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Research paper

# Taxon-specific epibenthic foraminiferal $\delta^{18}O$ in the Arctic Ocean: Relationship to water masses, deep circulation, and brine release

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ABSTRACT

We determined $\delta^{18}O_{Cib}$ values of live (Rose Bengal stained) and dead epibenthic foraminifera *Cibicidoides wuellerstorfi*, *Cibicides lobatulus*, and *Cibicides refulgens* in surface sediment samples from the Arctic Ocean and the Greenland, Iceland, and Norwegian seas (Nordic Sea). This is the first time that a comprehensive $\delta^{18}O_{Cib}$ data set is generated and compiled from the Arctic Ocean. For comparison, we defined Atlantic Water (AW), upper Arctic Bottom Water (uABW), and Arctic Bottom Water (ABW) by their temperature/salinity characteristics and calculated mean equilibrium calcite $\delta^{18}O_{equ}$ from summer sea-water $\delta^{18}O_w$ and *in situ* temperatures. As a result, in the Arctic environment we compensate for *Cibicidoides*- and *Cibicides*-specific offsets from equilibrium calcite of −0.35 and −0.55‰, respectively. After this taxon-specific adjustment, mean $\delta^{18}O_{Cib}$ values plausibly reflect the density stratification of principle water masses in the Nordic Sea and Arctic Ocean. In addition, mean $\delta^{18}O_{Cib}$ from AW not only significantly differs from mean $\delta^{18}O_{Cib}$ from ABW, but also $\delta^{18}O_{Cib}$ from within AW differentiates in function of provenience and water mass age. Furthermore, in shallow waters brine-derived low $\delta^{18}O_w$ can significantly lower the $\delta^{18}O_{Cib}$ of *Cibicides* spp. and thus $\delta^{18}O_{Cib}$ may serve as a paleobrine indicator. There is no statistically significant difference, however, between deeper water masses mean $\delta^{18}O_{Cib}$ of the Nordic Sea, and of the Eurasian and Amerasian basins, and no influence of low-$\delta^{18}O_w$ brines is recorded in Recent uABW and ABW $\delta^{18}O_{Cib}$ of *C. wuellerstorfi*. This may be due to dilution of a low-$\delta^{18}O_w$ brine signal in the deep sea, and/or to preferential incorporation of relatively high-$\delta^{18}O_w$ brines from high-salinity shelves. Although our data encompass environments with seasonal sea-ice and brine formation supposed to ultimately ventilate the deep Arctic Ocean, $\delta^{18}O_{Cib}$ from uABW and ABW do not indicate negative excursions. This may challenge hypotheses that call for enhanced Arctic brine release to explain negative benthic $\delta^{18}O$ spikes in deep-sea sediments from the late Pleistocene North Atlantic Ocean.



## 1. Introduction

### 1.1. Rationale

The recent rapid retreat of sea ice in the Arctic Ocean draws much public attention on the critical role of the Arctic in climate change. So it is a matter of ongoing and intense oceanographic research of how the rate of meltback and changing sea-ice volume affects water mass formation and circulation in the Arctic basins and the connected Nordic Sea (e.g., Rabe et al., 2009, 2013; Bauch et al., 2011a,b; Dodd et al., 2012). Sea-ice formation and brine rejection takes place on shallow continental shelves during the winter season (Aagaard et al., 1985). It is believed that today much of the deep Arctic Ocean is ventilated by brines (e.g., Jones et al., 1995).

One of the tracers used in oceanography to differentiate between melt water from surrounding continental ice and sea ice, river discharge and direct oceanic precipitation in contrast to marine water is the stable isotopic composition of oxygen in water ($\delta^{18}O_w$). Consequently many water samples have been analyzed for their stable oxygen isotopic composition over the last decades (cf. Schmidt et al., 1999). Similarly, much attention was directed to the Arctic's late Pleistocene and Cenozoic evolution (e.g., Thiede and Johannessen, 2008). In paleoceanography, one of the proxies for temperature and density of ancient water masses is the stable isotopic composition of foraminiferal calcite (e.g., Lynch-Stieglitz et al., 2006). Foraminifera calcify their tests in a constant relationship to $\delta^{18}O_w$ depending on temperature. Since $\delta^{18}O_w$ relates to salinity and the global continental ice volume, ideally $\delta^{18}O_{Cib}$ of benthic foraminiferal calcite can be used to reconstruct paleotemperatures and densities of former deep and bottom water masses. Unfortunately it is not an easy task to calibrate $\delta^{18}O_{Cib}$ via $\delta^{18}O_w$ with temperature of a given water body. The simpler part is calculating the temperature dependent equilibrium $\delta^{18}O_{equ}$ of inorganically precipitated calcite, though still a matter of discussion, since the temperature range below 10 °C is not well-defined during equilibrium calcification (Grossman, 2012). It is, however, more difficult to calibrate $\delta^{18}O_{Cib}$ from epibenthic

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deep-sea foraminifera with overlying bottom water $\delta^{18}O_w$ and temperatures (Shackleton, 1974; Matsumoto and Lynch-Stieglitz, 1999). Furthermore, contemporaneously recovered surface sediment and bottom water samples from the Arctic Ocean, including hydrographic data, such as salinity and temperature, are very rare.

Here we associate mean $\delta^{18}O_w$ values of the Arctic Ocean's principal water masses, collected during boreal summer seasons, with independently sampled epibenthic foraminiferal $\delta^{18}O_{Cib}$ values from the ocean floor bathed by these water masses. As a result we obtain a water mass specific $\delta^{18}O_{Cib}$ of Recent epibenthic foraminifera, which then can be used as modern analog for the interpretation of late Pleistocene benthic foraminiferal records from the Arctic Ocean and Nordic Sea; for example, to better evaluate suggestions, whether late Pleistocene epibenthic *Cibicidoides wuellerstorfi* and *Cibicides* spp. $\delta^{18}O_{Cib}$ values in the Nordic Sea have been influenced by brines (Vidal et al., 1998; Dokken and Jansen, 1999; Waelbroeck et al., 2006; Meland et al., 2008; Thornalley et al., 2010).

We present new $\delta^{18}O_{Cib}$ values of live epibenthic foraminifera *C. wuellerstorfi* and *Cibicides* spp. from the Arctic Ocean and the Nordic Sea in comparison to calcite equlibrium $\delta^{18}O_{equ}$, calculated from summer $\delta^{18}O_w$ water values and *in situ* temperature. It is the first time that a comprehensive $\delta^{18}O_{Cib}$ data set is generated and compiled from the Arctic Ocean. The $\delta^{13}C$ values of live *C. wuellerstorfi* and *Cibicides* spp. and $\delta^{13}C_{DIC}$ values are discussed elsewhere (Mackensen, 2013).

### 1.2. Oceanographic settings

The Arctic Ocean north of Fram Strait contains only about 1% of the World Ocean's water but represents 3% of its area. This is because almost 70% of the surface area of the Arctic Ocean is occupied by shallow continental shelves with average water depths between about 40 and 200 m, the width of which along the Eurasian coast exceeds 800 km in most places (Tomczak and Godfrey, 1994). Four principal water masses can be distinguished: Arctic Bottom Water (ABW), upper ABW (uABW), Atlantic Water (AW), and Arctic Surface Water (ASW).

ABW and uABW fill the deep basins up to about 900 m water depth. In the Eurasian basins, ABW sometimes is called Eurasian Basin Bottom Water to diffentiate from Canadian Basin Bottom Water, which is warmer (Fig. 1). Shallower than about 1500 m water depth in the Amerasian and 1300 m in the Eurasian basins, the intermediate uABW can be distinguished from ABW (Fig. 1). The uABW also is called upper Polar Deep Water or Arctic Intermediate Water. Here we use the $\sigma_{0.5}$-isopyc $= 30.444$ kg m$^{-3}$ to separate ABW from uABW (Jeansson et al., 2008). ABW is a mixture of water from (i) the inflow of Atlantic water over the Barents Sea shelf in the Eurasian basins, (ii) contributions from the Arctic shelves, and (iii) recirculated Greenland Sea Deep Water (GSDW) and Norwegian Sea Deep Water (NSDW) entering the central Arctic Ocean through Fram Strait (Jones et al., 1995).

The formation of ABW generally involves GSDW and Brine-enriched Shelf Water (BSW) from the Arctic shelves (Aagaard et al., 1985; Rudels and Quadfasel, 1991; Jones et al., 1995; Jones, 2001). GSDW is formed during winter in the central Greenland Sea in short-lived and small-scale events, during which intense vertical convection occurs and surface water sinks to the bottom. Formation of GSDW occurs in mainly open water (Rudels and Quadfasel, 1991; Hansen and Østerhus, 2000). BSW is formed on the Arctic shelves (Aagaard et al., 1985). There, during winter in coastal polynyas, salinization by sea-ice formation may produce dense but, due to riverine fresh water, relatively low-salinity and low-$\delta^{18}O$ brine (cf. Bauch et al., 2012). However, only on high-salinity shelves, such as the salty Barents and Kara Seas, brines are dense enough to contribute to deep-water formation (Bauch et al., 1995; Shapiro et al., 2003; Ivanov et al., 2004), and these brines may show

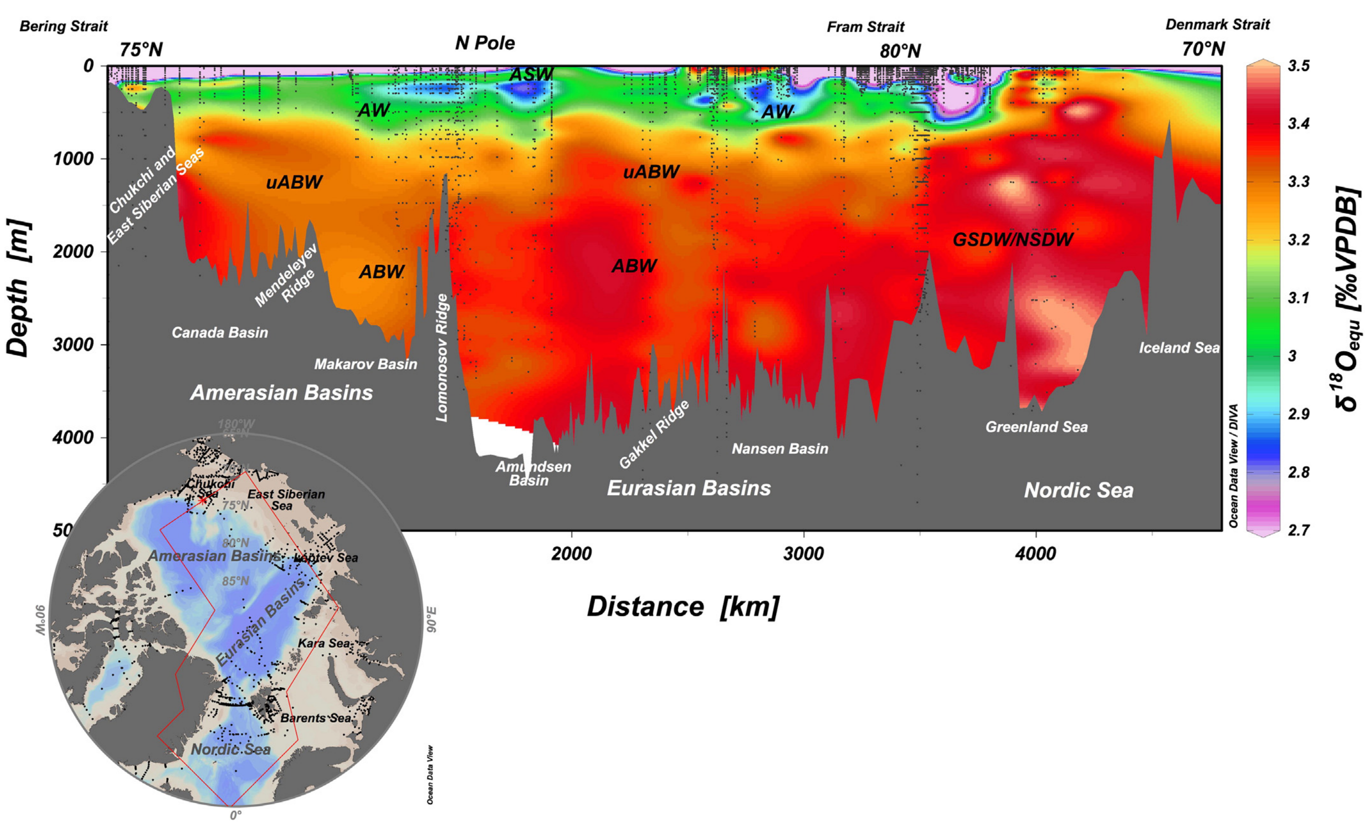


**Fig. 1.** Distribution of $\delta^{18}O_{equ}$, calculated from temperature and $\delta^{18}O_w$ after Matsumoto and Lynch-Stieglitz (1999) along a schematic section through the Arctic Ocean from Bering Strait broadly along Greenwich Meridian via the North Pole through Fram Strait into Greenland and Iceland Seas to Denmark Strait. Temperature and $\delta^{18}O_w$ are from different sources measured during summer seasons as compiled in Schmidt et al. (1999), Rabe et al. (2009), and from this study. ODV software is used to smooth and project values onto the section plane according to latitude and water depth (Schlitzer, 2011). Principle water masses are indicated: Arctic Surface Water (ASW), Atlantic Water (AW), Arctic Bottom Water (ABW), and upper ABW (uABW).

higher $\delta^{18}O_w$ values, because less low-$\delta^{18}O_w$ river water is incorporated. Jones et al. (1995) conclude that dense water, consisting of brines and primarily of waters entrained from the AW, contributes most to ABW. The second most important contributor to ABW is inflow of Atlantic water over the Barents Sea shelf, and least important is inflow of NSDW through Fram Strait (Jones et al., 1995). The residence time of ABW varies between about 250 to 450 years with highest ages in the more isolated deep Canada Basin (Schlosser et al., 1994; Macdonald and Bewers, 1996).

Above ABW, AW enters the Arctic through Fram Strait and occupies the depth range between about 700 and 150 m. The AW, also known as Atlantic Layer, is commonly defined as water with temperatures above 0 °C (Rudels et al., 2004). We chose $\sigma_0 < 27.97$ kg m$^{-3}$ to separate from uABW below and $\sigma_0 = 27.70$ kg m$^{-3}$ from ASW above (Jeansson et al., 2008), (Fig. 2). Within the uppermost layer of roughly 200 m of ASW it can be distinguished between a 25 to 50 m thick surface layer and a subsurface layer of Arctic Ocean halocline water below (Aagaard et al., 1985, 1987; Tomczak and Godfrey, 1994; Jones, 2001). The renewal age of the Atlantic layer and the halocline water is estimated at about 25 and about 10 years, respectively (Östlund, 1982; Macdonald and Bewers, 1996). At present, the central Arctic Ocean is permanently covered by sea ice, but the vast area of continental shelves surrounding the central basins is free of sea ice during summer. Most of the sea ice is formed during winter on the shelves (Aagaard et al., 1985). Since we will discuss the stable isotopic composition of biogenic calcite it is important to note that the calcite lysocline, Arctic Ocean wide, is situated below 4 km water depth, except at the shelf break depth in the Canadian basin where high nutrient and low dissolved oxygen concentrations suggest decomposition of organic matter (Jutterström and Anderson, 2005).

## 2. Material and methods

### 2.1. Water samples

On *Polarstern* cruises ARK XIX/1 in 2003, and ARK XXIII/2 in 2008, and *Heincke* cruise 153 in 2002, at 162 stations across the Fram Strait and in adjacent Greenland and Svalbard fairways, water was sampled with the aid of a water-sampling rosette to determine its $\delta^{18}O_w$ (Fig. 2). For salinity and temperature profiles obtained during *Polarstern* cruises in the Fram Strait see also Rabe et al. (2009) and Dodd et al. (2012). Samples of 100 ml water from 10 l Niskin bottles were drawn into glass vials, sealed with wax, and stored at 4 °C temperature until further treatment on shore. In the laboratory, 7 ml of water were equilibrated in 13 ml headspace with $CO_2$ gas by using a Finnigan equilibration device. Oxygen isotope equilibrium in the $CO_2$–$H_2O$ system was attained by shaking for 430 min at 20 °C. The equilibrated gas was purified and transferred to a Finnigan Delta-S gas mass spectrometer.

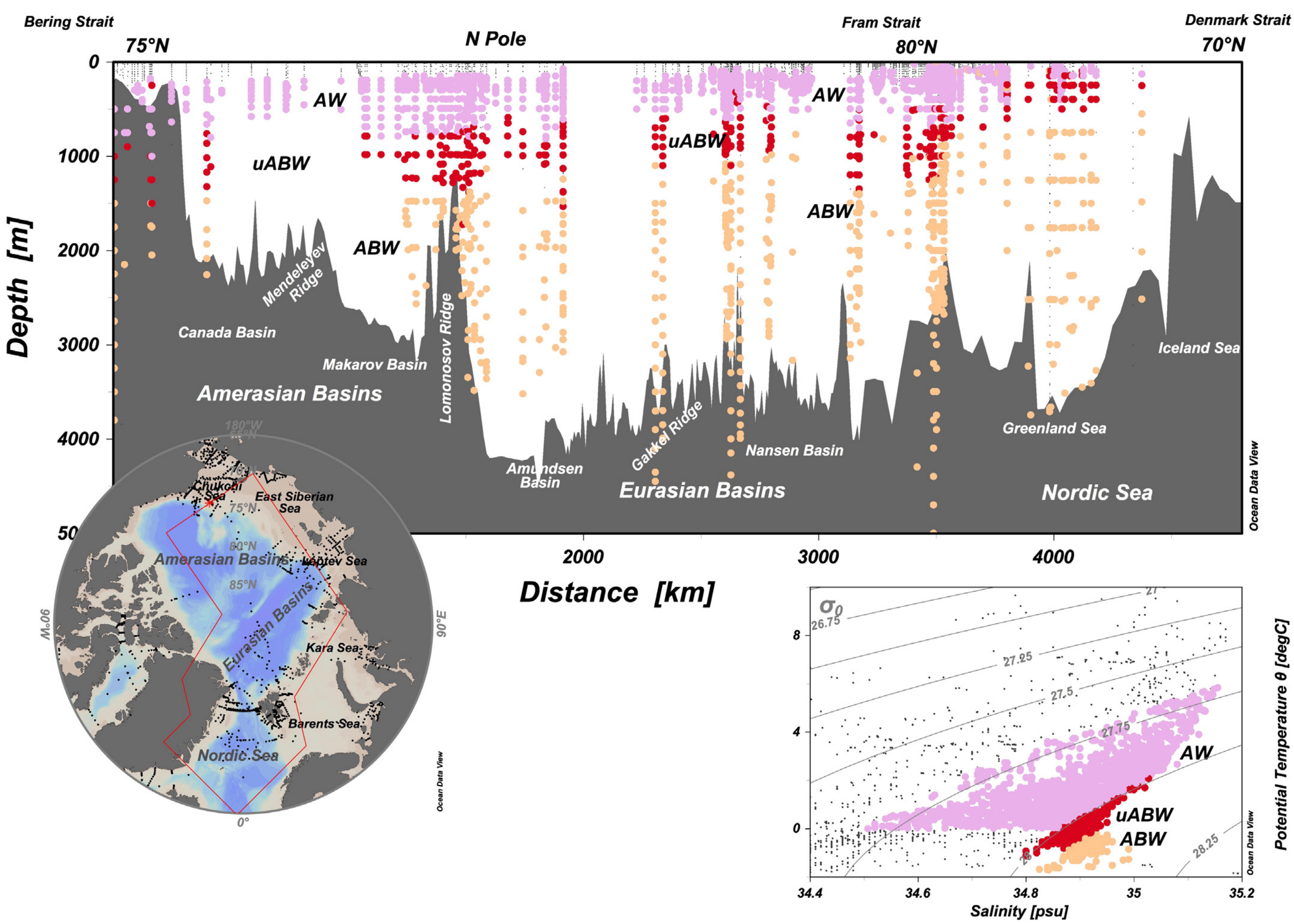


**Fig. 2.** Distribution of water masses as defined by potential density along a schematic section through the Arctic Ocean from Bering Strait broadly along Greenwich Meridian via the North Pole through Fram Strait into Greenland and Iceland Seas to Denmark Strait. Potential density is calculated from summer potential temperature and salinity from different sources as compiled in Schmidt et al. (1999) and Rabe et al. (2009). ODV software is used to calculate water mass patches and project values onto the section plane according to latitude and water depth (Schlitzer, 2011). Water masses are: Atlantic Water (AW), Arctic Bottom Water (ABW), and upper ABW (uABW).

Sample preparation and isotope measurements were calibrated against Vienna Standard Mean Ocean Water (VSMOW) and Standard Light Antarctic Precipitation (SLAP) standard waters. At least two replicates (including preparation and measurement) were run for each oxygen isotope determination. Results are reported in δ-notation versus VSMOW with a reproducibility of better than 0.03‰. In addition, we extracted $\delta^{18}O_w$, salinity, and temperature data from the global seawater oxygen-18 database (Schmidt et al., 1999). For detailed references see Appendix Table 1. Water $\delta^{18}O$ data were gridded with Data Interpolating Variational Analysis (DIVA; http://modb.oce.ulg.ac.be/projects/1/diva), and mapped with Ocean Data View (ODV) (Schlitzer, 2011).

For comparison with $\delta^{18}O_{Cib}$ (‰ VPDB), $\delta^{18}O_{equ}$ (‰ VPDB) equilibrium calcite values were calculated using summer (JAS) $\delta^{18}O_w$ (‰ VSMOW) and in situ temperatures (°C) from CTD measurements at the time of sampling. To calculate $\delta^{18}O_{equ}$, we used $\delta^{18}O_{equ} = -0.2004\ T + 3.2486 + (\delta^{18}O_w - 0.27)$. This linear equation after Matsumoto and Lynch-Stieglitz (1999) is derived from data of Kim and O'Neil (1997) on equilibrium oxygen isotopic fractionation of inorganically precipitated calcite, where T is in degree centigrade, and $\delta^{18}O_w$ (VSMOW) is transferred to VPDB by subtracting 0.27 (Hut, 1987).

For the sake of comparability with classic calibration studies we calculated $\delta^{18}O_{equ}$ after Shackleton's (1974) quadratic approximation of the data of O'Neil et al. (1969). Accordingly, we applied a correction of −0.20 to adjust $\delta^{18}O_w$ from SMOW to the PDB scale after Craig (1965) (cf. Bemis et al., 1998; Grossman, 2012). We similarly compared to values derived from McCorkle's et al. (1990) version of O'Neil et al. (1969) and the empiric calibration based on planktic foraminiferal cultures of Erez and Luz (1983). However, although the offsets varied significantly, no significant difference occurred in the relationship between calculated mean $\delta^{18}O_{equ}$ of defined water masses.

### 2.2. Sediment samples

Between 1987 and 2013 at 636 sites in the Arctic Ocean and Nordic Sea, surface sediment samples were recovered on 12 cruises with *PRV Polarstern*, three cruises with *PRV Araon*, and on one cruise with RV *Heincke*. 347 and 246 samples, from deeper than 150 m water depth, yielded dead and live specimens of epibenthic species, respectively, and were suitable for isotopic analysis in this study (App. Table 2; Fig. 4). Immediately after recovery on board ship, all samples, except those from *PRV Araon*, were preserved in Rose Bengal stained ethanol (1 g/L). On shore, samples were wet sieved through 63 μm, 125 μm, and 2 mm sized meshes. After drying at temperatures less than 60 °C, epibenthic foraminifera from the larger than 125 μm fraction, were selected for stable isotope analyses. One to three specimens of *C. wuellerstorfi* were separated and transferred, via the automated carbonate preparation device Kiel Carbo, into Finnigan MAT251 or MAT253 mass spectrometers. If no *C. wuellerstorfi* were recognized, i.e. on the shelves and upper slopes <750 m water depth in the Arctic Ocean and <1000 m in the Nordic Sea, closely related species *Cibicides lobatulus* and *Cibicides refulgens* (in the following lumped under *Cibicides* spp.) were analyzed (cf. Mackensen, 1997). No other species were considered in this study. The mass spectrometers were calibrated via international standard NBS19 to the PDB scale, and results are given in δ-notation versus VPDB. The precision of $\delta^{18}O$ measurements, based on an internal laboratory standard (Solnhofen limestone) measured over a one-year period together with samples, was better than ±0.08‰. In addition, we included 12 *C. wuellerstorfi* $\delta^{18}O$ values from unstained surface sediment samples taken in 1994 in the Mendeleyev Ridge area of the Amerasian basins (Poore et al., 1999). Foraminiferal $\delta^{18}O$ data were gridded with DIVA and mapped with ODV (Schlitzer, 2011).

## 3. Results

### 3.1. Water $\delta^{18}O_{equ}$ distribution

We separated three principal Arctic water masses, Atlantic Water (AW), upper Arctic Bottom Water (uABW), and ABW by their potential densities (Fig. 2) and projected accordingly grouped patches on a schematic section through the Arctic Ocean from the Bering Strait broadly along Greenwich Meridian via the North Pole through Fram Strait into Greenland and Iceland Seas to Denmark Strait (Fig. 2). Since for the purpose of this study it is not needed, we did not adjust Arctic water mass definitions to those typically used in the Nordic Sea. We then subdivided the area of investigation into four geographic regions: (i) the Nordic Sea (NOR), i.e. Greenland, Iceland and Norwegian seas, including surrounding continental margins; (ii) the Eurasian basins and margin, divided at 90°E into a western (WEU) and an eastern (EEU) part; and (iii) the Amerasian basins and eastern margins (AME). Finally we calculated mean $\delta^{18}O_{equ}$ values for each of these boxes (Table 1).

Means of summer $\delta^{18}O_{equ}$ values clearly reflect the general hydrographic geometry of the Nordic Sea and the Arctic basins (Table 1, Fig. 3). Mean $\delta^{18}O_{equ}$ values of shallow water masses are in all areas lower than corresponding mean values of deeper and denser water masses. As indicated by the standard errors of means, only the density stratification between uABW and ABW in the Amerasian basins, is statistical not significantly reflected in the means of summer $\delta^{18}O_{equ}$ values (Table 1, Fig. 3).

### 3.2. Foraminiferal $\delta^{18}O_{Cib}$ distribution

We plotted $\delta^{18}O_{Cib}$ values (App. Table 2) onto a stereographic projection of the northern Hemisphere north of 65°N (Fig. 4). In addition, we assigned $\delta^{18}O_{Cib}$ values according to geographic position and differentiated between shallow, intermediate, and deep water depths, coinciding with the water depth range occupied by AW, uABW, and ABW, as given by the specific water mass density fields defined by $\delta^{18}O_{equ}$ (Table 2; Fig. 3). This allowed us to roughly calculate water mass specific $\delta^{18}O_{Cib}$ means. Generally, on the continental shelf breaks and upper slopes off Greenland and Svalbard, mean benthic calcite $\delta^{18}O_{Cib}$ is low, whereas in the deep Nordic Sea and the western Eurasian basins, means are higher by 0.5 and 0.8‰, respectively (Figs. 3, 4; Table 2). In the eastern Eurasian and Amerasian basins, mean $\delta^{18}O_{Cib}$ values of water masses on upper slopes, i.e. AW and uABW, do not significantly differ from values of ABW on the lower slope and deeper ocean floor below (Table 2; Figs. 3, 4).

## 4. Discussion

### 4.1. $\delta^{18}O_{equ}$ versus $\delta^{18}O_{Cib}$ distributions

In the modern Nordic Sea *C. wuellerstorfi* does not thrive in water depths shallower than about 1000–1200 m, in detail depending on the trophic situation (e.g., Belanger and Streeter, 1980; Sejrup et al., 1980; Mackensen et al., 1985), and in the Arctic Ocean no Recent *C. wuellerstorfi* was found shallower than about 800 m water depth (e.g., Green, 1960; Vilks, 1969, 1989; Lagoe, 1977; Bergsten, 1994; Wollenburg and Mackensen, 1998; Poore et al., 1999). Consequently all $\delta^{18}O_{Cib}$ values from the shallowest boxes, i.e. <800 m water depth, are obtained from either *C. lobatulus* or *C. refulgens*, the only two *Cibicides* species we analyzed. Since the mid-depth range of the Arctic Ocean and Nordic Sea continental slopes usually is populated by *C. wuellerstorfi* and we always analyzed *C. wuellerstorfi* if available, the $\delta^{18}O$ means of these areas (boxes >700/800 m) are obtained from *C. wuellerstorfi*.

Comparison of mean $\delta^{18}O_{Cib}$ values with mean $\delta^{18}O_{equ}$ values revealed a systematic difference in taxon-specific deviations. In fact, the differences between all (live and dead) *Cibicides* spp. $\delta^{18}O$ means and AW $\delta^{18}O_{equ}$ means, except one (WEU), are by 0.2‰ larger than

**Table 1**
Mean $\delta^{18}O_{equ}$ values after Matsumoto and Lynch-Stieglitz (1999) as derived from data of Kim and O'Neil (1997), $\delta^{18}O_w$ and *in-situ* temperature of principle water masses. Means with standard deviations and standard errors are calculated according to water mass definitions as given in the text and oceanic region.

| | $\delta^{18}O_{equ}$ [‰VPDB] | n | Std. err | $\delta^{18}O_w$ [‰VSMOW] | Std. err | *In situ* T [°C] | Std. err |
|---|---|---|---|---|---|---|---|
| *Nordic Sea (65°–80°N, 20°W–40°E), incl. E-Greenland, W-Spitsbergen and W-Barents Sea slopes and shelf breaks* | | | | | | | |
| "AW" | 2.77 ± 0.25 | 649 | ±0.010 | 0.26 ± 0.15 | ±0.006 | 2.37 ± 1.28 | ±0.050 |
| "uABW" | 3.22 ± 0.13 | 172 | ±0.010 | 0.26 ± 0.06 | ±0.005 | 0.12 ± 0.55 | ±0.042 |
| "ABW" | 3.42 ± 0.26 | 343 | ±0.014 | 0.30 ± 0.24 | ±0.013 | −0.73 ± .19 | ±0.010 |
| *W-Eurasian Basins (80°–90°N, 20°W–90°E), incl. N-Greenland and N-Svalbard slopes and shelf breaks* | | | | | | | |
| AW | 2.97 ± 0.16 | 314 | ±0.009 | 0.22 ± 0.10 | ±0.006 | 1.13 ± 0.78 | ±0.044 |
| uABW | 3.23 ± 0.11 | 72 | ±0.013 | 0.25 ± 0.07 | ±0.008 | 0.00 ± 0.42 | ±0.050 |
| ABW | 3.35 ± 0.12 | 160 | ±0.009 | 0.25 ± 0.11 | ±0.008 | −0.61 ± .15 | ±0.012 |
| *E-Eurasian Basins (70°–90°N, 90°–140°E), incl. Kara and Laptev Seas slopes and shelf breaks* | | | | | | | |
| AW | 2.96 ± 0.13 | 291 | ±0.008 | 0.22 ± 0.05 | ±0.003 | 1.22 ± 0.67 | ±0.040 |
| uABW | 3.24 ± 0.06 | 111 | ±0.006 | 0.24 ± 0.04 | ±0.003 | −0.10 ± .27 | ±0.026 |
| ABW | 3.35 ± 0.03 | 132 | ±0.003 | 0.26 ± 0.03 | ±0.003 | −0.59 ± .08 | ±0.007 |
| *Amerasian Basins (70°–90°N, 140°E–130°W), incl. E-Siberian and Chukchi Seas slopes and shelf breaks* | | | | | | | |
| AW | 3.04 ± 0.18 | 215 | ±0.012 | 0.20 ± 0.17 | ±0.012 | 0.69 ± 0.42 | ±0.028 |
| uABW | 3.30 ± 0.14 | 43 | ±0.021 | 0.29 ± 0.14 | ±0.021 | −0.14 ± .13 | ±0.020 |
| ABW | 3.33 ± 0.05 | 48 | ±0.007 | 0.29 ± 0.05 | ±0.007 | −0.31 ± .12 | ±0.017 |

the difference between *C. wuellerstorfi* $\delta^{18}O$ and the $\delta^{18}O_{equ}$ of uABW and ABW (Tables 1, 2). Consequently, we adjusted *Cibicides* spp. and *C. wuellerstorfi* $\delta^{18}O$ by −0.55‰ and −0.35‰, respectively (Fig. 5). This resulted in $\delta^{18}O_{Cib}$ mean values that coincide within ±0.13‰ with mean $\delta^{18}O_{equ}$, except for AW on the western Eurasian basins margins (WEU), where corrected $\delta^{18}O_{Cib}$ means are much lower than mean $\delta^{18}O_{equ}$ (Fig. 5). As mentioned earlier, the fact that we used different equilibrium equations for comparison did not significantly change the order and relationships between values but would affect the absolute correction values.

In previous studies $\delta^{18}O_{Cib}$ has shown a great variety of deviations from $\delta^{18}O_{equ}$, varying from none (Bemis et al., 1998; Matsumoto and Lynch-Stieglitz, 1999; Mackensen et al., 2001) to up to +1.0‰ (Shackleton, 1974; Woodruff et al., 1980; Belanger et al., 1981; McCorkle et al., 1990). Much of the reported variability is due to the use of different equilibrium calibrations (cf. Bemis et al., 1998; Grossman, 2012). For example, a calibration of subantarctic eastern Atlantic *Cibicidoides* spp. needed no taxon-specific adjustment if calculated after Matsumoto and Lynch-Stieglitz (1999) but a correction of 0.64‰ if calculated after Erez and Luz (1983) (Mackensen et al., 2001).

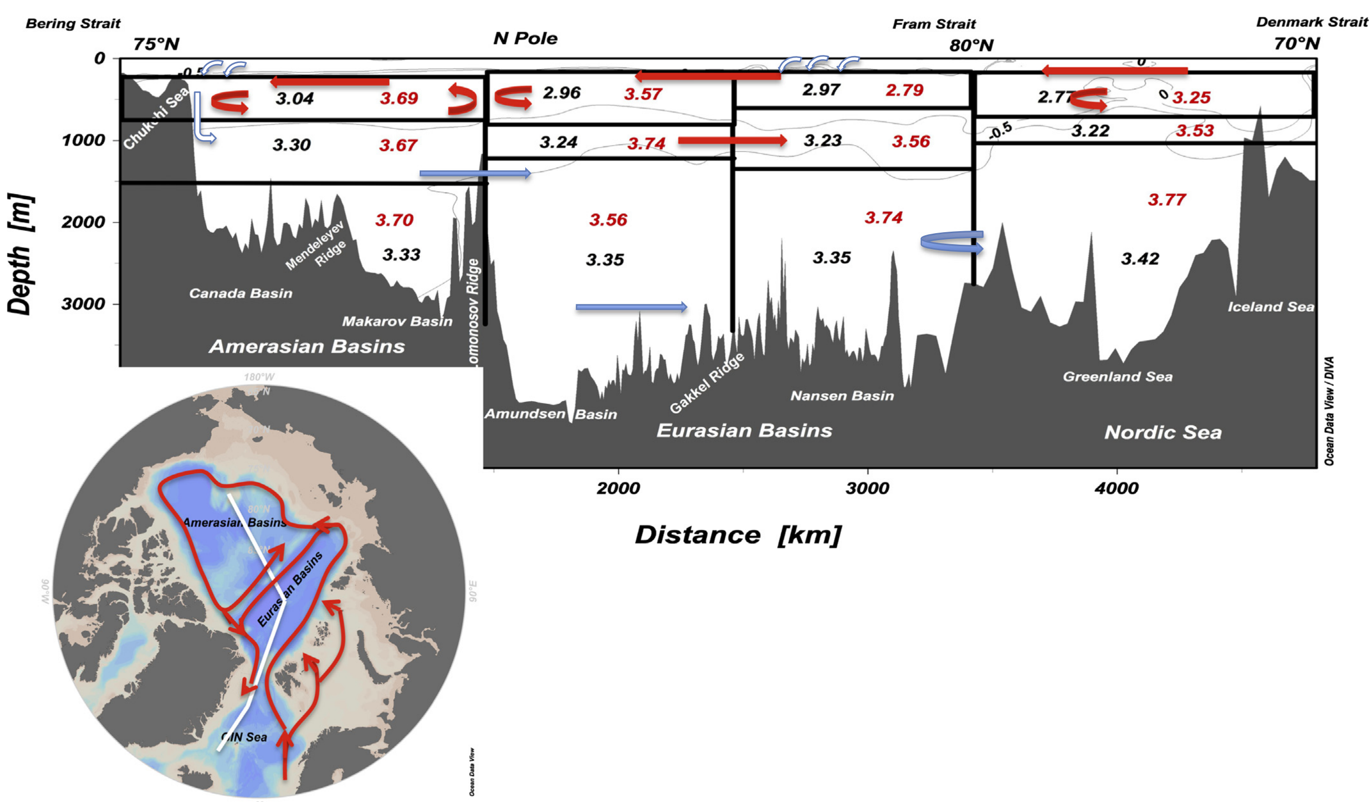


**Fig. 3.** Schematic section through the Arctic Ocean from Bering Strait broadly along Greenwich Meridian via the North Pole through Fram Strait into Greenland and Iceland Seas to Denmark Strait. The 0° and −0.5 °C isotherms are indicated. Boxes as discussed in the text and Tables 1 and 2 are given with mean $\delta^{18}O_{equ}$ (black) and live epibenthic $\delta^{18}O_{Cib}$ (red) values. Red and blue arrows schematically indicate generalized flow of Atlantic Water (AW) and deep water. White arrows indicate potential admixture of brine-enriched Arctic Surface Water and AW. Inlet map gives course of section plane and generalized circulation of AW, entering the Arctic Ocean through the eastern Fram Strait and the Barents and Kara seas. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

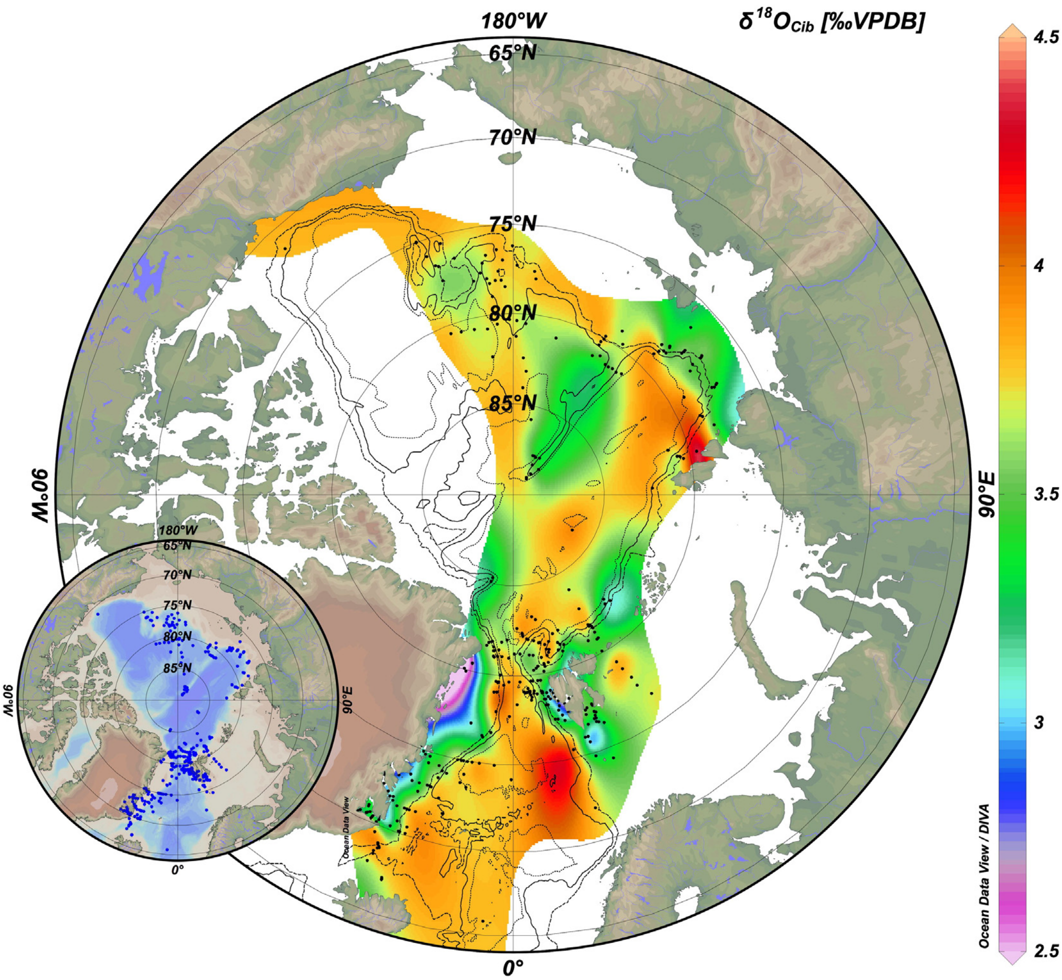


**Fig. 4.** $\delta^{18}O_{Cib}$ values of all (empty and Rose-Bengal stained) tests of *Cibicidoides wuellerstorfi* and *Cibicides* spp. (App. Table 2) from water depth > 150 m in a stereographic projection north of 65°N. Foraminiferal $\delta^{18}O$ data were gridded with DIVA and mapped with ODV (Schlitzer, 2011).

A taxon-specific offset may also be influenced by regional water mass specific factors, such as the carbonate ion concentration (Spero et al., 1997). So, in the Mackensen et al. (2001: Fig. 13) calibration of subantarctic eastern Atlantic $\delta^{18}O_{Cib}$ versus $\delta^{18}O_{equ}$, a possible dependence on $[CO_3^{2-}]$ of water masses was indicated. There a lower $\delta^{18}O_{Cib}$ and stronger deviation from equilibrium in North Atlantic Deep Water (higher pH and $[CO_3^{2-}]$) than in lower Circumpolar Deep Water (lower pH and $[CO_3^{2-}]$) would agree with the suggested decrease of $\delta^{18}O_{Cib}$ with increasing sea water $[CO_3^{2-}]$ (Spero et al., 1997). In the Arctic realm higher *Cibicides* spp. $\delta^{18}O_{Cib}$ corresponds to stronger deviation from $\delta^{18}O_{equ}$ in AW

**Table 2**
Mean $\delta^{18}O$ of live, i.e. Rose-Bengal stained, and all, i.e. empty and Rose-Bengal stained tests of *Cibicidoides wuellerstorfi* and *Cibicides* spp. in areas under investigation. Means with standard deviations and standard errors are calculated from boxes arranged according to water mass distribution and oceanic region.

| $\delta^{18}O$ [‰ VPDB] | Live | n | Std err | All | n | Std err |
|---|---|---|---|---|---|---|
| *Nordic Seas (65°–80°N, 20°W–40°E)* | | | | | | |
| "AW" 150–750 m | 3.25 ± 0.45 | 43 | ±0.069 | 3.22 ± 0.45 | 59 | ±0.059 |
| "uABW" 750–1100 m | 3.53 ± 0.15 | 8 | ±0.054 | 3.57 ± 0.16 | 10 | ±0.049 |
| "ABW" >1100 m | 3.77 ± 0.22 | 53 | ±0.030 | 3.78 ± 0.22 | 57 | ±0.028 |
| *W-Eurasian Basins (80°–90°N, 20°W–90°E)* | | | | | | |
| AW 150–700 m | 2.79 ± 0.45 | 15 | ±0.116 | 2.95 ± 0.48 | 20 | ±0.106 |
| uABW 700–1300 m | 3.56 ± 0.36 | 19 | ±0.082 | 3.52 ± 0.41 | 22 | ±0.088 |
| ABW >1300 m | 3.74 ± 0.25 | 30 | ±0.046 | 3.76 ± 0.24 | 41 | ±0.038 |
| *E-Eurasian Basins (70°–90°N, 90°–140°E)* | | | | | | |
| AW 150–800 m | 3.57 ± 0.24 | 11 | ±0.071 | 3.66 ± 0.35 | 13 | ±0.097 |
| uABW 800–1200 m | 3.74 ± 0.24 | 2 | ±0.172 | 3.47 ± 0.29 | 5 | ±0.129 |
| ABW >1200 m | 3.56 ± 0.36 | 7 | ±0.138 | 3.66 ± 0.32 | 15 | ±0.082 |
| *Amerasian Basins (70°–90°N, 140°E–130°W)* | | | | | | |
| AW 200–750 m | 3.69 | 1 | | 3.75 ± 0.13 | 3 | ±0.076 |
| uABW 750–1500 m | 3.67 ± 0.15 | 17 | ±0.037 | 3.70 ± 0.16 | 21 | ±0.036 |
| ABW >1500 m | 3.70 ± 0.23 | 37 | ±0.037 | 3.69 ± 0.27 | 48 | ±0.039 |

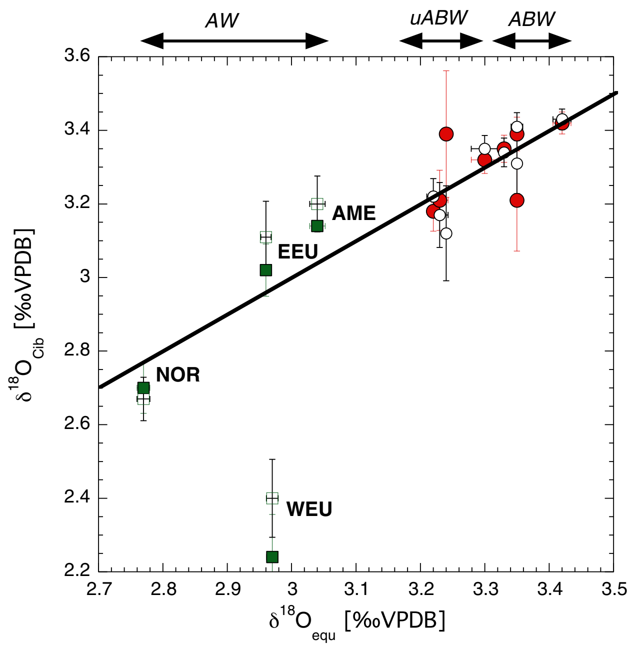


**Fig. 5.** Means of $\delta^{18}O_{Cib}$ of live (bold red dots) and all (open circles) *Cibicidoides wuellerstorfi* and *Cibicides* spp. (filled and open green squares) plotted versus summer mean $\delta^{18}O_{equ}$ as defined in Tables 1 and 2 and adjusted by −0.35 and −0.55‰. Bold line is slope of one. Error bars give standard errors. NOR, WEU, EEU and AME give values of Atlantic Water (AW) in Nordic Sea, western Eurasian basins, eastern Eurasian basins, and Amerasian basins, respectively. Range of $\delta^{18}O_{equ}$ values of principal water masses AW, upper Arctic Bottom Water (uABW) and ABW is indicated at the top above x-axis. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

compared to *C. wuellerstorfi* in ABW, which then would suggest a lower $[CO_3^{2-}]$ in shallower water than in ABW. Indeed, this agrees with the enhanced solubility of $CO_2$ in cold arctic surface water and the role of seasonally ice-free shelves and polynyas as a strong sink for atmospheric $CO_2$ (e.g., Yager et al., 1995; Anderson et al., 2004).

However, an enrichment of $^{18}O$ in live *C. wuellerstorfi* during calcification compared to equilibrium inorganic calcite precipitation, to the best of our knowledge, was not yet reported. In the classic study of Belanger et al. (1981) from the Nordic Sea, an offset of about 1‰ from equilibrium is suggested. Their $\delta^{18}O_{Cib}$ of *C. wuellerstorfi* is precisely in the range we measured but their $\delta^{18}O_{equ}$ was calculated at 4.8‰ using a $\delta^{18}O_w$ of 0.1‰ and the Epstein et al. (1953) paleotemperature equation. If instead the Matsumoto and Lynch-Stieglitz (1999) linear approximation of Kim and O'Neil's (1997) equation is applied on their $\delta^{18}O_w$ and temperature data, values around 3.4‰, which is very close to ours, are obtained. Application of McCorkle et al. (1990) and their quadratic approximation of O'Neil et al. (1969) results in 4.4‰. After taxon-specific correction of −0.55‰ for *Cibicides* spp. in AW and of −0.35‰ for *C. wuellerstorfi* in uABW and ABW, as suggested, $\delta^{18}O_{Cib}$ mean values plausibly reflect the density stratification of the principle water masses in the Nordic Sea and Arctic Ocean. However, a separation just by means of $\delta^{18}O_{Cib}$ values between uABW and overlying AW on the one hand, and uABW and underlying ABW on the other hand, is statistically not always significant (Fig. 5; Tables 1, 2).

It is worth mentioning that the *C. lobatulus* stable isotopic composition is widely used and regarded as reliable substitute for *C. wuellerstorfi*, which is generally absent from shallow shelf and upper slope environments (Mackensen et al., 1985, 1995; Holbourn et al., 2013). Although deviations of *C. lobatulus* $\delta^{13}C$ from the $\delta^{13}C_{DIC}$ of dissolved inorganic carbon are reported (Mackensen et al., 2000), not much is known about its behavior with regard to its oxygen isotopic fractionation. This may be due to the fact that many geochemical and paleoceanographic investigations do not differentiate between close relatives *C. lobatulus* and *Cibicidoides* spp. (cf., Schweizer et al., 2009, 2012). However, since there is clear evidence in our data set for a taxon-specific $^{18}O$ fractionation during calcification, we suggest differentiation between the shallow water species *C. lobatulus* (Walker and Jacob) and *C. refulgens* (de Montfort) on the one hand and the deep-water *C. wuellerstorfi* (Schwager) on the other.

## *4.2. AW and uABW on upper slopes*

### *4.2.1. Water $\delta^{18}O_{equ}$*

Even though we just roughly differentiated by oceanic region and water masses (Fig. 2), mean $\delta^{18}O_{equ}$ of AW on Nordic Sea slopes being 0.20‰ lower than on Barents, Kara and Laptev Sea slopes, and again 0.08‰ lower than on eastern Siberian slopes of the Amerasian basins, reasonably well reflects the cooling of AW on its way from the Nordic Sea into the Amerasian Arctic (Table 1, Fig. 3). Since the $\delta^{18}O_{equ}$ shift of 0.28‰ corresponds to a cooling of roughly 1 °C, which is about one third of what temperature measurements tell us, the trend of increasing $\delta^{18}O_{equ}$ due to decreasing temperatures may be counteracted by increasing amounts of fresh water from river runoff. This, however, is not indicated by accordingly decreasing $\delta^{18}O_w$ values from the upper slopes and shelf breaks (Table 1).

In contrast to AW, uABW $\delta^{18}O_{equ}$ values do not change from the Nordic Sea into the Eurasian basins but increase by 0.06‰ across the Lomonossov Ridge into the Amerasian basins (Table 1, Fig. 3). This is in agreement with a diminishing warming of uABW by AW (Jones et al., 1995; Jones, 2001).

### *4.2.2. Benthic $\delta^{18}O_{Cib}$*

Low mean benthic $\delta^{18}O_{Cib}$ on Greenland and Svalbard continental slopes of 3.25‰ between 150 and 750 m water depth, on the one hand reflect the warm but salty and relatively high-$\delta^{18}O_w$ of AW off the Barents Sea and Svalbard, and on the other hand, cold but less saline, relatively low-$\delta^{18}O_w$ off the Greenland continental ice cap and downstream of the Arctic outflow through Fram Strait (Table 2; Figs. 3–5).

In the eastern Eurasian basins (Table 2; Figs. 3, 5: EEU) at the Kara and Laptev Sea slopes mean $\delta^{18}O_{Cib}$ reached 3.57‰, and finally at the East Siberian Sea slopes (AME), 3.69‰. This increase in mean $\delta^{18}O_{Cib}$ of 0.12‰ would correspond to a decrease of water temperature of about 0.5 °C in AW at the upper slope, if a change in $\delta^{18}O_w$ is excluded. The calculated overall difference of 0.44‰ between AW $\delta^{18}O_{Cib}$ mean values at Nordic Sea upper slopes and $\delta^{18}O_{Cib}$ mean values at the East Siberian Sea slope corresponds to a temperature decrease by about 2 °C (Table 2, Figs. 3, 5). So we can summarize that the deduced cooling trend agrees with the decrease in temperature of Atlantic-derived water on its course from the entrance through Fram Strait into the Amerasian basins (Jones et al., 1995; Rudels et al., 2004).

However, a mean $\delta^{18}O_{Cib}$ of 2.79‰ on northern Greenland and Barents Sea continental slopes clearly is the lowest one calculated in this study, and more importantly it is lower than the mean of 3.25‰ calculated for the upper slopes of the Nordic Sea (Table 2, Figs. 3, 5). The $\delta^{18}O_{equ}$ of the Nordic Sea summer mean of our "AW" is by 0.2‰ lower than the Arctic Ocean summer mean of AW, and thus is well in line with a continuously cooling AW on its way north. Also, a mean $\delta^{18}O_w$ AW of 0.26‰ in the Nordic Sea versus 0.22‰ in the Eurasian basins calls for higher water temperatures in the Nordic Sea to compensate for in $\delta^{18}O_{equ}$ (Table 1, Fig. 5). So either an artifact during calculation of mean $\delta^{18}O_{Cib}$ or $\delta^{18}O_{equ}$, or an environmental influence not yet considered has to be taken into account.

Generally $\delta^{18}O_{equ}$ may vary independently from water temperature. Cold fresh water with low $\delta^{18}O_w$ will increase the $\delta^{18}O_{equ}$ by its low temperature but decrease $\delta^{18}O_{equ}$ due to its low $\delta^{18}O_w$. The $\delta^{18}O_{Cib}$ then just records the combined attenuated signal as is reflected in $\delta^{18}O_{equ}$ at the time of calcification. On the Eurasian shelves the mean $\delta^{18}O_w$ of river water varies between values of −16.3‰ (Ob River) and −19.4‰ (Lena River), whereas on the East Siberian shelf values range from −21.1‰ (Yana River) to −22.4‰ (Kolyma River) (Bauch et al., 2005). However, the runoff of east Siberian rivers into the East Siberian Sea and Amerasian basins is just about 10% of the Eurasian rivers runoff (Bauch et al., 2005). So in the end $\delta^{18}O_w$ of Eurasian shelf water may be lower than on the Amerasian shelves. If varying over seasonal or annual time scales, just small amounts of fresh water with −19 ± 3‰ may regionally modify an oceanic $\delta^{18}O_w$ of 0.25 ± 0.05‰ and thus the $\delta^{18}O_{equ}$ and $\delta^{18}O_{Cib}$. So our mean $\delta^{18}O_{equ}$ calculated from summer $\delta^{18}O_w$ and temperature measurements for about three decades do not necessarily reflect the interannual variations of river runoff as our $\delta^{18}O_{Cib}$ means may do. The latter values are less attenuated because of the lower number of sediment samples from less cruises collected with box or multiple corers compared to the large number of hydrographic data (Tables 1, 2).

Also, different sampling strategies for $\delta^{18}O_w$ and $\delta^{18}O_{Cib}$ on continental shelves with seasonal sea-ice formation and brine release may cause an artificial decoupling of $\delta^{18}O_{equ}$ and $\delta^{18}O_{Cib}$. So cold and dense brines formed on the shelves during winter by fresh and brackish surface water freezing and subsequent sea-ice formation with very low $\delta^{18}O_w$ but temperatures close to freezing would sink to the sea floor. Down there, later during spring, when food becomes available but brine still may be present in depressions and troughs on the shelf (cf. Bauch et al., 2012), epibenthic *Cibicides* spp. starts secreting its calcitic test, which then records low $\delta^{18}O_{Cib}$ due to the low $\delta^{18}O_w$ in spite of low bottom water temperatures. In contrast this low-$\delta^{18}O_w$ brine is not measured during regular oceanographic sampling of the water column, usually terminated at a water depth of about 5 to 10 m above the sea floor, and during the summer season.

A closer look at the individual 20 stations we arranged in the WEU box supposed to be bathed by AW reveals that only at six stations *Cibicides* spp. have low $\delta^{18}O$ values between 2.3 and 2.8‰, all of which were from very close to the glaciated NE Greenland coast or close to northern Svalbard's coast line. So local geography suggests that the $\delta^{18}O_{Cib}$ directly at the seafloor does not record $\delta^{18}O_w$ of brines from freezing of river runoff but rather of brines from freezing of seasonally and inter-annually varying glacier melt water. Holocene Greenland continental icecap $\delta^{18}O_w$ can be as low as -36‰; late Pleistocene values may reach -44‰ (Dansgaard et al., 1993).

In summary, (i) an artifact caused by incongruent sampling locations, different techniques, and different time periods, and (ii) glacier melt water as an environmental agent not explicitly considered in this study, are probably responsible for too low $\delta^{18}O$ values of *Cibicides* spp. in comparison to AW in the WEU compartment. This suggests that low $\delta^{18}O$ values of *C. lobatulus and C. refulgens* from shallow waters may be useful brine indicators.

In the intermediate water boxes, i.e. uABW, mean $\delta^{18}O_{Cib}$ does not change significantly although the overall trend as depicted in $\delta^{18}O_{equ}$ means is reflected (Tables 1, 2; Fig. 3). Note that we adjusted for the deepening influence of warm AW in the Amerasian compartment by deepening the division between uABW and ABW boxes.

### 4.3. ABW in deep basins

#### 4.3.1. Water $\delta^{18}O_{equ}$

Mean $\delta^{18}O_{equ}$ calculation of ABW reveal a range of 0.09‰ over all, 0.07 of which is due to an increase between the Arctic Ocean basins and the Nordic Sea (Table 1, Figs. 1, 3). Interestingly in the deep Nordic Sea mean summer $\delta^{18}O_w$ values of waters denser than $\sigma_{0.5}$ = 30.444 kg m$^{-3}$ vary around 0.30‰, i.e. 0.04‰ and just 0.01‰ higher than in the eastern Eurasian and Amerasian basins, respectively (Table 1). This would cause the calculated $\delta^{18}O_{equ}$ to rise by the same amount on the VPDB scale if temperatures are kept constant. Partly the higher $\delta^{18}O_{equ}$ in the deep Nordic Sea may also reflect slightly lower temperatures than in the Arctic Ocean (cf. Jeansson et al., 2008).

No significant change in mean $\delta^{18}O_{equ}$ from Eurasian to Amerasian basins is observed. This may either imply no change in $\delta^{18}O_w$ of ABW, or more likely that higher $\delta^{18}O_w$ is masked by the counteracting influence of higher bottom water temperatures in the Amerasian basins (cf. Bauch et al., 1995). Indeed, averaged summer $\delta^{18}O_w$ values of ABW, as defined in this study by the $\sigma_{0.5}$ = 30.444 kg m$^{-3}$ isopycnic line, in Amerasian basins are by 0.03‰ higher than in Eurasian basins (Table 1). Similarly, higher temperature and salinity have been reported in the Canadian basin than in the Eurasian basins (e.g., Jones et al., 1995).

#### 4.3.2. Benthic $\delta^{18}O_{Cib}$

Generally, mean $\delta^{18}O_{Cib}$ mirrors the same general pattern as $\delta^{18}O_{equ}$, with high values in the Nordic Sea and lower values in the Arctic basins (Table 2, Fig. 3). Unfortunately, due to low numbers of stations versus high variability of $\delta^{18}O_{Cib}$, the calculated standard error of means, indicates that these subtle differences between basins are statistically not significant (Table 2, Fig. 5). This, however, is in support of the uniform mean $\delta^{18}O_{equ}$ we calculated, due to the counteracting effects of increasing $\delta^{18}O_w$ and increasing temperatures, across Lomonossow Ridge between Eurasian and Amerasian basins (Table 1, Figs. 1, 3).

Analyses of Arctic Ocean halocline $\delta^{18}O_w$ and salinity suggest that sea-ice formation in the open ocean or in polynyas at the continental slope results in salinities of 32–34 and relatively high $\delta^{18}O_w$, whereas sea-ice formation in coastal polynyas on shallow shelves in close proximity to freshwater supply from Siberian river mouths produces salinities of 30–32 and lower $\delta^{18}O_w$ (Bauch et al., 2011b). Based on stable isotopes of benthic foraminifera from a brine-influenced shelf environment off Svalbard, and on $\delta^{18}O_w$ profiles in the Arctic Ocean, it was concluded that only brines with relatively high $\delta^{18}O_w$ might be salty and thus dense enough to significantly contribute to intermediate and deep waters (Bauch and Bauch, 2001; Rasmussen and Thomsen, 2009). Accordingly, sea-ice formation in river influenced coastal polynyas may not result in dense enough brines to ventilate deep basins, whereas comparatively high-$\delta^{18}O_w$ brine with only minor freshwater input may penetrate into the AW and uABW down to below 2000 m water depth. If so, it is not surprising that this is not reflected in our $\delta^{18}O_{Cib}$ data set from the Arctic basins seafloor bathed by ABW, because either high-$\delta^{18}O_w$ brine was similar to ABW $\delta^{18}O_w$ or low-$\delta^{18}O_w$ brine did not sink deep enough into the water column to significantly modify ABW $\delta^{18}O_w$. This hypothesis is further corroborated by data suggesting that in the Makarov Basin just about 0.2% of bottom water is derived from $\delta^{18}O_w$ river water (Bauch and Bauch, 2001). However, this seemed not to deplete the $^{18}O$ content of ABW in a measurable way (Bauch et al., 1995).

## 5. Conclusions

Since in the literature various corrections for taxon-specific "vital effects" are applied due to the use of different equilibrium calculations, limited calibration data, and lumping of *Cibicides* spp. and *Cibicidoides* spp., for the reconstruction of northern and high Arctic paleoenvironments, we suggest using the new taxon-specific corrections presented here. In the Arctic Ocean, water mass dependent differences in carbonate ion concentration may enhance a taxon-specific offset from equilibrium.

Low benthic $\delta^{18}O_{Cib}$ on the shelves off NE Greenland and north of Svalbard bathed by AW, may reflect admixture of low-$\delta^{18}O_w$ brine rejected during freezing of glacier melt water. In contrast, in the deep basins and on the lower continental slopes no influence of low-$\delta^{18}O_w$ brine is recorded in Recent deep-water $\delta^{18}O_{Cib}$. Possibly, low-$\delta^{18}O_w$ brine is strongly diluted in the deep sea and/or comparatively high-$\delta^{18}O_w$ brine from high-salinity shelves is preferentially incorporated and ventilates the deep Arctic Ocean.

The Recent $\delta^{18}O_{Cib}$ values from the Arctic Ocean suggest that if low-$\delta^{18}O_w$ brine formation during stadials in the Nordic Sea and the North Atlantic has taken place and may have been enhanced in the Arctic Ocean (Vidal et al., 1998; Dokken and Jansen, 1999; Meland et al., 2008; Thornalley et al., 2010; Dokken et al., 2013), then troughs and deeps on shallow boreal continental shelves should be the prime archives to look for *Cibicides* spp. $\delta^{18}O$ that may be significantly depleted. If high-$\delta^{18}O_w$ brines from high-salinity shelves should have ventilated the deep Nordic Sea and North Atlantic during stadials, no signal would be recorded in epibenthic $\delta^{18}O_{Cib}$. So, if negative $\delta^{18}O_{Cib}$ spikes in late Pleistocene deep-sea sediments were caused by low-$\delta^{18}O_w$ brines, they should have been denser than those from Recent Arctic Ocean low-salinity shelves.

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