The dissolved Beryllium isotope composition of the Arctic Ocean

M. Frank\textsuperscript{a,b,*}, D. Porcelli\textsuperscript{c}, P. Andersson\textsuperscript{d}, M. Baskaran\textsuperscript{e}, G. Björk\textsuperscript{f}, P.W. Kubik\textsuperscript{g}, B. Hattendorf\textsuperscript{h}, D. Guenther\textsuperscript{h}

\textsuperscript{a}Leibniz Institute of Marine Research, IFM-GEOMAR, Wischhofstrasse 1-3, 24148 Kiel, Germany
\textsuperscript{b}Institute for Isotope Geology and Mineral Resources, Department of Earth Sciences, ETH Zurich, CH-8092 Zurich, Switzerland
\textsuperscript{c}University of Oxford, Department of Earth Sciences, Parks Road, Oxford OX1 3PR, UK
\textsuperscript{d}Laboratory for Isotope Geology, Swedish Museum of Natural History, 104 05 Stockholm, Sweden
\textsuperscript{e}Department of Geology, Wayne State University, Detroit, MI 48202, USA
\textsuperscript{f}Department of Oceanography, Earth Science Center, Göteborg University, Box 460, SE-405 30 Göteborg, Sweden
\textsuperscript{g}Paul Scherrer Institute, c/o Institute for Particle Physics, ETH Zu¨rich, CH-8093 Zu¨rich, Switzerland
\textsuperscript{h}Laboratory for Inorganic Chemistry, ETH Hönggerberg, 8093 Zürich, Switzerland

Received 22 December 2008; accepted in revised form 7 July 2009; available online 14 July 2009

Abstract

We present the first comprehensive set of dissolved \(^{10}\)Be and \(^{9}\)Be concentrations in surface waters and vertical profiles of all major sub-basins of the Arctic Ocean, which are complemented by data from the major Arctic rivers Mackenzie, Lena, Yenisey and Ob. The results show that \(^{10}\)Be and \(^{9}\)Be concentrations in waters below 150 m depth are low and only vary within a factor of 2 throughout the Arctic Basin (350–750 atoms/g and 9–15 pmol/kg, respectively). In marked contrast, Be isotope compositions in the upper 150 m are highly variable and show systematic variations. Cosmogenic \(^{10}\)Be concentrations range from 150 to 1000 atoms/g and concentrations of terrigenous \(^{9}\)Be range from 7 to 65 pmol/kg, resulting in \(^{10}\)Be/\(^{9}\)Be ratios (atom/atom) between 0.5 and 14\(^{\text{C}}\). Inflowing Atlantic water masses in the Eurasian Basin are characterized by a \(^{10}\)Be/\(^{9}\)Be signature of \(7^{\text{C}}\). The inflow of Pacific water masses across the Bering Strait is characterized by lower ratios of 2–3 \(^{\text{C}}\), which can be traced into the central Arctic Ocean, possibly as far as the Fram Strait. A comparison of the high dissolved surface \(^{10}\)Be and \(^{9}\)Be concentrations (corresponding to low \(^{10}\)Be/\(^{9}\)Be signatures of \(\sim 2^{\text{C}}\)) in the Eurasian Basin with hydrographic parameters and river data documents efficient and rapid transport of Be with Siberian river waters across the Siberian Arctic shelves into the central Arctic Basin, although significant loss and exchange of Be on the shelves occurs. In contrast, fresh surface waters from the Canada Basin also show high cosmogenic \(^{10}\)Be contents, but are not enriched in terrigenous \(^{9}\)Be (resulting in high \(^{10}\)Be/\(^{9}\)Be signatures of up to \(14^{\text{C}}\)). This is explained by a combination of efficient scavenging of Be in the Mackenzie River estuary and the shelves and additional supply of cosmogenic \(^{10}\)Be via atmospheric fallout and melting of old sea ice. The residence time of Be in the deep Arctic Ocean estimated from our data is 800 years and thus similar to the average Be residence time in the global ocean.

© 2009 Elsevier Ltd. All rights reserved.

1. INTRODUCTION

The Arctic Ocean is a Mediterranean Sea that is highly sensitive to changes in global climate. This is manifested

most obviously in the variability of the extent and thickness of its sea ice cover (Winsor, 2001; Gerdes and Köberle, 2007) and of its freshwater cycle (Peterson et al., 2006). At the same time, the Arctic Ocean may itself exert an important control on the thermohaline circulation system in the North Atlantic through its freshwater export (Aagaard and Carmack, 1994).

The low winter temperatures in the Arctic, in combination with large shallow shelf areas, result in a high potential of deep water formation through direct cooling and brine
injection during sea ice formation (e.g. Aagaard et al., 1985; Bauch et al., 1995). The actual amount of deep water formed is, however, relatively small due to a freshwater lid that originates from inputs of Arctic river water and melting of sea ice (e.g. Aagaard et al., 1985; Anderson et al., 1994). The deeper Arctic Ocean water masses, in particular those of the Canada Basin, are renewed by waters of Atlantic origin through Fram Strait at comparatively low rates (e.g. Schlosser et al., 1995). Surface and shallow water exchange occurs with the Atlantic Ocean through Fram Strait and the Barents Sea and with the Pacific Ocean through the Bering Strait. In addition, there are surface water connections of the Arctic Ocean with the Atlantic and Pacific oceans through the Canadian Archipelago (e.g. Jones et al., 2003).

Many geochemical tracers have been used to investigate upper Arctic water masses, where highly saline influxes from the Pacific and Atlantic oceans mix with freshwaters originating from rivers and sea ice melting at a high spatial and temporal variability. These include dissolved oxygen isotopes (H$_2$O) to distinguish between sea ice meltwater and riverine sources (Bauch et al., 1995; Ekwurzel et al., 2001), nutrients, such as phosphate and nitrate (Wilson and Wallace 1990; Ekwurzel et al., 2001; Jones et al., 2008) and silicate (Bauch et al., 1995), which can be used to distinguish Arctic and Pacific water sources. Ba as a tracer for river waters (Guay and Falkner, 1997; Taylor et al., 2003), $^{14}$C to quantify ventilation times (Schlosser et al., 1997), and more recently high precision $^{234}$U/$^{238}$U ratios to trace river waters (Andersen et al., 2007), and Nd isotopes to trace input sources and water mass mixing (Andersson et al., 2008; Porcelli et al., 2009). Here we present dissolved Be isotopes as a new tracer for hydrographic processes in the Arctic Ocean.

The cosmogenic radionuclide $^{10}$Be ($T_{1/2} = 1.5$ million years) is produced at a near constant rate by spallation reactions induced by cosmic rays in the upper atmosphere and is then transferred onto the Earth’s surface by precipitation. There is a range of estimates for the global average production of $^{10}$Be of between $0.1 \times 10^6$ atoms/(cm$^2$ yr) (from deposition rates in ice cores (Raisbeck et al., 1978)) and $1.6 \times 10^6$ atoms/(cm$^2$ yr) (from precipitation collection (Somayajulu et al., 1984)). Intermediate values were obtained from theoretical calculations (Lal and Peters, 1967) and lake sediments (Brown et al., 1987). Despite the fact that the highest atmospheric production rates occur at high latitudes due to the weaker shielding of the cosmic ray flux by the Earth’s magnetic field, the highest $^{10}$Be-fluxes to the Earth’s surface are observed in mid latitudes due to the maximum in stratospheric/tropospheric exchange (STE) (Lal and Peters, 1967; Field et al., 2006) and due to the amount of precipitation at a given site (Baskaran, 1995). Due to atmospheric circulation patterns and the distribution of precipitation in the northern hemisphere, the deposition of $^{10}$Be in Greenland and in particular in the Arctic Ocean is very low (Field et al., 2006). For this study we adopted a value of $0.25 \times 10^6$ atoms/(cm$^2$ yr), which was obtained from $^{10}$Be-fluxes measured in Greenland from 1986 to 1990 (Stanzik, 1996) and from an atmospheric GCM run for the same period of time (Heikkilä et al., 2008).

In the open ocean the effects of variable flux to the Earth surface and variations in production rate over time are efficiently homogenized through water mass mixing and vertical exchange. The average $^{10}$Be residence time in the ocean is 500–1000 years resulting in relatively homogenous $^{10}$Be concentrations in Atlantic deep waters of 800–1000 atoms/g and up to 2000 atoms/g in the deep Southern and Pacific Oceans (Kusakabe et al., 1987, 1990; Measures et al., 1996; Frank et al., 2002). This distribution is mainly a function of sorption and remineralization processes of $^{10}$Be along the deep pathways of the global thermohaline circulation. In the Arctic Ocean the only available $^{10}$Be concentrations were measured on three profiles in the Eurasian Basin and did not show any significant variations from an average value of $500 \pm 100$ atoms/g throughout the water column (Strobl et al., 1999).

Beryllium is particle reactive in seawater, which explains its distinct depletion in open ocean surface waters and its systematic enrichment at depth (Kusakabe et al., 1987; Ku et al., 1990). Related to its particle reactivity, it is also known to show boundary scavenging in the ocean (advective transfer of $^{10}$Be from areas of low particulate fluxes to high particle flux areas) (Anderson et al., 1990). If mixing and residence times of water masses are, however, short relative to the residence time of $^{10}$Be, it can be a useful quasi-conservative tracer for water mass mixing, at least on ocean basin-scales (Measures et al., 1996; Frank et al., 2002). In the open Arctic Ocean the use of $^{10}$Be as a water mass tracer is expected to be possible because biogenic particulate fluxes are very low due to perennial sea ice cover and detrital particle fluxes from either eolian or riverine sources are also low compared with other major ocean basins.

The stable isotope $^9$Be is a trace component in all continental rocks, which is released by weathering and transferred to the ocean predominantly via rivers. Despite a study suggesting that a large fraction of the riverine Be flux is immobilized in estuaries and does not reach the open ocean (Kusakabe et al., 1991), it is likely that the riverine transport of dissolved $^9$Be, together with $^{10}$Be originating from precipitation on the continents, is a major source of Be to the ocean. Additional sources of $^9$Be are the partial dissolution of particulate material supplied by rivers or by eolian transport. Concentrations of $^9$Be are variable in surface waters, ranging from 5 to 35 pmol/kg, and are most depleted in the Pacific. Deep waters are uniform at values around 25 pmol/kg (Kusakabe et al., 1987; Ku et al., 1990).

The ratio between dissolved cosmogenic $^{10}$Be and terrigenous $^9$Be (expressed in atoms/atoms) has also been shown to vary systematically between deep ocean basins (Kusakabe et al., 1987, 1990, 1991; Ku et al. 1990; Measures et al., 1996; von Blanckenburg et al., 1996), and has mainly been inferred to be a function of the sorption and remineralization processes described above combined with relatively high inputs of $^9$Be to the smaller Atlantic Basin (e.g. von Blanckenburg et al., 1996). This results in ratios of $4 \times 10^{-8}$ in Atlantic waters, values of about $10 \times 10^{-8}$ in the Indian and Southern Oceans, $10-16 \times 10^{-8}$ in Pacific deep waters, and values of up to $33 \times 10^{-8}$ in surface waters of the Pacific Ocean.
We present the first comprehensive data set of dissolved $^{10}$Be and $^9$Be concentrations covering all major basins of the Arctic Ocean in order to investigate the processes controlling their distribution and their applicability as tracers.

2. MATERIAL AND METHODS

2.1. Sampling area and hydrography

The water masses in the upper Arctic Ocean between the surface and ~150 m depth are generally subdivided into the cold Polar mixed layer, which reaches depths of 30–50 m, and the underlying cold halocline, characterized by near freezing temperature and increasing salinity with depth (Figs. 2 and 3). This structure originates from the cooling and ice melting-driven freshening of the incoming Atlantic water, which sinks north of Svalbard to depths of about 150 m. This water is then advected with the general circulation into the main Arctic Basin. Farther into the basin this layer becomes covered with freshwater from the rivers and is also modified by the supply of cold water from the Siberian shelves injected into the halocline (Aagaard et al., 1985; Anderson et al., 1994). Whereas in the Eurasian Basin the halocline waters predominantly originate from the Siberian shelves, the upper halocline waters in the Canada Basin are dominated by modified Pacific waters, centered between 50 and 150 m, that are underlain by lower halocline waters of eastern Arctic origin (Shimada et al., 2005) (Fig. 3). At depths of between 200 and about 1700 m the Arctic Intermediate Water (AIW) is found, which is of Atlantic origin and is characterized by significantly higher temperatures (Aagaard et al., 1985; Karcher and Oberhuber, 2002). This layer can be further subdivided into the Atlantic layer, characterized by temperatures above 0°C and extending to about 750 m, and the Upper Polar Deep Water below (Rudels et al., 2004). The core of Atlantic waters with the...
Fig. 2. Salinity and temperature profiles in the Eurasian Basin and Fram Strait sites (modified from Andersson et al., 2008). Salinity (psu, solid lines) and potential temperature, $\theta$ (°C, dotted lines) depth profiles from: (a) Fram Strait, site #1, (b) North Svalbard, site #2, (c) Slope North Svalbard, site #3, (d) Nansen Basin, site #4. The western Atlantic inflows and the Fram Strait outflow are marked. In (d) the principal stratification of the Arctic Ocean is marked as: Polar mixed layer (PML) <50 m, cold halocline layer (CHL), 50–150 m, Atlantic water (AW), Upper Polar Deep Water (UPDW), and Eurasian Basin Deep Water (EBDW).

Fig. 3. Salinity and temperature profiles for the central Arctic Ocean and Canadian Basin sites (modified from Porcelli et al., 2009). The salinity and potential temperature structure for the deep profiles sampled at each basin in the upper 200 m (upper panel) and throughout the water column (lower panel). The three deep sampled sites are shown for the Canada Basin (site #C, thick line, site #D, medium line, and site #E, fine line). For the Amundsen Basin site #5 (Fig. 1), both the deep profile for station #21 (thick line), and the shallow profile for station #22 (thin line), are shown. The Makarov Basin profile corresponds to site #6 in Fig. 1. The uppermost 10–20 m, the PML, is well mixed and has lowered salinities followed by the (CHL) in the Eurasian Basins, with relatively constant low temperatures as salinity increases with depth. Below the CHL is the prominent temperature maximum of the Atlantic Layer (AL), defined by a temperature above 0 °C (Rudels et al., 2004). This is underlain by UPDW to 1700 m, and the quite uniform Polar Deep Water (PDW) at greater depths. In the Canada Basin, the CHL is composed of Pacific waters; Pacific Winter Waters (PWW), with salinities around $S = 33.1$ and temperatures near freezing, underlain by Pacific summer water with $S < 33$ and represented by a temperature maximum above the PWW (Steele et al., 2004).
highest temperatures is found at ~300 m water depth. The deep water below the AIW is also ultimately of Atlantic origin (Polar Deep Water – PDW) but has been modified so that it has a higher salinity and a lower temperature than the AIW, most likely due to minor contributions of brines formed during sea ice melting on the shelves (Aagaard et al., 1985; Schllosser et al., 1995).

Samples were collected in summer 2000 during the AWS 2000 expedition with the US coast guard icebreaker “Polar Star” and in summer 2001 during the Swedish Arctic Ocean 2001 (AO-01) expedition with the research ice breaker “Oden” (Fig. 1). During the AWS 2000 cruise two complete deep water profiles were sampled in the Canada Basin accompanied by three vertical profiles from the shelf and slope off Barrow, Alaska. During the AO-01 cruise a total of three complete profiles were sampled from the two major basins of the Eurasian part of the Arctic Ocean (Nansen and Amundsen Basin) and the Makarov Basin. Additional samples were taken in the Fram Strait during the AO-02 expedition in 2002. Two shallow profiles were sampled near Svalbard and a total of eight additional surface water samples were taken along the cruise track of AO-01. This set of 82 seawater samples is complemented by samples of the Lena, Ob, Yenisey and McKenzie rivers, which were taken above tidal influence in summer 2003 during the PARTNERS (Pan-Arctic River Transport of Nutrients, Organic Matter and Suspended Sediments) project (see Cooper et al., 2005 and http://www.ecosystems.mbl.edu/partners). Aliquots of the same samples from both cruises and the rivers were also subject to studies on the distribution of radiogenic Nd isotopes (Andersson et al., 2008; Porcelli et al., 2009), Hf isotopes (Zimmermann et al., 2009) and U isotopes (Andersen et al., 2007). In addition, water samples collected from the same stations on the AWS 2000 cruise were used to study the distribution of natural Th isotopes (Trimble et al., 2004; Trimble and Baskaran, 2005).

2.2. Analytical methods

The samples were taken with a standard CTD rosette by 10 L Niskin bottles on the AWS 2000 cruise. On the AO-01 expedition the samples were taken with 20–60 L Teflon-coated GoFlo bottles attached to a CTD rosette. Each sample of approximately 20 L was filtered through 142-mm diameter Millipore nitrocellulose membrane filters. The samples were weighed, acidified with 20 mL of concentrated HCl and 0.3 mL of a 9Be standard solution (1000 ppm) were added as a carrier. Separate 2 L filtered samples were taken at each depth and were acidified with 1 mL of concentrated HCl for the determination of the natural 9Be concentrations. The 20 L samples were spiked in the laboratory for Nd and U-series isotope work and 250 mg Fe carrier (FeCl3) was added. After several days of equilibration the dissolved metals were coprecipitated with Fe(OH)3 at pH 7–8. The samples were centrifuged after two days. The Fe(OH)3 precipitate was redissolved in 6 M HCl and the Nd, U, and Th fractions were separated by ion chromatographic methods while the Be was eluted together with Fe. The samples were evaporated to dryness and then redissolved in 6 M HCl. Fe was separated from Be on a Biorad AG 1 × 8 anion exchange column. The pH of the elute containing the Be was then adjusted to a pH of 2.5 with ammonia followed by addition of 1 mL of a saturated Na–EDTA solution to complex Al and other divalent cations (Henken-Mellies et al., 1990). After one hour the solution was brought to pH 7–8 and then left over night for the precipitation of Be(OH)2. Be(OH)2 was separated by centrifugation and washed with distilled water and centrifuged several times to reduce the remaining amount of B in the samples. The samples were then treated with concentrated HNO3 and dried three times. The residue was then mixed with AgNO3, ashed to BeO at 850 °C under vacuum, and prepared for Accelerator Mass Spectrometry (AMS) measurements at the AMS facility of ETH Zürich and the Paul Scherrer Institute following standard procedures. The concentrations were normalized to internal standard S555 with a nominal 10Be/9Be ratio of 95.5 ± 12 (Table 1). The propagated 1 SD statistical uncertainties of individual samples include the statistics of the 10Be “counts” and the reproducibility of repeated measurements, which were performed for each sample.

For the determination of the 9Be concentrations, 750 µL of a 0.03 g/g FeCl3 solution was added to 1 L aliquots and the samples were precipitated in the same way as for 10Be analysis. After washing of the precipitate and ion-chromatographic removal of Fe analogous to the 10Be preparation, a Rh/Ir standard was added and the samples were measured using ICP-MS at the Laboratory for Inorganic Chemistry of ETH Zürich. Several tests with known amounts of 9Be resulted in yields close to 100% and repeat measurements including separate precipitations of the same water samples provided an analytical uncertainty of the 9Be concentration measurements of 10%, which was applied for the calculation of the errors of 10Be/9Be ratios.

3. RESULTS

Salinity and potential temperature profiles for all stations of this study are provided in Table 1 and are plotted in Figs. 2 and 3 (Andersson et al., 2008; Porcelli et al., 2009). Further background oxygen isotope and silicate concentration data of the same profiles can be found in Zimmermann et al. (2009). The water column distributions of 10Be and 9Be concentrations, as well as 10Be/9Be ratios follow clearly resolvable and distinct patterns (Fig. 4 and Table 1).
Table 1
Concentrations of dissolved $^{10}$Be, $^{9}$Be, as well as the $^{10}$Be/$^{9}$Be ratio against water depth (m).

<table>
<thead>
<tr>
<th>Site: Station, water depth</th>
<th>Water depth (m)</th>
<th>Salinity (psu)</th>
<th>$\theta^\circ$ (°C)</th>
<th>$^{10}$Be conc. (atoms/g)</th>
<th>$^{9}$Be conc. (pmol/kg)</th>
<th>$^{10}$Be/$^{9}$Be ($\times 10^{3}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>AO-02</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1: Fram Strait, Station #26, 1379 m</td>
<td>8</td>
<td>52.00</td>
<td>−1.69</td>
<td>376 ± 43</td>
<td>22.1</td>
<td>2.83 ± 0.43</td>
</tr>
<tr>
<td>80°20.28’N, 97°20.73’W</td>
<td>50</td>
<td>32.88</td>
<td>−1.61</td>
<td>418 ± 45</td>
<td>25.6</td>
<td>2.71 ± 0.40</td>
</tr>
<tr>
<td>09-May-2002</td>
<td>100</td>
<td>34.08</td>
<td>−1.75</td>
<td>383 ± 43</td>
<td>21.9</td>
<td>2.90 ± 0.44</td>
</tr>
<tr>
<td>1324</td>
<td>34.91</td>
<td>−0.26</td>
<td>472 ± 48</td>
<td>12.3</td>
<td>6.37 ± 0.91</td>
<td></td>
</tr>
<tr>
<td><strong>AO-01</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2: LegI, Station #2, 158 m</td>
<td>8</td>
<td>34.52</td>
<td>3.89</td>
<td>678 ± 81</td>
<td>28.4</td>
<td>3.96 ± 0.62</td>
</tr>
<tr>
<td>80°25.85’N, 15°31.05’E</td>
<td>20</td>
<td>34.60</td>
<td>4.06</td>
<td>507 ± 70</td>
<td>16.7</td>
<td>5.04 ± 0.86</td>
</tr>
<tr>
<td>13-July-2001</td>
<td>40</td>
<td>34.78</td>
<td>3.71</td>
<td>505 ± 118</td>
<td>9.6</td>
<td>8.74 ± 2.22</td>
</tr>
<tr>
<td>17-July-2001</td>
<td>100</td>
<td>34.96</td>
<td>3.05</td>
<td>526 ± 59</td>
<td>14.1</td>
<td>6.20 ± 0.93</td>
</tr>
<tr>
<td>3: LegII, Station #1, 497 m</td>
<td>8</td>
<td>32.72</td>
<td>−0.97</td>
<td>467 ± 41</td>
<td>6.8</td>
<td>11.40 ± 1.52</td>
</tr>
<tr>
<td>80°16.70’N, 2622.97’E</td>
<td>80</td>
<td>34.86</td>
<td>2.42</td>
<td>535 ± 51</td>
<td>12.8</td>
<td>6.94 ± 0.96</td>
</tr>
<tr>
<td>17-July-2001</td>
<td>150</td>
<td>34.95</td>
<td>2.62</td>
<td>592 ± 38</td>
<td>13.2</td>
<td>7.45 ± 0.88</td>
</tr>
<tr>
<td>18-July-2001</td>
<td>300</td>
<td>34.96</td>
<td>2.21</td>
<td>553 ± 44</td>
<td>14.3</td>
<td>6.42 ± 0.82</td>
</tr>
<tr>
<td>4: LegII, Station #10, 4025 m</td>
<td>40</td>
<td>34.31</td>
<td>−1.73</td>
<td>492 ± 135</td>
<td>12.2</td>
<td>6.70 ± 0.96</td>
</tr>
<tr>
<td>80°46.93’N, 31°57.10’E</td>
<td>100</td>
<td>34.43</td>
<td>−1.62</td>
<td>682 ± 81</td>
<td>12.5</td>
<td>9.06 ± 1.41</td>
</tr>
<tr>
<td>18-July-2001</td>
<td>300</td>
<td>34.93</td>
<td>1.98</td>
<td>1161 ± 247</td>
<td>13.4</td>
<td>14.39 ± 3.38</td>
</tr>
<tr>
<td>5: LegII, Station #11, 4039 m</td>
<td>8</td>
<td>33.98</td>
<td>−1.58</td>
<td>460 ± 34</td>
<td>14.3</td>
<td>5.34 ± 0.66</td>
</tr>
<tr>
<td>80°16.87’N, 3339.81’E</td>
<td>30</td>
<td>34.24</td>
<td>−1.69</td>
<td>518 ± 52</td>
<td>22.9</td>
<td>3.76 ± 0.53</td>
</tr>
<tr>
<td>18-July-2001</td>
<td>100</td>
<td>34.46</td>
<td>−1.16</td>
<td>504 ± 52</td>
<td>11.9</td>
<td>8.90 ± 1.90</td>
</tr>
<tr>
<td>6: LegII, Station #12, 4027 m</td>
<td>50</td>
<td>34.87</td>
<td>0.53</td>
<td>348 ± 95</td>
<td>11.6</td>
<td>5.00 ± 1.45</td>
</tr>
<tr>
<td>80°44.04’N, 35°14.89’E</td>
<td>100</td>
<td>34.87</td>
<td>0.53</td>
<td>664 ± 70</td>
<td>11.6</td>
<td>9.55 ± 1.39</td>
</tr>
<tr>
<td>19-July-2001</td>
<td>200</td>
<td>34.20</td>
<td>−1.66</td>
<td>415 ± 37</td>
<td>13.3</td>
<td></td>
</tr>
<tr>
<td>7: LegII, Station #16, 3595 m</td>
<td>8</td>
<td>34.09</td>
<td>−1.76</td>
<td>460 ± 34</td>
<td>14.3</td>
<td></td>
</tr>
<tr>
<td>80°28.89’N, 49°10.71’E</td>
<td>80</td>
<td>34.08</td>
<td>−1.65</td>
<td>408 ± 48</td>
<td>11.3</td>
<td>6.00 ± 0.93</td>
</tr>
<tr>
<td>20-July-2001</td>
<td>80</td>
<td>34.08</td>
<td>−1.65</td>
<td>561 ± 66</td>
<td>11.3</td>
<td>8.24 ± 1.27</td>
</tr>
<tr>
<td>8: LegII, Station #20, 4410 m</td>
<td>8</td>
<td>32.63</td>
<td>−1.66</td>
<td>596 ± 40</td>
<td>34.6</td>
<td>2.86 ± 0.34</td>
</tr>
<tr>
<td>80°16.70’N, 82°54.31’E</td>
<td>80</td>
<td>33.03</td>
<td>−1.76</td>
<td>693 ± 50</td>
<td>45.7</td>
<td>2.52 ± 0.31</td>
</tr>
<tr>
<td>22-July-2001</td>
<td>30</td>
<td>33.03</td>
<td>−1.76</td>
<td>693 ± 50</td>
<td>45.7</td>
<td>2.52 ± 0.31</td>
</tr>
<tr>
<td>9: LegII, Station #21, 4400 m</td>
<td>75</td>
<td>33.74</td>
<td>−1.81</td>
<td>306 ± 53</td>
<td>18.1</td>
<td>2.81 ± 0.56</td>
</tr>
<tr>
<td>80°24.48’N, 95°22.78’E</td>
<td>300</td>
<td>33.74</td>
<td>1.31</td>
<td>462 ± 45</td>
<td>14.2</td>
<td>5.40 ± 0.75</td>
</tr>
<tr>
<td>22-July-2001</td>
<td>500</td>
<td>34.87</td>
<td>0.53</td>
<td>348 ± 95</td>
<td>11.6</td>
<td>5.00 ± 1.45</td>
</tr>
<tr>
<td>1000</td>
<td>500</td>
<td>34.87</td>
<td>0.53</td>
<td>664 ± 70</td>
<td>11.6</td>
<td>9.55 ± 1.39</td>
</tr>
<tr>
<td>22-July-2001</td>
<td>2500</td>
<td>34.92</td>
<td>−0.86</td>
<td>740 ± 132</td>
<td>13.9</td>
<td>8.84 ± 1.81</td>
</tr>
<tr>
<td>3900</td>
<td>34.93</td>
<td>−0.95</td>
<td>430 ± 36</td>
<td>14.3</td>
<td>4.99 ± 0.65</td>
<td></td>
</tr>
<tr>
<td>10: LegII, Station #22, 4372 m</td>
<td>10</td>
<td>31.36</td>
<td>−1.55</td>
<td>493</td>
<td>49.3</td>
<td></td>
</tr>
<tr>
<td>80°26.05’N, 109°50.65’E</td>
<td>20</td>
<td>31.51</td>
<td>−1.57</td>
<td>680 ± 47</td>
<td>47.9</td>
<td>2.36 ± 0.29</td>
</tr>
<tr>
<td>23-July-2001</td>
<td>40</td>
<td>33.52</td>
<td>−1.82</td>
<td>23.5</td>
<td></td>
<td></td>
</tr>
<tr>
<td>60</td>
<td>33.66</td>
<td>−1.83</td>
<td>168 ± 35</td>
<td>20.4</td>
<td>1.37 ± 0.32</td>
<td></td>
</tr>
<tr>
<td>100</td>
<td>34.18</td>
<td>−1.58</td>
<td>323 ± 32</td>
<td>12.0</td>
<td>4.47 ± 0.63</td>
<td></td>
</tr>
<tr>
<td>23-July-2001</td>
<td>8</td>
<td>31.10</td>
<td>−1.53</td>
<td>737 ± 48</td>
<td>58.7</td>
<td>2.08 ± 0.25</td>
</tr>
<tr>
<td>11: LegII, Station #24, 2779 m</td>
<td>8</td>
<td>30.85</td>
<td>−1.53</td>
<td>726 ± 47</td>
<td>58.9</td>
<td>2.05 ± 0.24</td>
</tr>
<tr>
<td>80°08.43’N, 132°33.24’E</td>
<td>80</td>
<td>30.85</td>
<td>−1.53</td>
<td>726 ± 47</td>
<td>58.9</td>
<td>2.05 ± 0.24</td>
</tr>
<tr>
<td>23-July-2001</td>
<td>80</td>
<td>30.85</td>
<td>−1.53</td>
<td>726 ± 47</td>
<td>58.9</td>
<td>2.05 ± 0.24</td>
</tr>
</tbody>
</table>

(continued on next page)
<table>
<thead>
<tr>
<th>Site: Station, water depth</th>
<th>Water depth (m)</th>
<th>Salinity (psu)</th>
<th>$\theta^a$ ($^\circ$C)</th>
<th>$^{10}$Be conc. (atoms/g)</th>
<th>$^9$Be conc. $^b$ (pmol/kg)</th>
<th>$^{10}$Be/$^9$Be (atom/atom)</th>
<th>Depth (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>LegII, Station #26, 1513 m</td>
<td>8</td>
<td>30.90</td>
<td>−1.57</td>
<td>652 ± 47</td>
<td>61.5</td>
<td>1.76 ± 0.22</td>
<td>90</td>
</tr>
<tr>
<td>87$^\circ$08.43'N, 132$^\circ$33.24'E</td>
<td>23-July-2001</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>LegII, North Pole, 4260 m</td>
<td>8</td>
<td>31.07</td>
<td>−1.60</td>
<td>783 ± 102</td>
<td>65.2</td>
<td>1.99 ± 0.33</td>
<td>90</td>
</tr>
<tr>
<td>87$^\circ$17.68,43'00.00'N, 132$^\circ$33.24'E</td>
<td>01-Aug-2001</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>6: LegII, Station #30, 3985 m</td>
<td>8</td>
<td>31.23</td>
<td>−1.60</td>
<td>841 ± 83</td>
<td>53.8</td>
<td>2.60 ± 0.36</td>
<td>90</td>
</tr>
<tr>
<td>87$^\circ$08.43,00'00.00'N, 132$^\circ$33.24'E</td>
<td>23-July-2001</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>LegII, Station #26, 1513 m</td>
<td>8</td>
<td>31.23</td>
<td>−1.60</td>
<td>841 ± 83</td>
<td>53.8</td>
<td>2.60 ± 0.36</td>
<td>90</td>
</tr>
<tr>
<td>87$^\circ$08.43,00'00.00'N, 132$^\circ$33.24'E</td>
<td>23-July-2001</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>AWS 2000</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>A: AWS Station #1, 50 m</td>
<td>5</td>
<td>29.31</td>
<td>4.57</td>
<td>200 ± 39</td>
<td>58.5</td>
<td>0.57 ± 0.12</td>
<td>90</td>
</tr>
<tr>
<td>71$^\circ$40.00'00.00'N, 134$^\circ$46.00'E</td>
<td>August-2000</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>B: AWS Station #2, 170 m</td>
<td>5</td>
<td>27.46</td>
<td>0.89</td>
<td>507 ± 61</td>
<td>18.2</td>
<td>4.63 ± 0.72</td>
<td>90</td>
</tr>
<tr>
<td>71$^\circ$49.00'00.00'N, 153$^\circ$55.00'E</td>
<td>August-2000</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>C: AWS Station #3, 3850 m</td>
<td>5</td>
<td>25.90</td>
<td>−1.16</td>
<td>911 ± 56</td>
<td>10.9</td>
<td>13.8 ± 1.63</td>
<td>90</td>
</tr>
<tr>
<td>75$^\circ$12.50'00.00'N, 149$^\circ$57.00'E</td>
<td>August-2000</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>D: AWS Station #4, 3894 m</td>
<td>5</td>
<td>25.90</td>
<td>−1.16</td>
<td>911 ± 56</td>
<td>10.9</td>
<td>13.8 ± 1.63</td>
<td>90</td>
</tr>
<tr>
<td>73$^\circ$45.00'00.00'N, 152$^\circ$00.00'E</td>
<td>August-2000</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>E: AWS Station #5, 1200 m</td>
<td>5</td>
<td>26.56</td>
<td>−1.20</td>
<td>837 ± 76</td>
<td>11.3</td>
<td>12.3 ± 1.66</td>
<td>90</td>
</tr>
<tr>
<td>72$^\circ$14.70'00.00'N, 155$^\circ$04.50'E</td>
<td>August-2000</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Arctic rivers (summer 2003)**

<table>
<thead>
<tr>
<th>River</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Depth (m)</th>
<th>$^{10}$Be conc. (atom/atom)</th>
<th>$^9$Beconc. (pmol/kg)</th>
<th>$^{10}$Be/$^9$Be (atom/atom)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mackenzie (67$^\circ$26.00'N, 133$^\circ$45.00'W)</td>
<td>73.1666</td>
<td>−133.75</td>
<td>683 ± 75</td>
<td>315.6</td>
<td>0.36 ± 0.05</td>
<td></td>
</tr>
<tr>
<td>Ob (66$^\circ$31.00'N, 66$^\circ$36.00'W)</td>
<td>72.5222</td>
<td>−66.6</td>
<td>8982 ± 375</td>
<td>1597.5</td>
<td>0.93 ± 0.10</td>
<td></td>
</tr>
<tr>
<td>Yenisey (69$^\circ$23.00'N, 86$^\circ$09.00'W)</td>
<td>72.3833</td>
<td>−130.5666</td>
<td>2935 ± 120</td>
<td>316.9</td>
<td>1.54 ± 0.17</td>
<td></td>
</tr>
<tr>
<td>Lena (66$^\circ$46.00'N, 123$^\circ$22.00'W)</td>
<td>71.7666</td>
<td>−146.7666</td>
<td>4416 ± 167</td>
<td>1176.4</td>
<td>0.62 ± 0.07</td>
<td></td>
</tr>
</tbody>
</table>

$^a$ Potential temperature.

$^b$ The analytical uncertainty of the $^9$Be concentration measurements is 10%.

$^c$ The uncertainty of the $^{10}$Be/$^9$Be (atom/atom) includes both the uncertainties of the $^{10}$Be and $^9$Be measurements depths with an * represent repeat AMS measurements of the same processed samples due to too low beam currents during initial runs.
Fig. 4. Profiles of dissolved $^{10}$Be (atoms/g) and $^9$Be (pmol/kg) concentrations, and $^{10}$Be/$^9$Be ($\times 10^{-8}$) against water depth (m). The vertical dashed lines mark average deep water values for $^{10}$Be concentrations (500 atoms/g), $^9$Be concentrations (12 pmol/kg) and $^{10}$Be/$^9$Be ($7 \times 10^{-8}$). For each location, the depth of the sea floor is given as a gray box. Note that the upper 500 m are plotted on an expanded scale, the end of which is marked by the horizontal dashed line. The $^{10}$Be concentration profiles plotted in gray mark the data previously obtained by Strobl et al. (1999) at essentially the same locations. For the four data points where duplicate $^{10}$Be measurements exist (see Table 1), weighted means were calculated and included in the plots of $^{10}$Be concentrations and $^{10}$Be/$^9$Be.
and $^9$Be concentrations below 20 m depth of around 500 atoms/g and 12 pmol/kg, respectively, resulting in a $^{10}\text{Be}/^9\text{Be}$ of $\sim 7 \times 10^{-4}$. At the surface, sites #2 and #3 differ markedly in their temperatures. The surface $^{10}\text{Be}$ and $^9\text{Be}$ concentrations at the shallow shelf site (#2) are elevated to values of 680 atoms/g and 28 pmol/kg, respectively, corresponding to high temperatures near 4 $^\circ\text{C}$. The resulting $^{10}\text{Be}/^9\text{Be}$ is low at $4 \times 10^{-4}$. In contrast, at site #3 on the N Svalbard slope the surface $^{10}\text{Be}$ concentration is indistinguishable from the average of 500 atoms/g, whereas $^9\text{Be}$
reaches its minimum (6.8 pmol/kg) of all surface water samples of this study. This results in an elevated \(^{10}\text{Be}/^{9}\text{Be}\) ratio of \(11.4 \times 10^{-8}\) and corresponds to a lower salinity (32.7 psu) and temperature (\(-0.97^\circ\C\)).

The profile in the Nansen Basin (site #4) is a composite of 2 stations (#10/11, Fig. 4 and Table 1) and shows relatively constant \(^{10}\text{Be}\) and \(^{9}\text{Be}\) concentrations around 500 atoms/g and 12 pmol/kg, respectively, as well as \(^{10}\text{Be}/^{9}\text{Be}\) values of \(7 \times 10^{-8}\). Salinities are above 34 psu throughout the profile and the surface layer (uppermost 100 m) is cold (\(-1.62\) to \(-1.73^\circ\C\)). At a water depth of 100 m the \(^{9}\text{Be}\) and \(^{10}\text{Be}\) concentrations of profiles 10 and 11 differ significantly but there is also a significant difference in temperature between the two neighboring locations (\(\theta = -1.6\) and \(-1.16^\circ\C\)) indicating a slight difference in water mass composition, i.e. a slight difference in the depth of the boundary between the cold halocline layer above and water mass composition, i.e. a slight difference in the depth (32.7 psu) and temperature (\(-0.97^\circ\C\)).

The profiles in the central Amundsen and Makarov Basins are characterized by very similar patterns with systematically high \(^{10}\text{Be}\) concentrations of 700–900 atoms/g in the upper 20–30 m, followed by a sharp drop to minimum values of 160–330 atoms/g at depths below 60 and 100 m and then an increase to values between 400 and 700 atoms/g, which is within the range of all Arctic deep waters (350–750 atoms/g). Similar to \(^{10}\text{Be}\), the \(^{9}\text{Be}\) concentrations in the upper 100 m at these sites reach high values of up to 65 pmol/kg at the surface and decrease rapidly to values below 25 pmol/kg at depths below 30 m. These surface waters are characterized by a low salinity of 31.2 psu which increases to 34.2 psu within the upper 100 m, while temperatures stay below \(-1.55^\circ\C\). Below 100 m water depth, the profiles of \(^{9}\text{Be}\) in both the Amundsen and Makarov Basins show near constant values of between 10 and 15 pmol/kg throughout the water column. The resulting \(^{10}\text{Be}/^{9}\text{Be}\) data show minima of between 1.4 and \(2.8 \times 10^{-8}\) in the upper 100 m, followed by intermediate values of \(5 \times 10^{-8}\) in the Atlantic layer at 300 m depth. Below 300 m the \(^{10}\text{Be}/^{9}\text{Be}\) is again in the range of all Arctic deep waters (5–9 \(\times 10^{-8}\)).

The profiles in the deep Canada Basin (#C–E, Figs. 1 and 2) show \(^{10}\text{Be}\) concentration profiles very similar to those of the Amundsen and Makarov Basins, with elevated concentrations of up to 900 atoms/g in the uppermost 25 m followed by pronounced minima of between 200 and 300 atoms/g at 35–150 m depth, again followed by an increase to concentrations of 300–500 atoms/g below 400 m depth. In pronounced contrast to the Eurasian Basin, however, the surface waters in the Canada Basin show low \(^{9}\text{Be}\) concentrations of around 10 pmol/kg in the uppermost 50 m, followed by an increase to values between 17 and 31 pmol/kg at 150 m depth underlain by again low concentrations of 9–15 pmol/kg to depths of 3700 m. This results in pronounced near surface (upper 25 m) maxima in \(^{10}\text{Be}/^{9}\text{Be}\) of up to \(14 \times 10^{-8}\), followed by low values of \(2.4–3.3 \times 10^{-8}\) between 35 and 85 m depth and a pronounced minimum between 1.1 and \(2 \times 10^{-8}\) at 125–150 m water depth, before the values again increase to values around \(7 \times 10^{-8}\). The low \(^{9}\text{Be}\) and high \(^{10}\text{Be}\) concentrations and \(^{10}\text{Be}/^{9}\text{Be}\) ratios near the surface coincide with salinities as low as 25.9 psu. Temperatures are relatively low at \(\sim 1.2^\circ\C\).

The two profiles on the shelf near the coast of Alaska (#A and B, Figs. 1 and 2) show significantly higher \(^{9}\text{Be}\) concentrations of up to 65 pmol/kg at the shallowest site #A near the coast and up to 30 pmol/kg at site #B. At the same time, the \(^{10}\text{Be}\) concentration exhibits values as low as 200 atoms/g close to the coast, which results in the lowest \(^{10}\text{Be}/^{9}\text{Be}\) (0.5 \(\times 10^{-8}\)) measured in the seawater samples reported in this study. While the surfaces of all sites in the Canada Basin exhibit low temperatures, the temperatures at site #A are much higher (4.5 °C).

For the four river samples (Fig. 1) maximum concentrations of both \(^{10}\text{Be}\) (9000 atoms/g) and \(^{9}\text{Be}\) (1600 pmol/kg) were found for the Ob river. Minimum \(^{10}\text{Be}\) concentrations of 700 atoms/g and \(^{9}\text{Be}\) concentrations of 300 pmol/kg were obtained for a sample from the McKenzie river, while the samples from the Lena and Yenisey rivers are intermediate between these two extremes. The origin of this variability is probably a combination of lithology and size of the drainage areas and river chemistry but without further data this cannot be confirmed. It is also clear that these few measurements are unlikely to represent the annual average values, but do provide a first estimate for the composition of Arctic river runoff. Despite the large range in \(^{10}\text{Be}\) and \(^{9}\text{Be}\) concentrations, the \(^{10}\text{Be}/^{9}\text{Be}\) in the river waters is consistently very low and only varies between 0.36 \(\times 10^{-8}\) and 1.54 \(\times 10^{-8}\).

4. DISCUSSION

The Be isotope composition shows a large and systematic variability in the upper 150 m of the Arctic water column, which is related to salinity and temperature variations influenced by riverine inputs, formation and melting of sea ice, the admixture of waters from the Arctic shelves, scavenging processes, exchange processes with the shelf sediments, as well as advection of waters from the Pacific and Atlantic Basins. At depths below 300–400 m the Be isotope variability is small and is clearly dominated by waters of Atlantic origin, and the constancy of the concentrations and isotope compositions of Be across the basin suggests that vertical exchange linked to particle adsorption/desorption processes (scavenging) does not play a significant role for the Be distribution below 400 m depth, most likely due to the very low particle concentrations.

4.1. Shallow water processes: riverine inputs and scavenging on the shelves

The surface water salinities obtained for the samples of this study range from 25.9 to 34.5 psu which is mainly a consequence of inputs from the large Arctic rivers combined with some contributions from sea ice melting. This
has been demonstrated for example using a combination of geochemical tracers such as dissolved Si concentrations and oxygen isotopes (cf. Bauch et al., 1995; Yamamoto-Kawai et al., 2008). Combined \(^{10}\)Be and \(^{9}\)Be concentrations should have a large potential to distinguish different freshwater sources due to their different input mechanisms.

4.1.1. Eurasian Basin

In Fig. 5, the salinities of the upper 300 m in the Eurasian Basin transect of AO-01 (Björk et al., 2002) are plotted together with the distribution of the concentrations of \(^{10}\)Be, \(^{9}\)Be, and of \(^{10}\)Be/\(^{9}\)Be. Decreased salinities in the Amundsen and Makarov Basins due to inputs of river waters are accompanied by high \(^{10}\)Be and \(^{9}\)Be concentrations and low \(^{10}\)Be/\(^{9}\)Be (2–3 \(\times\) \(^{10}\)Be Be content of 500–600 atoms/g) have lost \(^{10}\)Be prior to mixing with the river waters in the upper water column due to scavenging processes on the shelves. That would, however, also have led to a loss of \(^{9}\)Be, which is not observed. The data that mainly prevent a correlation between \(^{10}\)Be concentrations and salinity for the entire range between 31 an 35 psu are the too low \(^{10}\)Be concentrations for salinities between 33 and 34 psu at water depths between 50 and 150 m. This strongly indicates that the Atlantic waters were modified by mixing with halocline waters with very low \(^{10}\)Be concentrations either originating from the Siberian shelves or in some cases even from the Canadian Basin, which are not correspondingly low in their \(^{9}\)Be concentrations (see Section 4.2 below). Alternatively these waters may have exchanged with the shelf sediments and thereby changed \(^{10}\)Be and \(^{9}\)Be content, as well as the \(^{10}\)Be/\(^{9}\)Be ratio. These data show that \(^{9}\)Be isotopes only trace admixture of riverine waters in the Eurasian Basin at salinities below 33 psu.

4.1.2. Inter-annual variability in the Eurasian Basin

It has been demonstrated that the hydrography of the Eurasian Basin of the Arctic Ocean (here including the Makarov Basin), in particular the surface water masses and the cold halocline, have been highly variable on interannual time scales (Björk et al., 2002). It was shown that the cold halocline layer found during the AO-01 expedition was also present in the Makarov and Amundsen Basins during the joint RV Oden ‘91 and RV Polarstern expedition ARK VIII/3 in 1991, whereas in 1995 it was only found in the Makarov Basin and in 1996 it was not found at all in the Eurasian Basin. River runoff was clearly present in surface of the Amundsen and Makarov Basins in 1991, covering 45% of the Eurasian Basin, retreated in 1994, disappeared from the Eurasian Basin in 1996 and advanced again in 2001 (Anderson et al., 2004). Although coverage of the Eurasian Basin in 2001 was smaller than in 1991, the fraction of riverine runoff-derived waters reached more than 11% in the Makarov Basin in 2001. The fraction of river water only reached 7% in 1991 but in contrast to 2001 covered most of the Amundsen Basin (Anderson et al., 2004).

In order to investigate if changes in the surface freshwater distribution in the Eurasian Basin over time are also reflected in the \(^{10}\)Be concentrations on annual time scales, the variability of \(^{10}\)Be profiles obtained during the AO-01 cruise are directly compared to results obtained during the expedition ARK VIII/3 in 1991 (Strobl et al., 1999). The data from surface stations of the 2001 cruise show a clear correlation to the distribution of riverine waters and the
Fig. 5. Transect of the upper 350 m of (a) $^{10}\text{Be}$ concentrations (atoms/g), (b) $^9\text{Be}$ concentrations (pmol/kg) and (c) $^{10}\text{Be}/^9\text{Be}$ along the cruise track of the AO-01 cruise between Svalbard and the Makarov Basin in the central Arctic Ocean. The green labels denote the $^{10}\text{Be}$ results obtained in a previous study based on water samples taken during in the RV Polarstern cruise ARK VIII/3 in 1991 (Strobl et al., 1999). The contours show the salinity in the upper 350 m (Björk et al., 2002). The depth profiles are the same as in Fig. 4. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
associated salinity minimum, in which the concentrations of both Be isotopes are significantly increased (Fig. 5). There are ¹⁰Be profiles available from three sites sampled during the 1991 expedition, two of which (152 and 173) are essentially identical to the locations of sites #4 and #5 of the AO-01 cruise, respectively (Table 1 and Figs. 4 and 5). The data from 1991 reveal uniform ¹⁰Be concentrations of around 500 atoms/g throughout the deeper water column (Fig. 4), whereas the concentrations at 50 m depth were slightly higher in the Amundsen Basin than in the Nansen...
Basin, which is consistent with the decreased salinities in the Amundsen Basin in 1991 (Björk et al., 2002). These increased 10Be concentrations were clearly not as high as directly at the surface in 2001, but correspond very well to the salinities and concentrations obtained at water depths around 50 m in 2001 (Fig. 4), suggesting that the 10Be concentrations indeed directly reflect the hydrographic variability.

4.1.3. Canada Basin

In marked contrast to the Eurasian Basin, the lowest surface water salinities (as low as 26 psu in the uppermost 25 m) of the open Canada Basin are all marked by low 9Be concentrations of between 9 and 20 pmol/kg (Figs. 4 and 6). However, these correspond to high 10Be concentrations of up to 1000 atoms/g, resulting in 10Be/9Be ratios of up to 14 × 10^-8. This observation is consistent with the very efficient estuarine removal of 9Be originating from Mackenzie River waters (Fig. 7). Considering that riverine 10Be from the Mackenzie must be scavenged to exactly the same extent as 9Be, there must be an additional source of 10Be to the surface waters in the Canada Basin. The atmospheric flux of ~0.25 × 10⁶ atoms/(cm² yr) (Heikillä et al., 2008) supplies significant amounts of 10Be contributing to the observed maximum in the uppermost 25 m of the open Canada Basin. It is also possible that melting of sea ice contributed to the increased 10Be concentrations. Sea ice would provide fresh and cold waters containing no significant amounts of 9Be but large amounts of 10Be, ultimately also resulting from atmospheric fallout. Melt water originating from sea ice is generally found in the upper 20–30 m of the Arctic water column but comparison of salinity and oxygen isotopes in these samples excludes any large fraction of molten sea ice (Andersen et al., 2007; Zimmermann et al., 2009). It does, however, not require a lot of sea ice to add significant amounts of 10Be to the surface waters without significantly altering the oxygen isotope composition of the surface waters. If we assume that the average age of the sea ice in the Canada Basin was on the order of 3–5 years in 2000 (cf. Rigor and Wallace, 2004) at an average thickness of 2–3 m, the concentration of 10Be in the ice amounts to ~0.25–0.6 × 10⁷ atoms/g. Melting of 2 m of sea ice would thus add ~0.5–1.2 × 10⁷ atoms/(cm² yr) of 10Be to the surface corresponding to 2–5 times the local annual fallout and for the upper 25 m this addition would result in an increase of the 10Be concentrations by 200–500 atoms/g, which already approximately corresponds to the observations. At the same time the salinity would be diminished by only 2.8 psu, which would not result in a resolvable deviation from the relationship between oxygen isotopes and salinity in the Canada Basin (Andersen et al., 2007; Zimmermann et al., 2009).

In addition to the 10Be originating from atmospheric fallout, sea ice sometimes contains sediments ("dirty ice") that have very high concentrations of reactive metals including Fe, Al, 210Pb and adsorbed 10Be (Eisenhauer et al., 1994; Measures, 1999; Baskaran, 2005), which are released to the water column during melting/freezing cycles of sea ice. Despite the fact that there are no direct data available, it is also likely that 9Be is released from these sediments. Dirty ice may thus be an important contributor of both 10Be and 9Be, but its importance cannot be assessed further without information on the 10Be/9Be ratio released form the ice-bound sediments.

Thus, the observed distribution of Be isotopes in the uppermost part of the water column of the Canada Basin is best explained by a combination of scavenging processes of both 9Be and 10Be in the Mackenzie estuary and the addition of 10Be from atmospheric fallout and/or sea ice melting. In view of the residence time of Be in the uppermost 25 m of the water column in the Canada Basin on the order of several years, this interpretation is consistent with the observation of enhanced melting of multi-year sea ice in the Canada Basin during the summers between 1979 and 2002 (Rigor and Wallace, 2004), which must have contained high amounts of 10Be derived from atmospheric fallout.

The main reason for the contrasting behavior of Be in the Canada Basin with that in Eurasian Basin thus most likely is efficient removal of Be from Mackenzie river waters through coagulation at the river mouth and active scavenging in high-particle concentration zones, combined with a longer residence of the surface waters in the Canada Basin, which allows for a significant contribution of 10Be from atmospheric fallout and melting of old sea ice to the upper 25 m of the water column.

4.2. Pacific–Arctic exchange

Pacific waters enter the Canada Basin through the shallow Bering Strait, where modifications by sea ice formation may occur. In the Canada Basin, Pacific water occupies the
water depth of the cold halocline at 50–150 m, below which halocline waters from the eastern Arctic are found (Shimada et al., 2005). This Pacific inflow is reflected in a marked minimum in $^{10}\text{Be}$ concentrations of around 200 atoms/g at sites #2 to #5 over the depth interval 50–150 m. This is accompanied by $^9\text{Be}$ concentrations of between 11 and 15 pmol/kg at 50–85 m, resulting in low $^{10}\text{Be}/^{9}\text{Be}$ ratios of between 2.4 and $3 \times 10^{-8}$. At greater depths of 125–150 m within the cold halocline layer, $^{10}\text{Be}$ concentrations are higher (17–31 pmol/kg), resulting in very low $^{10}\text{Be}/^{9}\text{Be}$ ratios of between 1.1 and $2 \times 10^{-8}$.

The low $^{10}\text{Be}$ and $^9\text{Be}$ concentrations in the upper cold halocline are similar to highly Be-depleted Pacific surface waters. The nearest Pacific $^{10}\text{Be}$ and $^9\text{Be}$ data available for comparison were obtained in the Bering Sea at 53°N, 177°E (Kusakabe et al., 1990; Xu, 1994). The Bering Sea profile shows surface $^{10}\text{Be}$ concentrations of 400 atoms/g and $^9\text{Be}$ concentrations of 15 pmol/kg (Kusakabe et al., 1990; Xu, 1994). To reach the low values in the upper cold halocline of the Canada Basin these waters must, however, have been modified through exchange processes with particles or with the underlying sediments when passing through the shallow particle-rich Bering Strait waters. This is also visible in the surface $^{10}\text{Be}/^{9}\text{Be}$ in the Bering Sea of $4.5 \times 10^{-8}$ which is higher than the halocline ratios observed in the Canada Basin.

The salinities and temperatures at 125–150 m depth are consistent with the relatively fresh Pacific Winter Water, which has originally been supplied through the Bering Strait and is subsequently modified by sea ice formation along the coastal polynya between Point Barrow and Cape Lisburne (cf. Shimada et al., 2005). At this depth, however, additional contributions to $^9\text{Be}$ are required. The low $^{10}\text{Be}/^{9}\text{Be}$ ratios and high $^9\text{Be}$ concentrations suggest that river waters must have been admixed in the area where the Pacific Winter Water was modified through sea ice formation before it reached its observed depths of 125–150 m. The highly radiogenic Nd isotope composition found in the halocline waters excludes a significant fraction of riverine contributions from the Mackenzie River (Porelli et al., 2009). Given that Mackenzie River waters present in the Canada Basin have obviously also lost most of their dissolved Be in the estuary (see discussion above), the inflowing Pacific waters must either have received river water contributions from Siberia, such as from the Kolyma, which would be in agreement with the consistently radiogenic Nd isotope composition at this water depth (Porelli et al., 2009). Alternatively, these waters may have exchanged with the sediments in the Chukchi Sea and thereby modified their Be isotope composition, which would be consistent with the radiogenic Nd isotope composition of waters in the Chukchi Sea (Duhlqvist et al., 2007; Porelli et al., 2009).

The Pacific contributions to the upper halocline waters can clearly be traced across the Arctic Basin as evidenced by the low $^{10}\text{Be}$ concentrations and correspondingly low $^{10}\text{Be}/^{9}\text{Be}$ ratios at depths of 60–100 m water depth at profiles #5 and #6 in the Amundsen and Makarov Basins and possibly as far as the Fram Strait (profile #1). This interpretation is consistent with observations based on Ba contents at this depth (e.g. Taylor et al., 2003) and with significantly more Pacific-like Nd isotope compositions at 60 and 75 m depth than the waters above and below in the Amundsen Basin (Porcelli et al., 2009). The waters below the cold halocline are characterized by $^{10}\text{Be}$ and $^9\text{Be}$ concentrations typical for Atlantic-dominated water masses and do not show any influence of the Pacific inflow.

### 4.3. Atlantic–Arctic exchange

The inflow of Atlantic water can be traced in the entire Arctic Basin by its high temperature. In the Eurasian Basin (sites #5 and #6) the high temperature core of the Atlantic water layer is found at a depth of ~300 m and this water is characterized by $^{10}\text{Be}$ concentrations between 400 and 500 atoms/g and $^9\text{Be}$ concentrations of ~14 pmol/kg, corresponding to a $^{10}\text{Be}/^{9}\text{Be}$ ratio of $5 \times 10^{-8}$. Atlantic water flows into the Arctic Ocean through the Fram Strait and the Barents Sea (Andersson et al., 2008). The Fram Strait Branch is captured at site #2, clearly identified by its high temperatures and also by its Nd isotope characteristics (Andersson et al., 2008). This current supplies waters with slightly elevated $^{10}\text{Be}$ concentrations and significantly increased $^9\text{Be}$ concentrations of 28 pmol/kg at the surface and lower values in deeper waters. These data suggest that the inflow of Atlantic waters is characterized by somewhat elevated $^9\text{Be}$ and mean Arctic $^{10}\text{Be}$ concentrations resulting in an input $^{10}\text{Be}/^{9}\text{Be}$ signal of between 5 and $7 \times 10^{-8}$, which is indistinguishable from the deep water signal in the deep Arctic Ocean. This is consistent with a dominantly Atlantic origin of deep waters in the Arctic Ocean.

The outflow of the Arctic Ocean through Fram Strait (site #1) is characterized by significantly elevated $^9\text{Be}$ concentrations and relatively low $^{10}\text{Be}$ concentrations of 400 atoms/g, resulting in low $^{10}\text{Be}/^{9}\text{Be}$ ratios of between 2 and $3 \times 10^{-8}$ in the uppermost 100 m. These data indicate that there are significant contributions from river runoff at the very surface and some admixture of waters of Pacific origin in the halocline layer below (see Section 4.2), possibly modified by interaction with shelf sediments in the Canada Basin. This is consistent with the Nd isotope composition of the same samples (Andersson et al., 2008) and with the distribution of Ba concentrations in the upper water column of the eastern Fram Strait (Taylor et al., 2003). At a water depth of 1300 m, the Be isotope signature is indistinguishable from Arctic deep waters.

The deep waters in the Arctic Ocean below 500 m water depth show a much smaller variability in $^{10}\text{Be}$ and $^9\text{Be}$ concentrations than observed in the uppermost 500 m. The $^{10}\text{Be}$ concentrations are mostly within 500 ± 100 atoms/g and there is only the waters around 2500 m water depth in the central Arctic Basin (sites #4 to #6) which show a small increase above this value followed again by a decrease to values of ~500 atoms/g, which is also observable in the data by Strobl et al. (1999) (Fig. 4). This depth level corresponds to the upper part of the Polar Deep Water, but in the deep Canada Basin the $^{10}\text{Be}$ concentrations at depths around 2500 m are relatively low (300–400 atoms/g). This may indicate a small difference in the $^{10}\text{Be}$ concentration between deep waters on either side of the Lomonosov Ridge.
which inhibits a direct exchange of Arctic deep waters. Overall $^{10}\text{Be}/^{9}\text{Be}$ ratios range between 5 and $8 \times 10^{-8}$ in the Arctic deep waters, which is similar to the western North Atlantic Ocean ($5-7 \times 10^{-8}$, Ku et al., 1990), consistent with the dominance of Atlantic waters.

Comparison of our Be isotope data with those of the deep western North Atlantic, the deep Arctic concentrations of both $^{10}\text{Be}$ and $^{7}\text{Be}$ only amount to $\sim 50\%$ of the values in the northern Atlantic Ocean. The reason is unclear. It is possible that removal processes in the deep Norwegian Greenland Seas occur, as indicated by highly variable $^{10}\text{Be}$ concentrations between 300 and 1200 atoms/g (Strobl et al., 1999). Most likely this is combined with the shielding effect of Arctic sea ice, which only allows low amounts of Be from continental or atmospheric sources to arrive in the deep basin.

### 4.4. Residence time of Beryllium in the Arctic Ocean

From the available new data the inventories of $^{10}\text{Be}$ and its residence time in the deep Arctic Ocean can be calculated. It must be noted that this is a simplified approach, which assumes steady state and requires that advective input of $^{10}\text{Be}$ equals advective output. It also does not take into account potential effects of boundary scavenging (Anderson et al., 1990), but in view of the relatively slow deep water mass exchange can provide a useful first order estimate. The inventories ($\text{In}^{10}\text{Be}$) were calculated from the measured water column profiles of $^{10}\text{Be}$ extrapolating for depths at which no data are available and range from 15 to $25 \times 10^7$ atoms/cm$^2$ (Table 2). It is noted that the large variability in the upper 150 m of the Arctic water column discussed above does not influence the inventory at the deep sites but clearly adds uncertainty to the shallow sites. The residence times of $^{10}\text{Be}$ ($t^{10}\text{Be}$) were then calculated as follows:

$$t^{10}\text{Be} = \frac{\text{In}^{10}\text{Be}}{\text{Flux}^{10}\text{Be}_{\text{atm}}}.\,$$

We have applied the mean annual depositional flux of $^{10}\text{Be}$ from the atmosphere ($\text{Flux}^{10}\text{Be}_{\text{atm}}$) published by Heikkilä et al. (2008) for 75°N (0.25 $\times 10^6$ atoms/cm$^2$ yr). The calculated residence times for the deep Arctic Ocean range from 590 to 980 years and average 800 years (Table 2). For the shallower sites nearer to the coast and closer to the Fram Strait the dominating advection term at these water depths prevents calculation of meaningful residence times. The annual depositional flux of $^{10}\text{Be}$ from the atmosphere is expected to vary with latitude and amount of precipitation but there are no local data available, which limits the precision of this approach. Earlier studies on the depositional flux of $^{7}\text{Be}$ have shown that the annual depositional flux at the same latitude could vary by a factor of two due to different amounts of precipitation (Baskaran, 1995). Thus, the residence times calculated in this way are at best expected to be realistic within a factor of 2 and in case the atmospheric flux has been higher than the Heikkilä et al. estimate, the residence time estimates in Table 2 would be upper limits. Further uncertainty arises from the fact that a part of the atmospheric fallout will be deposited on sea ice, a large fraction of which is exported to North Atlantic via the Fram Strait. This results in an underestimation of the residence times of $^{10}\text{Be}$ presented here, in particular in the Eurasian Basin and the central Arctic Ocean. In addition, most of the deep waters in the Arctic Ocean were advected from the Atlantic Ocean that has seen higher atmospheric deposition rates of $^{10}\text{Be}$, which thus would cause an overestimation of the inventory and thus also the residence times at the deep locations. However, as discussed above, the Atlantic waters have probably lost a significant amount of dissolved Be in the Nordic Seas, which suggests that this source of uncertainty is small.

A second possibility to estimate residence times combines the water column inventories with $^{10}\text{Be}$-fluxes to Arctic Ocean surface sediments (Table 3). $^{10}\text{Be}$ concentrations in Arctic Ocean sediments have been obtained for a number of locations in the Nansen and Amundsen Basins (Eisenhauer et al., 1994; Aldahan et al., 1997), on the Lomonosov Ridge (Spielhagen et al., 1997, 2004; Frank et al., 2008), in the Makarov Basin (Spielhagen et al., 2004), and in the Canada Basin (Sellén et al., in press). Normalization of the $^{10}\text{Be}$ concentrations to $^{230}\text{Th}_{xs}$ concentrations ($C^{(\text{230}\text{Th}_{xs})}$) allows the calculation of depositional $^{10}\text{Be}$-fluxes to the sediments ($\text{Flux}^{10}\text{Be}_{\text{sed}}$), which are corrected for any sediment redistribution effects at each location (Frank et al., 1996; Francois et al., 2004) (Table 3):

$$\text{Flux}^{10}\text{Be}_{\text{sed}} = \frac{P \cdot z \cdot C^{(\text{10}\text{Be})}}{C^{(\text{230}\text{Th}_{xs})}}.\,$$

whereby $P$ stands for the production of $^{230}\text{Th}$ in the water column (2.63 dpm/(cm$^2$ yr 1000 m water column)), $z$ is water depth and $C^{(\text{10}\text{Be})}$ (atoms/g) and $C^{(\text{230}\text{Th}_{xs})}$ (dpm/g) are the measured concentrations of $^{10}\text{Be}$ and $^{230}\text{Th}_{xs}$ in the surface sediments, respectively. These calculations are robust even if some surface sediment was lost during coring because only the ratios between $^{10}\text{Be}$ and $^{230}\text{Th}_{xs}$ are part of the equation and all surface samples are less than 5000 years old, which renders any potential uncertainties arising from decay correction of the $^{230}\text{Th}_{xs}$ activities negligible. Combined $^{230}\text{Th}_{xs}$ and $^{10}\text{Be}$ data are available for five locations in the Arctic Ocean that correspond to locations

### Table 2

<table>
<thead>
<tr>
<th>Location</th>
<th>Water depth (m)</th>
<th>$^{10}\text{Be}$ inventory (10$^7$ atoms/cm$^2$)</th>
<th>Residence time (yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>#4: LegII, Station #11</td>
<td>4039</td>
<td>24.0</td>
<td>880</td>
</tr>
<tr>
<td>#5: LegII, Station #21</td>
<td>4400</td>
<td>24.8</td>
<td>980</td>
</tr>
<tr>
<td>#6: LegII, Station #30</td>
<td>3985</td>
<td>18.3</td>
<td>730</td>
</tr>
<tr>
<td>#C: AWS 2000 Station #3</td>
<td>3850</td>
<td>14.9</td>
<td>590</td>
</tr>
<tr>
<td>#D: AWS 2000 Station #4</td>
<td>3894</td>
<td>15.0</td>
<td>600</td>
</tr>
</tbody>
</table>
of the 10Be water column profiles in this work (Eisenhauer et al., 1990; Bohrmann, 1991; Hentzschel, 1992; Schäper, 1994) (Table 3). These fluxes are on the order of or somewhat lower than the assumed atmospheric deposition rate. 10Be residence times (τ10Be) can then be calculated from the following equation:

\[
\tau_{10Be} = \frac{\text{Inv}(10Be)}{\text{Flux}_{10Be_{sed}}}
\]

Low apparent residence times of 80 and 100 years are calculated for the Fram Strait and on the Yermak Plateau, respectively (Table 3), consistent with higher particulate fluxes and associated scavenging of 10Be, but these numbers are most likely not realistic due to the strong influence of advection. The calculated residence times in the deep Arctic Ocean (Amundsen and Makarov Basins) are on the order of 500 years, consistent with a quasi-conservative behavior of Be due to very low particulate fluxes caused by the sea ice cover and the large distance from land. The data from the shallower location on the Lomonosov Ridge in the central Arctic Basin also confirm a relatively long local residence time of 130 years for the upper 1000 m of the water column.

The estimates of the residence times of Be obtained from the two approaches can be directly compared for the Nansen and Makarov Basins at water depths below 3600 m and are in good agreement. These estimates are limited by the fact that they do not take into account advective exchange of deep water masses and variations in high latitude atmospheric deposition rates of 10Be. Nevertheless they provide clear support for the use of Be isotopes as tracers for water mass mixing in the open Arctic Ocean because they demonstrate that only limited vertical exchange of Be occurs in the water column.

5. CONCLUSIONS

We have acquired the first set of dissolved 10Be and 9Be concentrations covering the entire water column and all major sub-basins of the Arctic Ocean. This also includes most of the inflow and outflow connections of the Arctic Ocean with the Atlantic and Pacific Oceans, as well as the four largest Arctic rivers (Ob, Yenisey, Lena and Mackenzie).

There is a pronounced variability of 9Be and 10Be concentrations in the surface and subsurface waters caused by different sources of water masses and the origin of freshwater. In the Eurasian Basin the Be isotopes show that surface waters are predominantly a mixture between Siberian river runoff and seawater, whereby, favoured by short residence times of the river waters on the Siberian shelves, a large amount of the riverine Be is transported across the wide shelves without being removed by scavenging. The waters of the cold halocline above the Atlantic layer either originating from the Siberian shelves but sometimes even advecting from the Canada Basin are characterized by low 10Be concentrations and relatively high 9Be concentrations resulting in minimum 10Be/9Be near 2 × 10^-8.

In pronounced contrast, the freshwater component in the surface waters of the open Canada Basin is dominated by river waters from the Mackenzie which experienced efficient scavenging on the shelves prior to reaching the open Canada Basins as reflected by low 9Be concentrations in the fresh uppermost 25 m. At the same time, the 10Be concentrations at the surface are significantly increased by a combination of longer residence times of the surface waters in the Canada Basin, thus allowing a larger contribution from atmospheric fallout and sea ice melting than in the Eurasian Basin, which results in high 10Be/9Be ratios in the uppermost 25 m. The waters of the cold halocline mainly consist of modified Pacific waters, which at depths between 125 and 150 m apparently have exchanged with the Chukchi shelf or have received contributions from Siberian river waters through sea ice formation processes. The Be isotope data are consistent with advection of Canada Basin halocline waters across the Arctic Basin until Fram Strait.

The deep waters of the Arctic Ocean below 500 m water depth do only show minor variations in Be isotope composition and are consistent with the Atlantic Ocean being the dominant source of these waters. The average oceanic residence time of Be in the deep Arctic Ocean is estimated to 800 years, which is similar to the global average residence time.
Despite the fact that all processes influencing Be isotopes in Arctic seawater cannot be disentangled based on a single data set, the first data presented here suggest that Be isotopes have a potential to serve as a tracer for the sources of freshwater inputs, exchange processes on the shelves and water mass mixing in the Arctic Ocean. There are, however, significant gaps of data and knowledge on the geochemical behavior of Be on the shelves and consequently its balance in shallow waters of the Arctic Ocean that need to be closed before Be isotopes can be applied reliably. The main focus of future research will thus be a detailed investigation of Be isotopes in the dissolved and particulate fractions of waters on the Canadian and Siberian shelves including the variability of the riverine inputs.

ACKNOWLEDGMENTS

This project was supported by the Swiss National Fonds, ETH Zurich and US-NSF OPP (9996337). We are grateful for the logistical support from the Captain and crew of the U.S.CGC POLAR STAR for their help in obtaining samples from the Canada Basin during the AWS 2000 Cruise, and Bettina Zimmermann, Larry Borum, and Sarah Trimble for their help with sample collection in the field. We also thank the captain and crew of I/B Oden and the Swedish Polar Research Secretariat (SPRS) during the Arctic Ocean 2001 and 2002 (AO-01 and AO-02) expeditions. Support from Peter Winsor, Anna Nikolopoulou, Johan Soderkvist and Leif Anderson during sampling on I/B Oden is gratefully acknowledged. The project was partly financed by the Swedish Research Council (VR grant #621-2001-2616). Dorothea Bauch is thanked for discussions. Fabienne Scheidele 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. Constructive and helpful reviews by M. Christl and two anonymous reviewers significantly improved the manuscript.

REFERENCES


Alldahan A., Nisg S., Possnert G., Backman J. and Leif Anderson during sampling on I/B Oden is gratefully acknowledged. The project was partly financed by the Swedish Research Council (VR grant #621-2001-2616). Dorothea Bauch is thanked for discussions. Fabienne Scheidele 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. Constructive and helpful reviews by M. Christl and two anonymous reviewers significantly improved the manuscript.


Frank M., Gersonde R., Rutgers van der Loeff M., Kuhn G. and Mangini A. (1996) Late Quaternary sediment dating and quantification of lateral sediment redistribution applying...


*Associate editor: S. Krishnaswarri*