanning the 20-50 ka time interval, in order to reconstruct variations in atmospheric cosmogenic $^{10}$Be production rates and derive the related geomagnetic dipole moment modulation. A complementary approach consisting in $^{230}Th_{xs}$-normalized $^{10}$Be deposition rate determination on selected samples confirmed the reliability of the authigenic $^{10}$Be/$^{9}$Be record. This study constitutes the first successful comparison of the two widely-used normalization techniques of $^{10}$Be concentrations. For both methods, the presented results herein evidence a factor of ~2 cosmogenic nuclide overproduction linked to a minimum dipole moment associated with the Laschamp Excursion. The latter is stratigraphically constrained beneath Heinrich Event 4. Its age is estimated at ~41 ka on the basis of direct correlation between the series of rapid paleoclimatic events recorded in the Portuguese Margin sediments and in the Greenland ice sheet, and is confirmed by calibrated radiocarbon dating carried out on the same sediments.

The remarkable agreement between the authigenic $^{10}$Be/$^{9}$Be and the Greenland Ice cores $^{10}$Be deposition rate records attest to their global significance. This new authigenic $^{10}$Be/$^{9}$Be record has been combined with that previously obtained at the same site to produce a stacked record that is calibrated using absolute values of Virtual Dipole Moment determined on lava flows. This provides a reconstruction of dipole geomagnetic moment variations over the 20-50 ka interval, independent from paleomagnetically-constrained methods, that documents the Laschamp dipole low but fails to express any dipole low related to the Mono Lake Excursion.

This high resolution record responds to the necessity of supplementing the knowledge of the atmospheric $\Delta^{14}$C variations in the 30-45 ka interval during which the $^{14}$C calibration curve suffers from a lack of accurate data, and during which a discrepancy of about 5500 yr

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

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

## 1. Introduction

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

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

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

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

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

## <u>2. Material and methods</u>

### *2.1 Coring site and sediment description*

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

Using different methodologies, several studies have demonstrated that the upper part of the giant cores collected with the Calypso corer of the R/V *Marion Dufresne* are affected by significant oversampling (e.g. Thouveny et al., 2000; Skinner and McCave, 2003 and Szérémeta et al., 2004). This artefact has led to overestimations of apparent sedimentation rates (e.g. Thompson et al., 1999). Despite their perfect stratigraphic preservation, the upper 12 m of core MD95-2042 are affected by this artefact, which hampered the evaluation of accurate sedimentation rates and the interpretation of paleomagnetic and cosmogenic nuclides results.

During the MD-140 PRIVILEGE campaign of the R/V *Marion Dufresne* (2004) on the Portuguese Margin, a giant box corer CASQ ($0.25m^2$x12m) allowed to collect a high resolution, undisturbed sediment sequence (Core MD04-2811; Lat.: 37°48 N; Long.: 10°09 W; 3162m water-depth) at the site of the previously studied SU81-18 and MD95-2042 cores (Fig. 1; see references in caption). These cores were extensively studied for radiocarbon calibration (section 3.2) and therefore benefits from an excellent chronological control. Core MD04-2811 was thus selected in order to document a lack of the MD95-2042 record in the critical time interval (20-35 ka) and to assess accurate sediment fluxes using the $^{230}Th_{xs}$ normalization.

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

### *2.2 Paleomagnetic record and sampling strategy*

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

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

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

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

### *2.3 Analytical techniques*

*Leaching procedure: dissolution of the authigenic phase.*

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

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

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

*Beryllium isotopes chemical extraction and measurements.*

Once the sample is in solution, whatever the method used, Beryllium is then extracted in the form of Be acetylacetonate using organic solvent. After decomposition of these complexes, Be oxy-hydroxides are precipitated at pH 8 and oxidized to perform AMS measurements at ASTER, the French AMS national facility (installed at the CEREGE) which operates at 5MV. $^{10}Be$ concentrations are deduced from spiked $^{10}Be/^{9}Be$ measured ratios calibrated against the NIST 4325 Standard Reference Material with an assigned $^{10}Be/^{9}Be$ value of $2.79 \times 10^{-11}$ (Nishiizumi et al., 2007), and are decay-corrected using the $^{10}Be$ half-life of 1.387 ± 0.012 Ma (Chmeleff et al., 2010; Korschinek et al., 2010):

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

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

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

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

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

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

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

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

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

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

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

To calculate excess $^{230}Th$ activity ($^{230}Th_{xs}$), measured $^{230}Th$ concentration is corrected for detrital and authigenic $^{230}Th$ components. The fraction of $^{230}Th$ supported by the decay of detrital $^{238}U$ is estimated from the $^{232}Th$ sample concentration and an assumed average lithogenic $^{238}U/^{232}Th$ activity ratio of R = 0.6 ± 0.1 for the Atlantic Basin (Henderson and Anderson, 2003). Ingrowth of $^{230}Th$ from decay of authigenic uranium is estimated by assuming that authigenic uranium has an initial $^{234}U/^{238}U$ activity ratio equivalent to that of sea-water (i.e. 1.147, Andersen et al., 2010). $^{230}Th_{xs}$ activities were calculated as:

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

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

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

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

### *2.4 Using authigenic $^{10}Be/^{9}Be$ and $^{230}Th_{xs}$-normalized $^{10}Be$ fluxes as tracers of geomagnetic moment variations*

The particle-reactive $^{10}Be$ is mainly produced in the atmosphere through nuclear reactions (spallation reactions) on oxygen (O) and nitrogen (N), and is transferred to the Earth's surface in soluble form by precipitation (Raisbeck et al., 1981) within ~3 yrs (Baroni et al., 2011). Atmospheric $^{10}Be$ is ultimately removed from water via settling particles and is either deposited in marine and lacustrine sediments, or is efficiently retained by continental sediment components, and finally enters the oceans through various paths and deposits in deep sea sediments. Due to its short atmospheric residence time and chemical reactivity, $^{10}Be$ records of production enhancements should be synchronous in the geophysical reservoirs. In marine sediments, the $^{10}Be$ concentration results from a complex interplay between several processes: e.g. cosmogenic production, redistribution by atmospheric, riverine and oceanic transport, adsorption and deposition processes. Considering the adsorption processes, for example, the absolute $^{10}Be$ concentration is controlled by the scavenging efficiency and the specific surface of the settling sedimentary particles and thus depends on environmental conditions affecting their chemical and grain size composition. Consequently, $^{10}Be$ concentrations are meaningless and in order to account for these dependencies, a correction procedure is required. Soluble forms of $^{10}Be$ and $^{9}Be$ have different sources: while $^{10}Be$ is cosmogenic, the stable isotope $^{9}Be$ originates from partial dissolution of detrital, aeolian and riverine inputs (Brown et al., 1992). Once homogenized in seawater, both isotopes are scavenged with the same efficiency. Authigenic $^{10}Be/^{9}Be$ ratio of marine surface sediments reflects the Be isotope composition of the overlying deep waters, and its spatial variability is mainly controlled by the proximity to continental $^{9}Be$ inputs (Bourlès et al., 1989). Although the authigenic $^{9}Be$ normalization method has provided promising results in specific environments (Bourlès et al., 1989; Carcaillet et al., 2003; Carcaillet et al., 2004a; Carcaillet et al., 2004b; Lebatard et al., 2010), quantitative reconstructions of the $^{10}Be$ fluxes to the sediment requires taking into account syndepositional lateral transport of adsorbed Be. Because residence time of Be in the water column is about 500-1000 yrs, boundary scavenging and deepwater circulation must also contribute to the removal of dissolved Be and thus influence the $^{10}Be/^{9}Be$ ratio. Moreover, lateral transport of adsorbed Be can be significant in sediments heavily affected by focusing, such as drift deposits. However, at ocean margins where particle flux is high, the short Be residence time allows to record the $^{10}Be$ atmospheric flux variations with almost no signal attenuation.

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

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

## 3. Results

### 3.1 The authigenic $^{10}Be/^{9}Be$ ratios and RPI on the same sequence

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

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

### 3.2 Chronostratigraphy

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

Two major studies have established radiocarbon dates series on MD95-2042 (Bard et al., 2004; Shackleton et al., 2004, Electronic Appendix, Table A.1) that were compared with ages obtained from correlations with annually laminated Greenland ice cores (GRIP and GISP2). Among the paleoclimatic proxies measured along the MD95-2042 sequence, sea-surface temperature proxies (alkenones and planktonic $\delta^{18}O$) show marked variations which are in robust mutual stratigraphic agreement, and which correlate to Dansgaard-Oeschger cycles and Heinrich Events. For this study, the age control on the core MD95-2042 was updated by tuning the alkenone-derived sea-surface temperature record with the NGRIP oxygen isotopes record using a minimum number of tie points (R= 0.876, Fig. 3 and 4).

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

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

Bulk $^{10}Be$ concentrations obtained after total sample dissolution are comprised between $1.1x10^{-14}$ g/g and $1.8x10^{-14}$ g/g (see Table 2). Total mass flux ranges between 4.79 and 6.09 g/cm$^2$/ka, and is fully compatible with results obtained further northward on the Portuguese Margin by Thomson et al. (1999), who determine a total sediment accumulation flux of ~5 g/cm$^2$/ka during glacial times over the last 140 ka. Preserved vertical $^{10}Be$ deposition rates vary from $3.18x10^9$ to $6.59x10^9$ atoms/cm$^2$/ka (Fig. 5a). The $^{10}Be$ deposition rate downcore evolution is strictly similar to that of the authigenic $^{10}Be/^9Be$ ratios measured on the same samples: the characteristic increase with a maximum at 712 cm depth is indeed reflected in the $^{10}Be$ deposition rates record. The comparison between authigenic $^{10}Be/^9Be$ ratios and $^{230}Th_{xs}$-normalized $^{10}Be$ fluxes is illustrated in Figure 5b. The linear relationship between these two variables is highly significant (R = 0.912), albeit mainly carried by the maximum variability recorded during the main enhancement.

## 4. Discussion

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

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

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

Along core MD04-2811 the highest $^{10}Be/^{9}Be$ ratio peak - recorded ~40 cm beneath the MS peak corresponding to H4 - can thus undoubtedly be interpreted as the $^{10}Be$ atmospheric overproduction due to the geomagnetic dipole reduction associated with the Laschamp Excursion (LE). This 40 cm depth interval suggests that the LE occurred ~2 ka before Heinrich Event 4, which is dated at ~39 ka in North Atlantic sedimentary sequences (e.g. Thouveny et al., 2000). Indeed, the chronological reconstruction positions the $^{10}Be/^{9}Be$ Laschamp signature at 41.2 ± 1.6 ka (minimal uncertainty derived from GICC05 chronology). This age is coherent with the most recent independent radiometric age determinations obtained on volcanic material (40.4 ± 2.0 ka, Guillou et al., 2004; 40.7 ± 0.9 ka, Singer et al., 2009). It also coincides tightly with the age of the maximum $^{10}$Be flux in Summit ice cores (Muscheler et al., 2005; Svensson et al., 2008). This suggests that the $^{10}$Be residence time in the water column in such settings is short enough to preserve the simultaneity of the record in both geological reservoirs, considering age uncertainties and resolution.

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

### *4.2 Significance of the $^{10}$Be/$^{9}$Be record*

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

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

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

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

Lal's algorithm (1988) has been used to convert the authigenic $^{10}$Be/$^{9}$Be record into a relative Virtual Dipole Moment (VDM) record that is compared to the GLOPIS-75 stacked paleomagnetic record (Fig. 6c). This reconstruction and the GLOPIS-75 curve agree relatively well over the 30-48 ka time interval, although the cosmogenic record points out possible higher intensities for the 20-30 ka period and beyond 48 ka than indicated in the paleomagnetic records of the GLOPIS-75 stack.

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

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

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

*Calibrating to virtual dipole moments*

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

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

## **5. Conclusion**

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

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

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

The authigenic $^{10}Be/^{9}Be$ ratio records of neighboring cores MD04-2811 and MD95-2042 cores were compiled and averaged over a 1 ka sliding window. The obtained stack was calibrated using absolute virtual dipole moment (VDM) values, determined from lava flow paleointensities, to provide a $^{10}Be$ derived VDM record. This allows to characterize the Laschamp dipole low by a decrease from ~$10.5x10^{22}$ $A.m^2$ at ~49 ka to ~$2.6x10^{22}$ $A.m^2$ at ~41 ka. The latter estimate is concordant with radiometric ages obtained on excursional lava flows. This record fails to document a significant VDM reduction at the age of the Mono Lake Excursion (32-34 ka).

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

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# Figure Captions :

**"The Laschamp geomagnetic dipole low expressed as a cosmogenic $^{10}Be$ atmospheric overproduction at ~41 ka."**

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

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

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

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

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

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

**Fig. 6.** (a and b) Relative temporal variations of $^{10}Be$ atmospheric production rate proxies over the 20-50 ka interval. All records are normalized to their respective mean values over the studied time interval. (a) Black dots represent the authigenic $^{10}Be/^{9}Be$ record and the associated 1σ uncertainty. Grey dots represent the $^{230}Th_{xs}$-normalized $^{10}Be$ fluxes. (b) Black dots represent the authigenic $^{10}Be/^{9}Be$ record and the associated 1σ uncertainty, and grey line represents the Greenland $^{10}Be$ flux in Summit ice cores on GICC05 time scale (Finkel et al., 1997; Yiou et al., 1997; composite GRIP-GISP2: Muscheler et al., 2005). (c) Relative temporal variations of geomagnetic field intensity over the 20-50 ka interval. Solid lines represents the 1σ uncertainty envelop of the GLOPIS-75 stack: raw data from Laj et al., (2004) are normalized to $8x10^{22}$ $A.m^2$ (most recent GLOPIS VADM value) and plotted on the GICC05 time scale using the NGRIP, GRIP and GISP2 synchronization match points dataset (Rasmussen et al., 2006). Dots are the authigenic $^{10}Be/^{9}Be$-derived VDM record ($M/M_0$) calculated using the Lal's algorithm (1988): $P/P_0 = (M/M_0)^{-0.5}$, where $P/P_0$ is the authigenic $^{10}Be/^{9}Be$ ratio divided by the surface sample value of $(1.36±0.35)x10^{-8}$, undistinguishable within uncertainties from the long-term average $(1.44 \pm 0.043)x10^{-8}$.

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

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

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

Highligths to the manuscript: "The Laschamp geomagnetic dipole low expressed as a cosmogenic $^{10}Be$ atmospheric overproduction at ~41 ka."

by L. Ménabréaz *, N. Thouveny, D.L. Bourlès, P. Deschamps, B. Hamelin and F. Demory.
*Earth and Planetary Science Letters*

- Cosmogenic $^{10}Be$ concentration was measured along a marine sediment core (20 -50ka)
- Authigenic $^{9}Be$ and excess $^{230}Th$ were used to normalize $^{10}Be$ concentrations
- A doubling of the cosmogenic production is revealed at ~41ka (NGRIP age)
- Geomagnetic field intensity variations are reconstructed from $^{10}Be$ production rates
- The dipole low of Laschamp excursion is confirmed at 41 ka.