Irina Polovodova Asteman, Helena L. Filipsson and Kjell Nordberg: Tracing winter temperatures over the last two millennia using a NE Atlantic coastal record. Climate of the Past Discussions.

The manuscript presents a record of environmental and climate change from the Eastern North Atlantic region. The data are of good quality and although some local conditions may be expected, the authors argue well that the record does indeed relate to more regional conditions. One of the most interesting aspects is that the record may indeed represent winter conditions. The subject fits well to Climate of the Past and overall the manuscript is well written, although the general significance of the data could be made clearer and some additional explanations are in order. I thus recommend publication following moderate revisions.

1.) Relevance: The authors need to explain better why the study is important. It is mentioned in the introduction that only few high-resolution records of late Holocene conditions exist from the eastern North Atlantic region. But records also exist from other regions, both Iceland and the western North Atlantic and the Labrador Sea region. Why is the Eastern North Atlantic region important? Please add a short explanation, what is special/different about this region compared to other areas. How can this study improve our general understanding of the late Holocene climate of the North Atlantic and which mechanisms control climate and ocean variability?

2) Bottom Water Tempertures: Page 3, line 28-31. It is stated that the water exchange only occurs during winter. Does any change in salinity or temperature conditions of the bottom waters occur during spring/summer?

Explain more clearly whether the Bottom Water Temperatures actually represent winter conditions (mention this also in the abstract). As this is a central part of the work, it needs to be explained very clearly.

Seasons used in the bottom-water temperature reconstruction (p9): Traditionally the winter season is described through the months DJF, but here the period JFM is used. Why? Is there a local environmental reason for this, purely due to available data, or …? Similarly an explanation should be given for the use of May-August as the summer period, but this is normally JJA. It is not directly stated in paragraph 4.3 that these periods correspond to “winter” (JFM) and “summer” (MJJA), but in the following discussion (paragraph 5) winter temperatures are mentioned, so I assume that this is the case? However, it needs to be stated clearly and explained properly

3) General interpretation and potential link to the North Atlantic Oscillation: 3a) The section on the influence of the North Atlantic Oscillation (NAO) on the Gullmar Fjord (p 11, lines 1.6) should be moved to the introduction, with reference also to modern data from NE Europe/NE Atlantic. No reference to NAO during past climate periods should be mentioned as fact before this is discussed in the following paragraphs.

3b) The potential role of the NAO is discussed for the MCA and LIA. But what about the RWP and the DACP? Several studies have indicated that climate during these periods may also be linked to the NAO, and the manuscript would benefit form a more in-depth discussion – and reference to a wider range of previously published studies. It is also noteworthy that the authors only refer to work that shows comparable conditions as seen in Gullmar Fjorden, omitting any other studies. The authors should also look towards studies on the Late Holocene from further afield, e.g. Portugal, East Greenland, the Labrador Sea.

4) Hypoxia: On P. 7, line 5 and again Page 10, line 17-18 it is mentioned that *C. laevigata* has become a rare species in the Gullmar Fjord since 1990. One page 7 no explanation is given, page 10 the phenomenon is explained through hypoxia. However on page 15 a discussion is raised, whether it is due to hyposix and if yes, why. The discussion is certainly relevant but the fact that first a statement is made and later a discussion is raised, makes it confusing and somewhat messy.

I would suggest just to refer to "see discussion" instead of jumping the gun on p.7 and 10. Also, the discussion on p 15 does not really fit well to the remaining text, and a solution may be to move this hypoxia discussion to its own, separate paragraph.

With respect to this discussion, the authors basically explain the hypoxia as due to climate change. However, what about the increased nutrient supply seen due to more intensified farming seen in the general region, may this also play a role? Please discuss.

5) Conclusion paragraph: The paragraph should be expanded with a synopsis on the discussion on the processes driving the climate change.

Minor comments:

Foraminiferal species: add author name to the species name the first time a species is mentioned: i.e., *Cassidulina laevigata* d'Orbigny, 1826; *Adercotryma glomerata* (Brady, 1878); *Hyalinea balthica* (Schröter in Gmelin, 1791).

P5, line 14; reservoir correction: How many bivalve shells and from many sites in the Gullmar Fjord is is reservoir correction based on?

P.9, line 24: add reference for timing of the foraminiferal growth season.

P10, line 22-25: add references for the mentioned climatic intervals.

Page 15: Could the stronger recent warming of the Marlangen Fjord region be due to a more direct link to the northward flow of Atlantic water compared to Gullmar Fjord, which is not in direct contact with the core of the Atlantic water?

Figures and figure captions:

All terms and abbreviations should be explained.

Fig 1: explain abreviations for current names. Fig 1a: land masses are shown in a very pale gray – it would be easier to see, if landmasses were shown in a slightly darker colour.

Fig. 2: BWT needs to be explained wither in the figures or the figure captions, as it should be possible to understand the figures without reading the main text.

Fig 3A: I cannot distinguish between the upper and lower symbol; please make them more different.

Fig. 5: explain BWT, RWP, DA, LIA etc in the figure caption. Mark the present BWT range on the figure.

Fig. 6: explain the pink and blue inervals.

Fig. 7: Here "bottom water temperature" is written in full (not giving the abbreviation) – be consistent.

Some additional comments are provided as comments in pdf file of the manuscript (only relevant pages).

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Tracing winter temperatures over the last two millennia using a NE Atlantic coastal record

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Abstract. We present 2500 years of reconstructed winter temperatures by using a fjord sediment archive from the NE Atlantic region. The study is based on a ca. 8-m long sediment core from Gullmar Fjord (Sweden), dated by ^{210}Pb and AMS ^{14}C and analysed for stable oxygen isotopes (δ^{18}O) measured on shallow infaunal benthic foraminiferal species *Cassidulina laevigata*. The bottom water temperatures (BWTs), calculated by using a palaeotemperature equation of McCorkle et al (1997), range between 2.7 - 7.8°C and are within the annual temperature variability, instrumentally recorded in the deep fjord basin since the 1890s. The record demonstrates a warming during the Roman Warm Period (~350 BCE – 450 CE), variable bottom water temperatures during the Dark Ages (~450 – 850 CE), positive bottom water temperature anomalies during the Viking Age/Medieval Climate Anomaly (~850 – 1350CE) and a long-term cooling with distinct multidecadal variability during the Little Ice Age (~1350 – 1850 CE). The fjord BWT record also picks up the contemporary warming of the 20th century, which does not stand out in the 2500-year perspective and is of the same magnitude as the Roman Warm Period and the Medieval Climate Anomaly.

1 Introduction

The climate variability over last two millennia has been widely recognized as crucial for the understanding of the present and future climate responses to anthropogenic forcing (e.g. Cunningham et al., 2013; Pages2k, 2013; McGregor et al., 2015; Abram et al., 2016). To evaluate how significant regional climate changes are or if observed temperature anomalies are unprecedented in view of long-term climate evolution, there is a need for long historical instrumental climate records. A major limiting factor for the reconstructions of past climate changes, both by using proxy data and paleoclimate modelling, is often a lack of such long instrumental records, which if available seldom reach beyond the 20th century. The North Atlantic region plays in this respect a paramount role for climate variability by modulating the European climate through e.g. changing intensity of the Atlantic Meridional Overturning Circulation (AMOC) (e.g. Eiríksson et al., 2006; Lund et al., 2006; Park and Latif, 2008; Trouet et al., 2009). Hence, high-resolution paleoceanographic records, which preferably overlap with instrumental observations and historical data, are needed from the North Atlantic. At the same time many of the marine records available from the region to date ~~naturally~~ tend to have low temporal resolution due to their location in the deep-sea or within the open shelf areas. At the same time, crucial knowledge has been gained from temperature proxy datasets available ~~to date~~ from the North Atlantic and northern hemisphere, which represent either composite records of different climate characteristics with various temporal resolution or are a combination of historical and proxy data; with generated data sets mostly reflecting summer conditions ~~in~~ at higher latitudes (e.g. Moberg et al. 2005; Gunnarsson et al., 2011; Butler et al., 2013; Cunningham et al., 2013; PAGES2K, 2013, 2017; Sicre et al., 2014; Linderholm et al., 2015). In contrast, based on instrumental records, increased winter temperatures have been suggested as an important driver of the most recent warming (Cage and Austin, 2010) and, hence, climate proxies incorporating winter signal are needed. Sediment archives of the fjord inlets located within the North Atlantic region offer the potential of high-resolution records of maritime climate, acting as sediment traps resolving climate variability at nearly annual resolution (Howe et al., 2010). Yet to date, there are relatively

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few such high-resolution paleoclimate records from the Eastern North Atlantic fjords spanning the late Holocene (among others Mikalsen et al., 2001; Klitgaard-Kristensen et al., 2004; Cage and Austin, 2010; Filipsson and Nordberg, 2010; Hald et al., 2011, Kjennbakken et al., 2011; Faust et al., 2016).

Herein, we present a high-resolution winter temperature proxy record from the Gullmar Fjord, on the west coast of Sweden, which illustrates the climate development in NW Europe over the last ~2500 years. The reconstructed temperatures are based on stable oxygen isotopes ($\delta^{18}O$) measured in shells (tests) of a shallow infaunal foraminifer *Cassidulina laevigata* and reflect the deep-water temperatures in the fjord basin. The fjord has a >100-yr long record of instrumental observations from the deepest basin, performed since 1869 (Fig. 2A-C); furthermore, >100-yr long time series of air temperature observations are also available for Stockholm, Sweden and central England. These instrumental observations of bottom water - and air temperatures are used to evaluate the accuracy of the reconstructed climate variability for the last century provided by the fjord sediment archive.

2 Study area

Gullmar Fjord is a Skagerrak fjord inlet, which is 28 km long and 1-2 km wide and oriented south-west to north-east (Fig. 1). The maximum basin depth is 118.6 m. The fjord is located at critical latitude picking up fluctuations between cold and temperate climates and has almost no tidal activity. The adjacent Skagerrak largely determines the local hydrography so that the deep (basin) water, which is typically exchanged in the fjord during the winter, originates from the North Sea surface water flowing into the Skagerrak with the present-day current circulation system (Svansson, 1975; Nordberg, 1991). The 42-m deep sill at the fjord entrance restricts the water exchange and results in water column stratified due to salinity differences (Fig.1C). At the surface (<1m) there is a thin layer of river water from the Örekilsälven (Fig.1), which does not significantly impact the fjord hydrography (SMHI, 1994; Arneborg, 2004). Below, at 1-15 m water depth, there is a brackish water mass (S=24-27), primarily derived from the brackish Baltic current flowing northward along the Swedish west coast. The brackish water mass has a residence time of 20-38 days in Gullmar Fjord (Ar,neborg et al., 2004). A more saline water mass (S=32-33) at ~15-50 m is derived from the Skagerrak and has mean residence time of 29-62 days (Arneborg et al., 2004). The last and deepest layer (>50 m), referred herein as deep water or basin water, is more stagnant, with little seasonal and inter-annual changes in salinity ranging between 34 and 35 and inter-annual temperature variability of 4 - 8°C (Fig. 2A, B). The deep water temperatures vary between the years depending on the temperature of the inflowing water mass but remain stable seasonally (Fig. 2D). The deep-water salinities seasonally do not vary much from the average value of 34.5 (Fig. 2B). The stratification of the water column is strengthened during the summer by the development of a strong thermocline. The deep-water exchange of the fjord basin water takes place once a year during winter, mostly between January and March. Due to a presence of a sill and the large basin volume, the temperature and salinity of the inflowing North Sea/Skagerrak water, is preserved in the basin until the next turnover, which often occurs in the following year (Arneborg et al., 2004). The benthic foraminifers reproduce and grow in the fjord during the following spring-summer-autumn (Gustafsson and Nordberg 2001),

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incorporating the winter-signal in their shells. This results in a stable oxygen isotope signal mainly reflecting winter temperatures of the North Sea surface water and the Skagerrak intermediate water.

After an extensive deep-water exchange event in the fjord the oxygen level starts to decline in June, and the lowest oxygen levels normally develop between November and January, usually dropping below 2 ml O_2 l^{-1} indicating hypoxic conditions, but anoxia has not been recorded (Fig. 2F). The first ever documented severe hypoxic event was noted in February 1890 by Pettersson and Ekman (1891). In the following, severe hypoxic events (≤1 ml O_2 l^{-1}) were measured in 1906, 1961/62, and 1973/74 (Fig. 2C) but due to the low observation frequency and duration of these events are not well documented. Since 1979, multiple episodes of more frequent severe hypoxia lasting for at least 3 months have been observed. These events occurred in 1979/80, 1983/84, 1987/88, 1988/89, 1990/91, 1994/95, 1996–1998, 2008, 2014/2015 (e.g. Filipsson and Nordberg 2004a; Polovodova Asteman and Nordberg 2013; SMHI SHARK-database, 2017).

The severe hypoxia makes the fjord basin hostile for large burrowing organisms but allows benthic meiofaunas to thrive. This lowers sediment bioturbation and results in well-preserved environmental sediment archive. The fjord basin has high sediment accumulation rates, which provide a high temporal resolution corresponding to 1-6 years per 1-cm thick sediment sample. Finally, the fjord sediment archive is characterized by the diverse and abundant foraminiferal faunas and dinoflagellate cysts, which have already provided some insights in climate evolution and associated environmental changes on the Swedish west coast during the last two millennia (Filipsson & Nordberg, 2004a; Harland et al., 2006; Nordberg et al. 2009; Filipsson and Nordberg 2010; Polovodova et al., 2011; Harland et al., 2013; Polovodova Asteman & Nordberg, 2013; Polovodova Asteman et al., 2013).

3 Material and Methods

This study is based on a composite record of two sediment cores: GA113-2Aa and 9004, which were both collected at 116 m water depth at the same site in the deepest Gullmar Fjord basin (58°17.570' N, 11°23.060' E) (Fig. 1), for which the long-term hydrographic observations are available (Fig. 2A-C). The core 9004 (731-cm long) was taken with a gravity corer (Ø=7.6 cm) ~~a~~ onboard ~~of~~ *R/V Svanic* in July 1990. The core GA113-2Aa (60-cm long) with an intact sediment-bottom water interface was recovered by using a Gemini corer (Ø=8 cm) in June 1999 from the *R/V Skagerak*. In the laboratory both cores were split in two halves and sectioned in 1-cm intervals. One half was used for bulk sediment geochemistry (TC, TN and C/N ratio), stable oxygen and carbon isotopes, dinoflagellate cysts- and benthic foraminiferal faunal analyses. Another half was stored as an archive at the Department of Geosciences, University of Gothenburg. The TC and stable carbon isotope data from both cores are published in Filipsson and Nordberg (2010), dinoflagellate cysts data are discussed in Harland et al. (2006, 2013), while C/N and foraminiferal assemblage data are presented in Filipsson & Nordberg (2004a), Polovodova et al. (2011) and Polovodova Asteman et al. (2013). We also present data from the gravity core G113-091, collected at the same location as GA113-2Aa & 9004 aboard of R/V Skagerak in September 2009, and used herein only (similar to our previous study) to create a composite age model for the cores GA113-2Aa and 9004 (Polovodova Asteman et al., 2013; see below).

In addition to the above-mentioned cores, we also use six surface samples (0-1cm) collected at five stations in the Skagerrak (OS4, OS6, OS14, 9202 and 9205) and one station in the Gullmar Fjord (G113-091a: the same location as for GA113-2Aa & 9004) in 1992-93 and 2009, respectively (Fig.1B, C; Table 1). All surface samples were stained by rose Bengal to distinguish individuals presumably living at the moment of sampling from the empty foraminiferal shells.

3.1 Sediment core dating and age model

The age model for the composite GA113-2Aa and 9004has been previously published in Filipsson and Nordberg (2010) with further revisions by Polovodova et al. (2011) and Polovodova Asteman et al. (2013). Eleven intact marine bivalve shells were recovered in life position from the core 9004 and were subject to the AMS ^{14}C analysis (Fig. 3A; Table 2). All ^{14}C dates were obtained through analysis at the Ångström Laboratory (Uppsala University, Sweden) and originally were calibrated using the marine calibration curve (Reimer et al., 2004; Bronk Ramsey, 2005). Ages were normalized to δ^{13}C of − 25‰ according to Stuiver and Polach (1977), and a correction corresponding to δ^{13}C= 0‰ (not measured) versus PDB has been applied. Herein we present ages recalibrated by using Calib Radiocarbon Calibration software v. 7.1 (Stuiver et al., 2017: http://calib.org/calib/), the most recent marine calibration curve (Reimer et al, 2013) and a reservoir age of 500 yr (ΔR=100±50), which has been obtained on pre-bomb marine bivalve shells from the Gullmar Fjord, provided by the Natural History Museums in Gothenburg and Stockholm (Nordberg and Posnert, unpubl. data). All ages are presented as median probability with 1-σ error margin (Table 2). Two dates at 98 cm and 313 cm showed minor age reversals and were omitted from the final age model (Table 2). The core GA113-2Aa was dated by using ^{210}Pb and a constant rate of supply (CRS) model (Appleby and Oldfield, 1978), which suggested that the core material was deposited between ca. 1915 and 1999 (Fig. 3A). For details regarding GA113-2Aa age model see Filipsson and Nordberg (2004a).

Together the cores GA113-2Aa & 9004 proved to be a continuous sediment record with no gap in between based on correlation of the stable carbon isotopes (δ^{13}C) and benthic foraminiferal species *C. laevigata, Adercotryma glomerata* and *Hyalinea balthica* with respective data from the core G113-091 (Fig. 3B herein; Polovodova Asteman et al., 2013; Polovodova Asteman and Nordberg. 2013). The composite record of GA113-2Aa & 9004 spans from approximately 350 BCE to 1999 CE (Table 2, Fig. 3A), and includes the late Holocene climate events such as the Roman Warm Period (RWP: ~350 BCE – 450 CE), the Dark Ages Cold Period (DA: ~450 – 850 CE), the Viking Age/Medieval Climate Anomaly (VA/MCA: ~850 – 1350 CE), the Little Ice Age (LIA: ~1350 – 1850 CE) and the contemporary warming from 1850 CE to present (Lamb, 1995; Filipsson and Nordberg, 2010; Harland et al., 2013; Polovodova Asteman et al., 2013; Helama et al., 2017). We add the Viking Age to the Medieval Climate Anomaly following the approach of Filipsson and Nordberg (2010), based on historical evidence that warming in Northern Europe began earlier than 1000 CE, which allowed Vikings to reach the NE coast of England and loot the monastery of Lindisfarne in 793 CE (Morris, 1985). For further details on chronology of the cores GA113-2Aa and 9004 see Filipsson and Nordberg (2004a), Polovodova et al., 2011; and Polovodova Asteman et al. (2013).

Combining the long gravity core with the 60 cm long Gemini core, which includes the sediment-bottom water interface and, hence, the intact core top, resulted in a high-resolution temporal record of almost 1-year cm^{-1} sample for the upper part of the record and <10 years cm^{-1} sample for the deepest part of the record. Calculations from the ^{210}Pb analyses and the AMS-^{14}C dates suggest sediment accumulation rates of ~9 mm year^{-1} in the most recent sediments and approximately ~2.8 mm year^{-1} in the compacted deepest part of the gravity core (Fig. 2). Hence, due to high accumulation rates the upper 60 cm of the record can be directly compared to instrumental hydrographic and meteorological data (Figs 6, 7).

3.2 Stable oxygen isotopes

We measured δ^{18}O on tests of shallow infaunal foraminifer *Cassidulina laevigata* from the core top samples and from the ca. 8-m long G113-2Aa - 9004 record (Fig.1B). Between 12 and 20 specimens of *Cassidulina laevigata* were picked from each sample for the analysis. In total 6 and 425 samples were analysed for stable oxygen isotopic composition for the surface sediments and composite G113-2Aa - 9004 record, respectively. All samples were measured at the Department of Geosciences, University of Bremen, Germany, using a Finnigan Mat 251 mass spectrometer equipped with an automatic carbonate preparation device. Isotope composition is given in the usual δ-notation and is calibrated to Vienna Pee Dee Belemnite (V-PDB) standard. The analytical standard deviation is <0.07‰ for δ^{18}O based on the long-term standard deviation of an internal standard (Solnhofen limestone).

The temperature was reconstructed using the salinity: $\delta^{18}O_w$ relationship established by Fröhlich et al. (1988) (eq. 1), which is representative for this region (Filipsson, unpubl. data). An average salinity value of 34.4 (range 33-35) was used in equation 1, based on instrumental measurements between 1896 and 1999 for the fjord deep-water (station Alsbäck). The salinity (*S*) was assumed to be constant over the investigated time period.

$$\delta^{18}O_w = 0.272 \times S - 8.91 \qquad (1)$$

To calculate temperatures the paleotemperature equation by McCorkle et al. (1997) was applied (eq. 2). This equation is more appropriate to the temperature range observed in temperate fjord basin than the more commonly used linear equation by Shackleton (1974), which produces unrealistically high temperatures in our study (see results section). The bottom water temperature in degrees Kelvin (T °K) was calculated as follows:

$$T^oK = \sqrt{\frac{2.78*10^3}{\ln(\frac{\delta^{18}O_c+1000}{0.97006*\delta^{18}O_w-29.94}+1000)} + \frac{2.89}{10^3}} \qquad (2)$$

Here, $\delta^{18}O_c$ stands for stable oxygen isotopic ratio ^{18}O/^{16}O measured in calcite tests of *C. laevigata*, while $\delta^{18}O_w$ is the isotopic composition of water calculated from equation 1 and converted from SMOW to V-PDB by subtracting 0.27‰ (Bemis et al., 1998)

Finally, to convert reconstructed temperatures to degrees Celsius, equation 3 was used:

$$T°C=T°K-273.15 \quad (3)$$

Since 1990 *C. laevigata* has become a rare species in the Gullmar Fjord deep basin (Fig. 6), which resulted in short gap in the most recent part of the record. Similar gaps in $\delta^{18}O$ and, hence, bottom water temperature data are also seen for the earlier part of the record and are due to absence or very low abundances of *C. laevigata* (Fig. 6).

3.3 Hydrographical and meteorological instrumental data

Long-term hydrographical instrumental data for temperature, salinity and dissolved oxygen concentration [O_2] for the fjord basin (average for 110-118 m w.d.) were extracted from the Swedish Meteorological and Hydrological Institute (SMHI) SHARK database (https://www.smhi.se/klimatdata/oceanografi/havsmiljodata/marina-miljoovervakningsdata). Some of the Gullmar instrumental data is also available from the Water Quality Association of the Bohus Coast (BVVF) (http://www.bvvf.se/), while the data prior to 1958 come from Engström (1970). The Skagerrak hydrography data for the stations adjacent to OS4-6, 9202, 9205 and OS14 were obtained from the International Council for the Exploration of the Seas (ICES: http://www.ices.dk/marine-data/).

Meteorological observations of air temperature were also obtained for Stockholm (https://www.smhi.se/klimatdata) and the Central England (http://www.metoffice.gov.uk/), which both have the longest historical meteorological records going as far back as the 18th century.

4 Results

4.1 Core tops

To obtain an error estimate and to facilitate the choice of the paleotemperature equation we used stained (living) tests of *Cassidulina laevigata* from the core top samples collected in the Gullmar Fjord and the adjacent Skagerrak. Calculated bottom water temperatures based on the $\delta^{18}O_c$ values from the living (stained) *C. laevigata* were compared to ICES and SMHI hydrography data from the adjacent stations (Fig. 4A). Also the $\delta^{18}O_c$ values predicted from the chosen equation (see below)

were used to estimate the reliability of our temperature reconstruction (Fig. 4B). *Cassidulina laevigata* has been previously suggested to calcify 0.19‰ lower than equilibrium (Poole et al., 1994). Our $\delta^{18}O_c$ data from the core tops demonstrate an offset, ranging between 0.01‰ and 0.27‰ (mean 0.15‰), compared with $\delta^{18}O_c$ predicted using the palaeotemperature equation from McCorkle et al (1997) (Fig. 4B). Applying the mean correction of +0.15‰ to the Gullmar $\delta^{18}O_c$ record results in bottom water temperatures ~0.5-1°C higher than those recorded by instrumental observations in the fjord (Fig. 2A), while uncorrected $\delta^{18}O_c$ values produce temperatures close to observations. Taking the latter into the account and because based on available data it is difficult to estimate how large the correction should be, we further report the uncorrected $\delta^{18}O_c$ values both for the core tops and for the sediment cores. Instead, we use a median value (0.7°C) of the range in produced temperature offset (Fig. 4A) as an error margin for our paleotemperature reconstructions (Figs 5-6).

Instrumental temperature data from ICES and SMHI were used to calculate $\delta^{18}O_c - \delta^{18}O_w$ for the core top samples to facilitate the choice of a paleotemperature equation. Plotting $\delta^{18}O_c$-$\delta^{18}O_w$ versus observed temperature data for different paleotemperature equations (Fig. 4C) allows estimating which of the equations gives the best possible agreement with the core top data and, hence, is the most appropriate for temperature reconstructions. Figure 4C shows that $\delta^{18}O$ values from the NW Skagerrak (OS4 and OS6) are clearly in better agreement with equations by Hays and Grossman (1991) and McCorkle et al. (1997), while the central Skagerrak samples (9202 and 9205) plot close to the linear equation by Shackleton (1974). The Gullmar Fjord (G113-091) and the OS14 station, taken just outside the fjord, occupy a space in between the Shackleton equation and those by Hays and Grossman (1991) and McCorkle et al. (1997). This suggests that applying the Shackleton equation for Gullmar Fjord and Skagerrak will result in temperatures higher than observations, which has been also observed for *Cibicidoides* and *Planulina* from Florida Straits (Marchitto et al., 2014). Indeed, when testing the Shackleton equation on our dataset, the temperatures are warmer than the ICES hydrographic observation data by 1.5-2°C. In contrast, the equation by Bemis et al. (1998) applied to the core top $\delta^{18}O_c$ data produces the coldest temperatures, which are 0.9-1.9°C colder than observations. In turn, it appears that by using Hays and Grossman (1991) or McCorkle et al. (1997) equations, the corresponding calculated temperatures come closer to observations. Both equations are nearly identical for the temperature range 5-8°C (Fig. 4C) observed between 1890 and 2001 (Fig. 2) and by exercising both equations on Gullmar Fjord $\delta^{18}O_c$ record the almost identical paleotemperature curves are produced. This is rather curious because the equation of Hays and Grossman (1991) is based on meteoric calcite of non-biogenic origin. For this reason, in the current paper we apply the McCorkle et al (1997) equation for the paleotemperature reconstructions.

4.2 Composite record of G113-2Aa and 9004 sediment cores

The $\delta^{18}O$ record from the Gullmar Fjord shows both decadal and centennial variability for the last 2500 yr (Fig. 5) and can be divided into five major isotopic intervals: 1) For the lower part of the record at 802–592 cm, corresponding to ~350 BCE – 450 CE, the $\delta^{18}O_c$ values are generally lower (~2.4‰) than the long-term average of 2.7‰. 2) Between 598 and 475 cm (~ 425 – 900 CE) the $\delta^{18}O$ record demonstrates a considerable variability (Fig. 5), starting with higher $\delta^{18}O_c$ (2.8-3‰) at 598–574 cm (~ 425 – 525 CE), which then become lower (~2.4‰) at 574–529 cm (~ 525 – 700 CE) and increase again (~3.0‰) between 529 and 497 cm (~ 700 – 825 CE). 3) The 475–302 cm interval (~ 900 – 1350 CE) displays again lower $\delta^{18}O_c$ (~2.4-2.5‰), which are below the long-term average. 4) From 302 to 53.5 cm (~ 1350 – 1900 CE) the stable oxygen isotope record increases again with the majority of the $\delta^{18}O_c$ values being ~3.1–3.2‰ and exceeding the long-term average. Within this interval the highest $\delta^{18}O$ values of >3.2‰ are found between 300 and 170 cm (~1350CE – 1580 CE). 5) Finally, the $\delta^{18}O$

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record becomes lower again (~2.4‰) between 53.5 and 5 cm (~ 1900 and 1996 CE). We did not find enough specimens of *Cassidulina laevigata* to perform isotopic analyses for samples between 5 and 0 cm (1996-1999).

Shifts of ~0.25‰ in $\delta^{18}O_c$ occur throughout the Gullmar Fjord $\delta^{18}O$ record, which according to the equation of McCorkle et al. (1997) used herein, may potentially indicate a temperature variability of ~1°C. A corresponding salinity change is rather small (0.02), calculated using the mixing line by Fröhlich et al. (1988) and by applying the $\delta^{18}O_c$ range of 2.6-2.85 and a corresponding temperature range of 4.9-5.9°C. Such salinity changes are well within the amplitude of inter-annual variability (1-1.5), recorded by instrumental salinity observations since the 1890 (Fig. 2). Foraminifera precipitate their tests during several months (e.g. Filipsson et al., 2004) and thus integrate the inter-monthly salinity signal, which together with annual variability is minimal according to the instrumental data. For the upper part of the record 1-cm sediment slice integrates one or possibly two growing seasons of *C. laevigata* and, hence, records a potentially higher variability of both salinity and temperature. In the deepest part of the record, however, a single 1-cm sample may correspond to ~7-10 years and, thus, more likely averages inter-annual salinity and temperature variability providing "a more smoothed" signal.

Stable carbon isotopes ($\delta^{13}C$) data from the composite G113-2Aa – 9004 record (Filipsson and Nordberg, 2010) were plotted against the oxygen isotope data presented herein, to investigate the potential relationship between the two e.g. due to different water masses (Suppl. Fig. 1). No such relationship was found (Suppl. Fig. 1), which indicates that our $\delta^{18}O$ record mainly reflects fjord deep-water temperatures.

4.3 Reconstructed bottom water temperatures (BWTs)

The resulting calculated bottom water temperature record is plotted as both as absolute temperature values (Fig. 5) and as anomaly from the mean value (5.4°C), based on the instrumental temperatures observed between 1961 and 1990 (Fig. 6). With very few outliers, the reconstructed temperature range (2.7 - 7.8°C) is within the present-day annual variability, documented from instrumental temperature measurements in the fjord deepest basin since 1890 (Fig. 2A-C). Observed annual temperatures registered between 1890 and 1996 (which corresponds to the uppermost part of the composite G113-2Aa – 9004 record) vary between 3.0 and 8.3°C, which gives an amplitude of 5.3°C. Corresponding instrumental 1890-1996 temperatures for foraminiferal growth season (April-August) show a 3.0 - 7.2°C range with an amplitude of 4.2°C. When studying the reconstructed temperatures over the last 2500 years the corresponding amplitude, i.e. the difference between the maximum (7.8°C) and the minimum (2.7°C) temperatures is 5.1°C (Fig. 5). Also when plotting the reconstructed bottom water temperatures for the period 1890 –1999 versus corresponding instrumental bottom water temperatures as annual average and means for May-August (Fig. 7B) and January-March (Fig. 7C), the calculated bottom water temperatures and hydrographic data agree with each other rather well in terms of amplitude. An increased agreement, however, is reached when comparing the reconstructed data to the hydrographic winter (Jan-March) temperature (Fig. 7C), which is not surprising considering the fjord hydrography and a season when deep-water exchanges typically occur (see Study Area

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parameters.

Figure 3: Age model of the studied Ga113-2Aa & 9004 record (A) and ~~correlation~~ comparison of foraminiferal and isotopic data with core G113-091, taken at the same location in ~~2009,~~ to prove show the absence of a gap between GA113-2Aa and 9004 (B), according to Polovodova Asteman et al (2013).

Figure 4: Comparison of reconstructed temperatures and $\delta^{18}O$ values measured in stained *C. laevigata* from the core tops collected in Gullmar Fjord (G113-091) and the Skagerrak (OS4, OS6, OS14, 9202, 9205) to hydrographic temperature data (A) and to $\delta^{18}O$ predicted from palaeotemperature equation (B) by McCorkle et al (1997). C: Temperature vs. $\delta^{18}Oc$ – $\delta^{18}Ow$, together with the paleotemperature equations from Shackleton (1974), Hays and Grossman (1991), Kim and O'Neil (1997), McCorkle et al. (1997), and Bemis et al. (1998).

Figure 5: A 2500-year long $\delta^{18}O$ record and reconstructed winter bottom water temperatures from Gullmar Fjord. Thick lines show 3-point running mean for both curves, and dashed lines indicate 1) a long-term average of 2.4‰ for $\delta^{18}O$ record and 2) 5.4°C - a mean for instrumental bottom water temperatures registered between 1961 and 1990. Grey shaded areas in BWTs indicate a median offset (0.7°C) in instrumental versus reconstructed temperatures obtained for rose Bengal stained *C. laevigata* from the core tops (see Fig. 4A), used herein as an error margin.

Figure 6: Reconstructed bottom water temperatures (as anomaly against the 1961-1990 instrumental mean of 5.4°C) from Gullmar Fjord compared against other temperature proxy records: annual northern hemisphere temperatures (Moberg et al., 2005), bottom water temperatures from Malangen Fjord in NW Norway (Hald et al., 2011) and Loch Sunart in Scotland (Cage and Austin, 2010), spring sea surface temperatures from Chesapeake Bay, E North Atlantic Ocean (Cronin et al., 2003), annual temperatures reconstructed for continental Europe (Pages2K, 2013) and the reconstructed NAO record from Trondheim Fjord, W Norway (Faust et al., 2016). Also are shown relative abundances of foraminifer *Cassidulina laevigata* in the fjord with abundance minima and respective gaps in temperature reconstruction linked to the positive NAO index (arrows). For location of these proxy records see Fig. 1A. Grey shaded areas in Gullmar Fjord BWT anomalies indicate a median offset (0.7°C) in instrumental versus reconstructed temperatures (see Fig. 4A) obtained for rose Bengal stained *C. laevigata* from the core tops, used herein as an error margin.

Figure 7: Comparison of the winter bottom water temperatures reconstructed from Gullmar Fjord record to instrumental basin water temperatures measured in the deepest fjord basin: the annual mean (a), mean for May-August (b) and mean for January-March (c).