Pleistocene variations of beryllium isotopes in central Arctic Ocean sediment cores

Emma Sellén a,⁎, Martin Jakobsson a, Martin Frank b, Peter W. Kubik c

a Department of Geology and Geochemistry, Stockholm University, 106 91 Stockholm, Sweden
b IFM-GEOMAR, Leibniz Institute of Marine Sciences, Wischhofstrasse 1-3, 24148 Kiel, Germany
c Paul Scherrer Institute, c/o Institute for Particle Physics, ETH-Zürich, CH-8093 Zürich, Switzerland

A R T I C L E   I N F O

Article History:
Accepted 28 March 2009
Available online 23 April 2009

Keywords:
beryllium isotopes
Arctic Ocean sedimentation rates
fluxes

A B S T R A C T

Neogene marine sediments can be dated via decay of the cosmogenic radionuclide 10Be. Two cores from the Alpha and Mendeleev Ridges in the Arctic Ocean have been analyzed for seawater-derived beryllium (Be) isotopes in order to date the sediments and to calculate sedimentation rates. The decrease of 10Be concentration in the cores was used to calculate first order sedimentation rates. To eliminate the dilution effect of beryllium caused by short-term changes in sedimentation rate and grain size, the 10Be concentrations were normalized to the terrigenous stable isotope 9Be determined in the same sample aliquot. The measured 10Be concentrations yield low average sedimentation rates for the Alpha and Mendeleev Ridges of 2.3 mm ka−1 and 2.7 mm ka−1, respectively. Sedimentation rates calculated from the 10Be/9Be ratios result in similarly low values, ranging from 0.2 to 6.8 mm ka−1 for the Alpha Ridge core and from 1.9 to 6.9 mm ka−1 for the Mendeleev Ridge core. However, amino acid racemization dating for the past 150 ka of a core adjacent to the Mendeleev Ridge core studied here indicates significantly higher sedimentation rates than calculated from the downcore decrease of 9Be and 10Be/9Be. If such higher rates also prevailed at the locations of our cores, for which there is biostratigraphic evidence, either the supply of 10Be was much lower than assumed or that of 9Be was much higher. This could imply that the signature of the deep waters in this part of the Arctic Ocean compared to today was largely different for most of the past approximately one million years with a significantly lower 10Be/9Be ratio. Our study also addresses the variability of beryllium isotopes in sediment cores across the Arctic Ocean through a comparison of previously published results. Calculated 10Be fluxes reveal low values in the Eurasian Basin and highest values in the Eurasian Basin, near the Fram Strait. The decrease of Be isotopes in the two studied Amerasian Basin cores may thus have been caused by environmental factors such as significantly reduced inflow of Atlantic waters in the past, reduced input of 9Be and/or increased input of 10Be from the shelves, combined with a more efficient sea ice shielding in this part of the Arctic Ocean.

© 2009 Elsevier B.V. All rights reserved.

1. Introduction

The cosmogenic radionuclide 10Be has been used as a dating tool in chronostratigraphic studies of Neogene marine sediments (Amin et al., 1975; Bourlès et al., 1989; Aldahan and Possnert, 2000; Frank et al., 2008). 10Be is produced in the upper atmosphere through cosmic ray-induced spallation of oxygen and nitrogen atoms. The half-life of 10Be has been determined to 1.51±0.06 Ma (Hofmann et al., 1987), but there is an ongoing debate on the correct value. Recent suggestions are ranging from 1.34 to 1.36 Ma (Fink and Smith, 2007; Nishizumi et al., 2007). There are, however, some inherent problems with these recent results that need to be resolved (Fink and Smith, 2007) before there will be a generally agreed revision of the half-life of 10Be. After a short residence time of about one year in the stratosphere 10Be is transferred to the troposphere from where it is quickly removed to the Earth’s surface by precipitation. The global average atmospheric production rate of 10Be has been estimated to 1.21±0.26×106 atoms cm−2 year−1 from precipitation collections (Monaghan et al., 1985/86). For the purpose of this study the production rate of 10Be can be considered constant for the late Holocene but it has mostly been higher by on average 30%, sometimes even up to 70%, during the rest of the Pleistocene as a function of variations in the strength of the Earth’s magnetic field (c.f. Frank et al., 1997). These variations need to be taken into account when applying 10Be for dating Neogene marine sediments and for the reconstruction of past variations of 9Be isotope in seawater.

Two different approaches to derive age constraints based on 10Be have been applied in previous studies of Arctic Ocean cores. The first assumes that secular variations of 10Be concentrations with depth in a core broadly represent glacial–interglacial cycles in that glacial periods are represented by low 10Be concentrations caused by a combination of increased Arctic Ocean sea ice cover and dilution caused by high accumulation rates of ice rafted debris (IRD). This secular glacial–interglacial variation permits the assignment of Marine Isotope Stages (MIS) to the downcore 10Be record. The second
approach uses the downcore decrease in $^{10}$Be concentration to calculate average sedimentation rates. This assumes a near constant supply of $^{10}$Be through time and a consistent initial concentration at the sediment surface. In other words, the downcore decrease in $^{10}$Be concentrations must be caused by radioactive decay only. For this approach, $^{10}$Be is often normalized to seawater-derived stable $^9$Be, which is supplied from the continents and in seawater behaves chemically identical to $^{10}$Be thus allowing elimination of disturbing dilution effects from short-term changes in sedimentation rates. The accuracy of this normalization method is dependent on a constant initial sea water $^{10}$Be/$^9$Be, which may in fact change as a function of circulation patterns given the oceanic residence time of Be of 500–1000 years (Kusakabe et al., 1987; Kusakabe et al., 1990; Measures et al., 1996).

The former method has been applied on some Arctic Ocean short piston cores (<10 m core length) primarily covering the Pleistocene (Eisenhauer et al., 1994; Aldahan et al., 1997; Spielhagen et al., 1997; Spielhagen et al., 2004), whereas the latter provided the main chronostratigraphic tool for the Neogene part of the IODP Leg 302, the ACEX, drill core from the Lomonosov Ridge in the central Arctic Ocean (Backman et al., 2008; Frank et al., 2008).

Here we present the first study of $^{10}$Be in cores retrieved from the Alpha and Mendeleev Ridges in the Amerasian Basin of the Arctic Ocean (Fig. 1). The cores were acquired during the Healy–Oden Trans-Arctic Expedition (HOTRAX) in 2005 (Darby et al., 2005). The purpose of this study is to investigate the possibility to use $^{10}$Be for dating Quaternary sediment cores from the Amerasian Basin and to address the variability of $^{10}$Be concentrations and fluxes in sediments across the Arctic Ocean. Therefore, we also include previously published results from beryllium isotope studies of cores from the Eurasian part of the Arctic Ocean including the Lomonosov Ridge, the Fram Strait and the Norwegian Sea (Fig. 1).

2. Material and methods

The two cores measured for $^{10}$Be in this study are HLY0503-09JPC and HLY0503-14JPC, hereafter referred to as 09JPC and 14JPC. The cores were retrieved from topographic highs on the Mendeleev Ridge and the Alpha Ridge, respectively, during HOTRAX 2005 with the Jumbo Piston Corer system on USCGC Healy (Darby et al., 2005) (Fig. 1). The results are compared with published results from previous studies on beryllium isotopes in Arctic Ocean cores (Fig. 1; Table 1).

2.1. Sampling and Be isotope measurements of the HOTRAX cores

The lithostratigraphies for both 09JPC and 14JPC are characterized by a cyclic alternation of dark brown and lighter colored yellowish brown layers. These sediment cycles have been interpreted to reflect interglacial and glacial variability and have been observed throughout the Arctic Ocean (e.g. Poore et al., 1993; Phillips, 1997; Jakobsson et al., 2000; Polyak et al., 2004; O’Regan et al., 2008; Adler et al., 2009—this volume). The samples for beryllium isotope analyses from 09JPC and 14JPC were therefore systematically taken from typical dark brown sediment units interpreted to represent interglacial periods. The reason for this was to avoid the influence of secular glacial–interglacial variations in the $^{10}$Be record given that our initial primary intention was to calculate sedimentation rates using the decay of $^{10}$Be. A total of 11 samples were taken from core 09JPC, and 20 samples from core 14JPC. All samples consisted of 1-cm thick slices and contained approximately 5 cc of sediment. The samples were dried, homogenized, and subjected to an established weak leaching procedure involving hydroxylamine hydrochloride, which was developed to extract seawater-derived trace metal isotope compositions, such as those of Nd or Pb (Bayon et al., 2002; Gutjahr et al., 2007). A similar
A procedure has been developed to extract the seawater $^{10}\text{Be}$/${}^{9}\text{Be}$ from sediments in the North Atlantic Ocean (Robinson et al., 1995; Carcaillet et al., 2004; Leduc et al., 2006). This procedure is supposed to selectively dissolve the authigenic (bottom water water-derived) trace metal fraction from marine sediments located in thin early diagenetic ferromanganese coatings. For the samples of this study we used the method of Gutjahr et al. (2007) as described in Knudsen et al. (2008), where the method was used successfully to extract seawater-derived $^{10}\text{Be}$/${}^{9}\text{Be}$ in the North Atlantic Ocean.

In order to measure the $^{10}\text{Be}$ concentration by isotope dilution, a precisely known amount of a $^{9}\text{Be}$ carrier (about 1 mg $^{9}\text{Be}$) was added to a weighed aliquot of the leached sediment sample. The leached samples were then further purified and prepared for AMS analyses following a previously available method (Henken-Mellies et al., 1990; Frank et al., 1994). The $^{10}\text{Be}$/${}^{9}\text{Be}$ concentrations were measured at the Zürich AMS facility of the Paul Scherrer Institute and the ETH Zürich, Switzerland. The $^{10}\text{Be}$/${}^{9}\text{Be}$ ratios that were measured at the accelerator for determination of the $^{10}\text{Be}$ concentrations by isotope dilution were normalized to the internal AMS standard S555 with a nominal $^{10}\text{Be}$/${}^{9}\text{Be}$ ratio of 95.5 $\times 10^{-12}$. The 1 sigma uncertainties of individual $^{10}\text{Be}$ measurements take into account both the counting statistics of the $^{10}\text{Be}$ “events” and the reproducibility of repeated measurements, which were performed for each sample.

The natural authigenic $^{9}\text{Be}$ data were acquired at the ICPMS-Laboratory of the Institute of Geosciences, University of Kiel, using an AGILENT 7500cs ICP-MS instrument following established standard procedures (Garbe-Schönberg, 1993). The measurements were performed on an aliquot of the same solution as used for $^{10}\text{Be}$ measurement in order to guarantee that the determined natural $^{10}\text{Be}$/${}^{9}\text{Be}$ represents a true seawater ratio. Repeat measurements of samples processed separately through the leaching procedure indicate a reproducibility within 5%.

### 2.2. Measurements in previous studies

The upper 152 m of the ACEX sediment cores were sampled at 2–10 m intervals for beryllium isotope analyses. The samples were dried, homogenized and leached with 6 M HCl in order to extract the seawater derived $^{10}\text{Be}$ from the sediments (Henken-Mellies et al., 1990; Frank et al., 2008). These samples were also analyzed for the stable isotope $^{10}\text{Be}$ using the leaching procedure of Gutjahr et al. (2007) but the $^{9}\text{Be}$ and $^{10}\text{Be}$ concentrations were not acquired using the same leaching method. Given that the 6 M HCl leach is more aggressive than the hydroxylamine hydrochloride leach, the adsorbed $^{10}\text{Be}$ is almost extracted quantitatively, whereas this is clearly not the case for the weaker leaching method that aims to recover the correct seawater $^{10}\text{Be}$/${}^{9}\text{Be}$ and not the quantitatively correct $^{10}\text{Be}$ concentration. The $^{10}\text{Be}$/${}^{9}\text{Be}$ ratios in Frank et al. (2008) were therefore suitable for dating the ACEX core but, as outlined in the publication, do not represent a true seawater $^{10}\text{Be}$/${}^{9}\text{Be}$ ratio and cannot be compared directly with the $^{10}\text{Be}$/${}^{9}\text{Be}$ results of this study.

Cores PS2178, PS2185 and PS2200 were measured for $^{10}\text{Be}$ concentrations at the AMS facility of ETH Zürich following a 6 M HCl leaching of the sediments (Spieglmann et al., 1997; Spieglmann et al., 2004). $^{9}\text{Be}$ was not included in these studies. The sediments of

---

**Table 1**

<table>
<thead>
<tr>
<th>Core name</th>
<th>Position</th>
<th>Water depth (m)</th>
<th>Core length (m)</th>
<th>Location</th>
<th>Number of $^{10}\text{Be}$-samples</th>
<th>Sample interval</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>09JPC</td>
<td>172.2766 W, 79°35.605 N</td>
<td>2783</td>
<td>15.39</td>
<td>Mendeleev Ridge</td>
<td>11³</td>
<td>0.5–2.5 m</td>
<td>Darby et al. (2005)</td>
</tr>
<tr>
<td>14JPC</td>
<td>149°2.041 W, 84°18.190 N</td>
<td>1856</td>
<td>11.22</td>
<td>Alpha Ridge</td>
<td>20³</td>
<td>App. 60 cm</td>
<td>Darby et al. (2005)</td>
</tr>
<tr>
<td>ACEX 2A</td>
<td>131°21.901 E, 87°35.271 N</td>
<td>1209</td>
<td>27.69</td>
<td>Lomonosov Ridge</td>
<td>25³</td>
<td>2–4 m</td>
<td>Frank et al. (2008)</td>
</tr>
<tr>
<td>ACEX 4C</td>
<td>130°11.381 E, 87°52.055 N</td>
<td>1288</td>
<td>40.19</td>
<td>Lomonosov Ridge</td>
<td>6⁴</td>
<td>App. 5 m</td>
<td>Darby et al. (2005)</td>
</tr>
<tr>
<td>96/09-1pc</td>
<td>143°26.62 E, 80°24.87 N</td>
<td>927</td>
<td>2.70</td>
<td>Lomonosov Ridge</td>
<td>9⁵</td>
<td>10–20 cm</td>
<td>Backman et al. (1997)</td>
</tr>
<tr>
<td>PS2185 (BC and KAL)</td>
<td>145°55.6 E, 87°32.2 N</td>
<td>1051</td>
<td>7.68</td>
<td>Lomonosov Ridge</td>
<td>92</td>
<td>5 cm</td>
<td>Spieglmann et al. (1997)</td>
</tr>
<tr>
<td>PS2178 (BC and KAL)</td>
<td>159°42.6 E, 88°1.5 N</td>
<td>4008</td>
<td>8.30</td>
<td>Makarov Basin</td>
<td>64</td>
<td>5 cm</td>
<td>Spieglmann et al. (2004)</td>
</tr>
<tr>
<td>PS2200 (BC and KAL)</td>
<td>14°00.0 W, 85°19.0 N</td>
<td>1074</td>
<td>7.7</td>
<td>Morris Jesup Rise</td>
<td>106</td>
<td>5 cm</td>
<td>Spieglmann et al. (2004)</td>
</tr>
<tr>
<td>PS2220 (BC and KAL)</td>
<td>4°39.5 E, 83°38.6 N</td>
<td>3682</td>
<td>6.8</td>
<td>Nansen Basin</td>
<td>55</td>
<td>5–20 cm</td>
<td>Aldahan et al. (1997)</td>
</tr>
<tr>
<td>PS2213 (BC and PC)</td>
<td>8°2.6 E, 80°27.6 N</td>
<td>853</td>
<td>13.0</td>
<td>Yermak Plateau</td>
<td>79</td>
<td>5–20 cm</td>
<td>Aldahan et al. (1997)</td>
</tr>
<tr>
<td>23059</td>
<td>41°13.9 W, 70°18.3 N</td>
<td>2281</td>
<td>5.22</td>
<td>Norwegian Sea</td>
<td>52</td>
<td>5–10 cm</td>
<td>Eisenhauer et al. (1994)</td>
</tr>
<tr>
<td>23235</td>
<td>1°18.5 E, 78°51.5 N</td>
<td>2456</td>
<td>8.06</td>
<td>Fram Strait</td>
<td>80</td>
<td>5–10 cm</td>
<td>Eisenhauer et al. (1994)</td>
</tr>
<tr>
<td>PS1524</td>
<td>26°12.9 E, 85°21.8 N</td>
<td>3646</td>
<td>4.2</td>
<td>Gakkel Ridge</td>
<td>33</td>
<td>5–10 cm</td>
<td>Eisenhauer et al. (1994)</td>
</tr>
<tr>
<td>PS1533</td>
<td>15°10.7 E, 82°19.5 N</td>
<td>2030</td>
<td>4.9</td>
<td>Yermak Plateau</td>
<td>48</td>
<td>5–10 cm</td>
<td>Eisenhauer et al. (1994)</td>
</tr>
</tbody>
</table>

---

**Table 2**

<table>
<thead>
<tr>
<th>Core name</th>
<th>Methods used for age determination</th>
<th>Average sedimentation rates (cm ka$^{-1}$)</th>
<th>Reference(s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>09JPC</td>
<td>Nanofossils and amino acid racemization correlated from 08JPC</td>
<td>1.53</td>
<td>Backman and Fornaciari (pers. comm.) and Kaufman et al. (2008)</td>
</tr>
<tr>
<td>14JPC</td>
<td>Presence of nanofossil <em>Emiliana huxleyi</em></td>
<td>1.96</td>
<td>Backman and Fornaciari (pers. comm.)</td>
</tr>
<tr>
<td>ACEX 2A/4C</td>
<td>Beryllium-dating</td>
<td>1.45</td>
<td>Frank et al. (2008)</td>
</tr>
<tr>
<td>96/09-1pc</td>
<td>Stratigraphic correlation with 96/12-1pc</td>
<td>1.23</td>
<td>Jakobsson et al. (2001)</td>
</tr>
<tr>
<td>2185</td>
<td>Carbon and oxygen isotopes, paleomagnetic inclination, nanofossils, planktic foraminifera, $^{14}$C, coarse fraction, $^{10}\text{Be}$</td>
<td>1.55</td>
<td>Spieglmann et al. (2004)</td>
</tr>
<tr>
<td>2178</td>
<td>$^{10}\text{Be}$, coarse fraction, paleomagnetic inclination, nanofossils, planktic foraminifera, $^{14}$C, coarse fraction, $^{10}\text{Be}$</td>
<td>2.55</td>
<td>Spieglmann et al. (2004)</td>
</tr>
<tr>
<td>2200</td>
<td>Carbon and oxygen isotopes, paleomagnetic inclination, $^{14}$C, coarse fraction, $^{10}\text{Be}$, planktic foraminifera</td>
<td>1.43</td>
<td>Spieglmann et al. (2004)</td>
</tr>
<tr>
<td>2208</td>
<td>$^{10}\text{Be}$ compared to oxygen isotope profile, paleomagnetic record, nanofossils, microfossils, $^{14}$C</td>
<td>1.92</td>
<td>Aldahan et al. (1997)</td>
</tr>
<tr>
<td>2213</td>
<td>$^{10}\text{Be}$ compared to oxygen isotope profile, paleomagnetic record, planktic foraminifera, $^{14}$C, coarse fraction, $^{10}\text{Be}$, planktic foraminifera</td>
<td>2.98</td>
<td>Aldahan et al. (1997)</td>
</tr>
<tr>
<td>23059</td>
<td>Oxygen isotopes</td>
<td>1.8</td>
<td>Eisenhauer et al. (1994)</td>
</tr>
<tr>
<td>23235</td>
<td>Nanofossils, $^{230}$Th</td>
<td>2.6</td>
<td>Eisenhauer et al. (1994)</td>
</tr>
<tr>
<td>1524</td>
<td>Oxygen isotopes, $^{14}$C, $^{230}$Th</td>
<td>0.96</td>
<td>Eisenhauer et al. (1994)</td>
</tr>
<tr>
<td>1533</td>
<td>Oxygen isotopes, $^{14}$C, $^{10}\text{Be}$, $^{230}$Th</td>
<td>3.1</td>
<td>Eisenhauer et al. (1994)</td>
</tr>
</tbody>
</table>
cores PS1533, PS1524, 23235, and 23059 in the study of Eisenhauer et al. (1994) were prepared following the same 6 M HCl leaching procedures to measure $^{10}$Be concentrations.

The AMS work by Aldahan et al. (1997) on cores PS2208 and PS2213 was carried out at the Uppsala Tandem Laboratory using 200 mg sediment samples and a 200 µg $^9$Be carrier. In their study, the $^{10}$Be was extracted from the samples using a total dissolution (HCl/HF) procedure, which guarantees complete extraction of all $^{10}$Be contained in the samples including potentially in situ produced $^{10}$Be that formed prior to the supply of the sediments to the ocean (for further details on this method see Ning et al. (1994)).

Sedimentation rates for all the previously measured cores included in this study have been calculated independently using a range of different dating methods (Table 2).

3. Results

3.1. Beryllium concentrations, $^{10}$Be/$^9$Be and calculated sedimentation rates based on the radioactive decay of $^{10}$Be in the HOTRAX cores

The $^{10}$Be concentrations measured in core 09JPC (Fig. 2) showed a value of $13.5 \times 10^7$ atoms g$^{-1}$ at the surface and then, with the exception of some elevated values between 1000 and 1200 cm core depth, decrease nearly exponentially to $0.7 \times 10^7$ atoms g$^{-1}$ at 1530 cm core depth. The first approach to obtain sedimentation rates was to apply an exponential regression through all measured values, yielding an average sedimentation rate for core 09JPC of 2.7 mm ka$^{-1}$. The $^{10}$Be concentrations in core 14JPC are generally lower than in 09JPC. The uppermost measured $^{10}$Be concentration is $11.4 \times 10^7$ atoms g$^{-1}$ and the lowest concentration at 1130.5 cm is $0.07 \times 10^7$ atoms g$^{-1}$. Calculating an average sedimentation rate for 14JPC using the same approach, including all values in an exponential regression, results in 2.3 mm ka$^{-1}$, which is very similar to core 09JPC. The scatter around the exponential decrease in core 14JPC is, however, much larger than in 09JPC and does not allow a precise estimate of the sedimentation rate. In order to reduce this scatter and potential variations caused by short term dilution effects due to fluctuations of coarser grained sedimentation (Frank et al., 2008), the $^{10}$Be concentrations were normalized to authigenic $^9$Be obtained from the same leached aliquot (Fig. 3). The resulting near surface $^{10}$Be/$^9$Be in the two cores is somewhat lower than the present day deep water $^{10}$Be/$^9$Be data but within the range of presently observed Arctic water column data (Frank et al., 2009). The regression lines for the entire $^{10}$Be/$^9$Be data sets result in average sedimentation rates of 1.9 mm ka$^{-1}$ for core 09JPC and 1.6 mm ka$^{-1}$ for core 14JPC but the scatter around the exponential regression line is still large and similar to that of the $^{10}$Be concentration profiles. The correlation between the regression fit and the data in core 14JPC does not improve by normalization to authigenic $^9$Be. Given that there is no a priori reason to assume that the sedimentation rates have been constant over time, rather than drawing a single regression line through all of the measurements resulting in an average sedimentation rate for the entire data sets, the data were divided into subsections in order to obtain regressions with better correlation coefficients (Fig. 3). If the overall decrease of the $^{10}$Be/$^9$Be data with depth in core is due to age, this should account for any changes in sedimentation rates and provide a higher resolution age–depth model. Using this approach, the sedimentation rates in core 09JPC range from 1.9 mm ka$^{-1}$ to 6.9 mm ka$^{-1}$ with the highest sedimentation rate occurring between about 700 and 1160 cm. For 14JPC the same method resulted in values ranging between 0.2 mm ka$^{-1}$ and 6.8 mm ka$^{-1}$, with the highest deposition rate found between 125 cm and 677 cm. These results may also indicate a hiatus in core 09JPC somewhere between 1160 cm and 1300 cm although a drop in sedimentation rates to 0.4 mm ka$^{-1}$ may not be unrealistic in view of the variations observed in 14JPC. The oldest sediments, based on these calculations, are approximately 8.2 Ma for core 09JPC and 10.8 Ma for core 14JPC. However, the regression lines are based on a very limited number of measurements which show a significant scatter. This results in relatively large uncertainties of the sedimentation rates, which, however, provide an indication of the possible variability within the cores.

3.2. Variations in surface sediment $^{10}$Be concentrations across the Arctic Ocean to the Norwegian Sea

In order to better constrain the observed downcore variability, we evaluate environmental influences on the seawater Be isotope composition in the Arctic Ocean. A comparison of measured $^{10}$Be concentrations...
from the sediment surfaces in the central Arctic Ocean, the Fram Strait and the Norwegian Sea indicates that there are substantial and systematic variations (Fig. 4). However, a direct comparison between all published 10Be results from Arctic Ocean cores is not possible as different methods have been used to derive 10Be concentrations (see section Measurements in previous studies).

It has been shown that the weak leach method (Gutjahr et al., 2007) provides approximately four times lower 10Be concentrations than obtained by a 6 M HCl leach on the same sediment material from the North Atlantic Ocean (Knudsen et al., 2008). If we assume that this ratio is also applicable to Arctic Ocean sediments and take it into account for the surfaces of cores 09JPC and 14JPC, the 10Be concentrations show significantly lower values in the Amerasian Basin than in the Eurasian Basin. The concentrations in surface sediments on the Lomonosov Ridge are all essentially higher than found in the HOTRAX cores from the Amerasian Basin, with the exception of 96/09-1pc. The Lomonosov Ridge cores in general have lower surface concentrations than observed in cores located further to the south in the Eurasian Basin and the Fram Strait. Core PS2178, retrieved from the Makarov Basin close to the Lomonosov Ridge, has beryllium concentrations in the uppermost sediments that are more comparable to the cores from the ridge crest than the two Amerasian Basin cores, possibly indicating sediment redistribution via slumping or that the sedimentation regime here is more similar to that of the ridge crest. The highest surface concentration of 10Be is found in core PS1533 (1.15×10^9 atoms g^-1) from the Yermak Plateau, which is about an order of magnitude higher than in core 14JPC from the Alpha Ridge.

3.3. Comparison of 10Be fluxes across the Arctic Ocean

A comparison of the geographical variations of 10Be from different sites and sections within the cores can only be made based on 10Be fluxes. In order to calculate reliable fluxes, independent age models derived from methods other than based on decay of 10Be are required (Table 2). For the two HOTRAX cores 09JPC and 14JPC, the highest sedimentation rates obtained from the 10Be/9Be analyses were 6.9 mm ka^-1 for 09JPC and 6.8 mm ka^-1 for 14JPC, which are comparable to the first estimates of Amerasian Basin sedimentation rates by Clark et al. (1980). However, the lowest values of 0.4 mm ka^-1 for 09JPC and 0.2 mm ka^-1 for 14JPC are even lower than those previously suggested sedimentation rates. The mm ka^-1 scale sedimentation rates of Clark et al. (1980) have been questioned as they were solely based on the identification of the Brunhes/Matuyama paleomagnetic inclination reversal (Backman et al., 2004). Several more recent studies have shown that the paleomagnetic inclination change assumed to be the Brunhes/Matuyama in the central Arctic Ocean stratigraphy likely represents a younger excursion event (Løvlie et al., 1986; Bleil and Gard, 1989; Jakobsson et al., 2001; Nowaczyk et al., 2001; Nowaczyk et al., 2003; Spielhagen et al., 2004).

Amino acid racemization dating for the past 150 ka and calcareous nanofossil abundance studies correlated from 08JPC (Fig. 5) provides estimated average sedimentation rates for core 09JPC of 1.5 to 1.6 cm ka^-1 (Kaufman et al., 2008; pers. comm. Backman and Fornaciari). The earliest appearance of E. huxleyi in 14JPC indicates an approximate sedimentation rate of 1.9 cm ka^-1 but only very few nanofossils were found in this core (Backman and Fornaciari, pers. comm.).

Fig. 3. 10Be/9Be ratios and calculated sedimentation rates for cores 09JPC (above, red) and 14JPC (below, blue). Sedimentation rates and ages are based on the slope of the regression lines. The black dashed line indicates a pronounced step in the beryllium values which may represent a hiatus. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Fig. 4. Beryllium surface concentrations and fluxes corrected for 10Be production variations in cores from the central Arctic Ocean, Fram Strait and the Norwegian Sea. The surface concentrations for 09JPC, 14JPC and ACEX2A/4C have been decay corrected. Blue circles mark the lowest values and red circles mark the highest values. Cores 2185 and ACEX 2A/4C do not have surface values but instead the uppermost available value is displayed. No fluxes are available for 23235 and 23059. Light blue lines and blue numbers indicate the sea ice residence time in years (Rigor et al., 2002).
values are one order of magnitude higher than the average rates calculated from the downcore decrease of $^{10}\text{Be}$ and $^{10}\text{Be}/^{9}\text{Be}$ assuming that the decrease was solely caused by radioactive decay. This suggests that environmental factors were responsible for the low downcore values of $^{10}\text{Be}$ concentrations and $^{10}\text{Be}/^{9}\text{Be}$, which will be discussed in the Discussion and conclusion section.

In order to account for differences caused by sedimentation rates, $^{10}\text{Be}$ fluxes were calculated for the sediment surfaces, i.e. Holocene fluxes, and for MIS 2–4, and MIS 5.5, as well as the average fluxes for the entire cores. For data points from the past 200 ka, the average MIS production rate changes provided in Frank et al. (1997) were applied and for the average fluxes of all data of the past 200 ka, a correction factor of 1.22 was applied (Frank et al., 1997). For the average of data older than 200 ka the same correction factor of 1.22 accounting for an overall higher production rate caused by an overall lower magnetic field intensity was applied (Valet et al., 2005). Density values for cores 23235 and 23059 are not available and, thus, fluxes for these two cores were not calculated.

The average $^{10}\text{Be}$ fluxes (Fig. 4) calculated from all the samples in each included core show the lowest value in the Amerasian Basin ($1.5 \times 10^8$ atoms cm$^{-2}$ ka$^{-1}$). Higher fluxes are found towards the Lomonosov Ridge and further to the south, where the highest average flux occurs in core PS1533 ($1.98 \times 10^9$ atoms cm$^{-2}$ ka$^{-1}$) on the Yermak Plateau. However, the average $^{10}\text{Be}$ fluxes do not show a linear increase towards the Fram Strait, but rather a marked step given that all the cores from the Gakkel Ridge, Nansen Basin, Morris Jesup Plateau, Yermak Plateau and the Fram Strait show rather similar $^{10}\text{Be}$ fluxes. Variations of the whole cores’ average $^{10}\text{Be}$ fluxes between coring sites can not be explained by larger amounts of older sediments with very low $^{10}\text{Be}$ concentrations in the deeper parts of some of the cores because for all the studied cores, including the long ACEX drill core, decay corrected initial $^{10}\text{Be}$ concentrations were used to calculate $^{10}\text{Be}$ fluxes.

The $^{10}\text{Be}$ values for MIS 1, 2–4 and 5.5 are available for all cores except 09JPC, for which MIS 2–4 and 5.5 have not been sampled; MIS 2–4 was not sampled in core 96/09-1pc; and the ACEX drill core record does not include data for MIS 1 and 5.5. The average fluxes for the Holocene show the highest value in core PS1533 ($2.7 \times 10^9$ atoms cm$^{-2}$ ka$^{-1}$) from the Yermak Plateau and the lowest $^{10}\text{Be}$ fluxes are found in PS2185 ($2.8 \times 10^8$ atoms cm$^{-2}$ ka$^{-1}$) from the Lomonosov Ridge. The calculated fluxes for MIS 5.5 show the lowest $^{10}\text{Be}$ fluxes in the Amerasian Basin on the Alpha Ridge ($1.1 \times 10^8$ atoms cm$^{-2}$ ka$^{-1}$) and the highest in core PS1533 ($5.1 \times 10^9$ atoms cm$^{-2}$ ka$^{-1}$) from the Yermak Plateau. In general, the lowest values are observed in the Amerasian Basin, the intermediate values in cores from or close to the Lomonosov Ridge, with the highest values in the Eurasian Basin for all intervals. An exception is the relatively low fluxes calculated for core PS1524 on the Gakkel Ridge when compared to the other Eurasian Basin cores. The cores PS2178, PS2208 and PS1524 are located at high water depths (Table 1) which may suggest that some of them have been subject to sediment redistribution which may have resulted in increased sedimentation rates and thus increased $^{10}\text{Be}$ fluxes.

4. Discussion and conclusion

4.1. Environmental influences on $^{10}\text{Be}$ concentrations and seawater $^{10}\text{Be}/^{9}\text{Be}$

The large discrepancy between sedimentation rates derived from the amino acid racemization, the nannofossil abundance study and apparent radioactive decay of $^{10}\text{Be}$ in cores 09JPC (see Fig. 5) and 14JPC requires an investigation of environmental factors that are superimposed on the radioactive decay. Such potential factors include variations in production of $^{10}\text{Be}$ in the atmosphere or changes in transport of this cosmogenic nuclide from the atmosphere to the oceans. Previous studies revealed prominent secular variations with high $^{10}\text{Be}$ concentrations during interglacials and low ones during

Fig. 5. Correlation between magnetic susceptibility records from HOTRAX cores 08JPC and 09JPC from the Mendeleev Ridge and a comparison between age model for 09JPC based on amino acid racemization (Kaufman et al., 2008) and the estimated average sedimentation rate for the interval based on $^{10}\text{Be}/^{9}\text{Be}$ (dashed line).
Variations in atmospheric production related to changes in geomagnetic field intensity have to be taken into account for the reconstruction of secular variations of $^{10}\text{Be}$ from marine sediments. The maximum effect of changes in production rate over the past about 300 ka is a near doubling of the $^{10}\text{Be}$ production rate during brief geomagnetic excursions such as the Laschamps or Iceland Basin events (Frank et al., 1997; Carcaillet et al., 2004; Knudsen et al., 2008). Overall the Earth's magnetic field was weaker by on average 30% for most of the past 2 Ma (Valet et al., 2005) which resulted in ~22% higher production rates of $^{10}\text{Be}$. Correction for this variation in production rates even enhances the minimum $^{10}\text{Be}$ concentrations and fluxes, as well as $^{10}\text{Be}/^{9}\text{Be}$, that we observe in the older parts of cores 09JPC and 14JPC (see Fig. 6). Changes in $^{10}\text{Be}$ production rate were also taken into account for the calculation of the average $^{10}\text{Be}$ fluxes of all cores. The spatial distribution of $^{10}\text{Be}$ surface concentrations and fluxes of the Arctic Ocean measured in sediment cores show that the Amerasian Basin must have received much lower pre-Holocene $^{10}\text{Be}$ fluxes which correspondingly resulted in lower deep water $^{10}\text{Be}/^{9}\text{Be}$ during the pre-Holocene (Fig. 4). The $^{10}\text{Be}$ surface concentrations and fluxes also become higher towards the Lomonosov Ridge and further south. The observed spatial pattern does not support an influence of atmospheric transport (Fig. 4), although we cannot completely exclude that it may also have played a minor role.

A dilution of $^{10}\text{Be}$ concentrations may occur during glacial periods due to the increased sedimentation of coarser material caused by larger influx of Ice Rafted Debris (IRD) from icebergs (Eisenhauer et al., 1994; Aldahan et al., 1997). However, normalizing to $^{9}\text{Be}$ should to a large extent remove this dilution effect. The most likely reason for the observed minima is thus a significantly reduced influx of Atlantic waters with relatively high $^{10}\text{Be}$ and $^{10}\text{Be}/^{9}\text{Be}$ into the deep Arctic Ocean combined with a reduced input of $^{10}\text{Be}$ from the shelves during the glacial periods. This is consistent with a scenario obtained from Nd isotope data on the Lomonosov Ridge for the past 15 Ma (Haley et al., 2008) and may have had an even stronger impact on the deep Amerasian basin which is further away from and deeper than the inflowing Atlantic waters. Furthermore, continental shelves make up as much as ~53% of the Arctic Ocean area and during glacial periods with more than 100 m lower sea level than at present and large ice sheets occupying some of the continental shelves, the Arctic Ocean was practically without shallow shelves (Jakobsson, 2002) thus inhibiting the supply of $^{10}\text{Be}$ from the surrounding continents. At the same time, terrigenous $^{9}\text{Be}$ was most likely transported to the central Arctic Ocean in larger amounts as a result of glacial erosion, thus enhancing the low $^{10}\text{Be}/^{9}\text{Be}$. However, we do not observe any drastically increased values in the seawater derived $^{9}\text{Be}$ record over the past one million years, which suggests that the low $^{10}\text{Be}/^{9}\text{Be}$ values are most likely a consequence of low $^{10}\text{Be}$ supply.

None of the environmental factors discussed above seems to fully answer the question why there were such pronouncedly low $^{10}\text{Be}$ concentrations and $^{10}\text{Be}/^{9}\text{Be}$ in the sediments of the Amerasian Basin of the Arctic Ocean.

4.2. Effects of sea ice shielding on $^{10}\text{Be}$ concentrations and $^{10}\text{Be}/^{9}\text{Be}$

In the modern deep Arctic Ocean, which is dominated by waters of Atlantic origin, the residence time of deep waters is a few hundreds of years (e.g. Schlosser et al., 1995) which are well homogenized with respect to Be isotopes (Frank et al., 2009). However the low fluxes of $^{10}\text{Be}$ to the seafloor that have been observed for parts of the Arctic Ocean imply that the pre-Holocene conditions must have been quite different.

The presence of sea ice has been suggested to restrict the amount of $^{10}\text{Be}$ that can be transferred into the deep Arctic Ocean (Eisenhauer et al., 1994). Sea ice thickness and residence time varies greatly across the Arctic Ocean (Fig. 4), which may have had an impact on the spatial distribution of $^{10}\text{Be}$ deposition in the deep sea sediments. The shortest sea ice residence time (~1 year) prevails in the Norwegian Sea and the longest residence time (~6 years) is found within the Beaufort Gyre in the Amerasian Basin (Rigor et al., 2002). The whole core average $^{10}\text{Be}$ fluxes show lower values for 14JPC and 09JPC. However, there does not seem to be similarly low values for the average Holocene in the Amerasian Basin, whereas the $^{10}\text{Be}$ fluxes for the colder MIS 2–4 periods show a similar pattern to the whole core average fluxes. If sea ice indeed is the main factor preventing deposition of the cosmogenic isotope of beryllium in parts of the Arctic Ocean our results indicate a more efficient ice shielding in the Amerasian Basin than further south in the pre Holocene Arctic Ocean.

**Fig. 6.** Decay-corrected initial $^{10}\text{Be}/^{9}\text{Be}$ ratios for 09JPC (dots) and 14JPC (open circles) plotted against ages derived from nannofossil biostratigraphy (pers. comm. Backman and Fornaciari).
When comparing our measured fluxes of cosmogenic $^{10}$Be to average sedimentation rates for all cores obtained by methods other than $^{10}$Be-dating, it is apparent that a pronounced drop to mm ka$^{-1}$ scale sedimentation rates in the Amerasian Basin cannot be observed. Therefore, it seems reasonable to suggest that the sea ice and its turnover time have had a strong influence on the $^{10}$Be accumulation in the sediments. If the high independently determined sedimentation rates are correct, it follows that the $^{10}$Be decay cannot be used directly to calculate sedimentation rates in the most severely sea ice covered areas of the Arctic Ocean, such as the Beaufort Gyre, at least for the past 1 Ma. Furthermore, there may have been a complete cutoff from continental $^{10}$Be supply due to a reduced hydrological cycle, as well as a severe restriction of $^{10}$Be supply with water mass input from the Atlantic Ocean during glacial times, which is in accordance with results obtained on the basis of Nd isotope variations (Haley et al., 2008). In addition, the large ice shelves that appeared to have covered the interior parts of the Amerasian Arctic Ocean (Jakobsson et al., 2008b) during glacial times together with a reduced hydrological cycle would effectively enhance the differences with respect to beryllium concentration between the Amerasian Basin cores and cores retrieved in the Eurasian Basin and further towards the Fram Strait. The fact that such an influence of environmental parameters on the $^{10}$Be distribution has not been observed in the shallow ACEX core from Lomonosov Ridge, where the downcore decrease of $^{10}$Be is essentially only a function of radioactive decay (Frank et al., 2008), supports the strong local control of these effects. These conclusions will have to be confirmed by further independent evidence for the stratigraphic framework and the corresponding high sedimentation rates of these cores. If these sedimentation rates will be confirmed, the $^{10}$Be isotope results presented here provide strong evidence for fundamentally different past environmental conditions such as stronger shielding by sea ice and reduced inflow of Atlantic waters.

Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.gloplacha.2009.03.024.

Acknowledgements

Support was received from Svenska Sällskapet för Antropologi och Geografi (SSAG) through a grant financed by the Andrée Foundation. Jutta Heinze is thanked for smooth chemical preparations, Ulrike Westerström at the IFG Kiel is thanked for measurements of the $^{10}$Be concentrations and Fabian Scheifele at ETH Zurich helped with the final preparations of the $^{10}$Be samples for the AMS measurements. The analysis of the $^{10}$Be concentrations was performed at the Zurich AMS Facility jointly operated by the Swiss Federal Institute of Technology, Zurich and Paul Scherrer Institut, Villigen, Switzerland. This is a contribution from the Bert Bolin Centre for Climate Research (BCCB).

References


