Holocene sea subsurface and surface water masses in the Fram Strait – Comparisons of temperature and sea-ice reconstructions

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ABSTRACT

Two high-resolution sediment cores from eastern Fram Strait have been investigated for sea subsurface and surface temperature variability during the Holocene (the past ca 12,000 years). The transfer function developed by Husum and Hald (2012) has been applied to sediment cores in order to reconstruct fluctuations of sea subsurface temperatures throughout the period. Additional biomarker and foraminiferal proxy data are used to elucidate variability between surface and subsurface water mass conditions, and to conclude on the Holocene climate and oceanographic variability on the West Spitsbergen continental margin. Results consistently reveal warm sea surface to subsurface temperatures of up to 6 °C until ca 5 cal ka BP, with maximum seawater temperatures around 10 cal ka BP, likely related to maximum July insolation occurring at that time. Maximum Atlantic Water (AW) advection occurred at surface and subsurface between 10.6 and 8.5 cal ka BP based on both foraminiferal and dinocyst temperature reconstructions. Probably, a less-stratiﬁed, ice-free, nutrient-rich surface ocean with strong AW advection prevailed in the eastern Fram Strait between 10 and 9 cal ka BP. Weakened AW contribution is found after ca 5 cal ka BP when subsurface temperatures strongly decrease with minimum values between ca 4 and 3 cal ka BP. Cold late Holocene conditions are furthermore supported by high planktic foraminifer shell fragmentation and high δ18O values of the subpolar planktic foraminifer species Turborotalita quinqueloba. While IP25-associated indices as well as dinocyst data suggest a sustained cooling due to a decrease in early summer insolation and consequently sea-ice increase since about 7 cal ka BP in surface waters, planktic foraminiferal data including stable isotopes indicate a slight return of stronger subsurface AW inﬂux since ca 3 cal ka BP. The observed decoupling of surface and subsurface waters during the later Holocene is most likely attributed to a strong pycnocline layer separating cold sea-ice fed surface waters from enhanced subsurface AW advection. This may be related to changes in North Atlantic subpolar versus subtropical gyre activity.

1. Introduction

In the Arctic, effects of global climate change occur more rapidly and severe than in other regions on Earth due to the polar amplification (Manabe and Stouffer, 1980). Over the past few decades, the Arctic sea-ice cover has been shrinking continuously while more heat is delivered to the high north through different atmospheric and oceanic mechanisms. Climate model projections not only predict a change from perennial to more seasonal sea-ice cover in the Arctic but also globally rising sea level with significant feedbacks to global climate (e.g., Bengtsson et al., 2006).

The Fram Strait between Greenland and Svalbard is the only deepwater connection where warm and cold surface to deep water masses exchange between the Arctic and the world’s oceans. Northward flowing warm and saline Atlantic Water (AW) via the eastern Fram Strait strongly contributes to the Arctic Ocean’s heat budget.

Reconstructions of past climate and oceanographic conditions...
are essential for understanding and modelling of the current and future climate. Extending the record of ocean temperatures beyond the era of instrumental measurements facilitates improved knowledge about the long-term mechanisms of heat advection into the Arctic Ocean and water mass stratification. This also includes their forcing factors such as insolation, sea-ice extent, ocean current strength, and sea-level changes. During recent decades, the eastern Fram Strait as the major gateway between the northern North Atlantic and the Arctic Ocean has been studied extensively to better understand Holocene environmental changes (Hald et al., 2004; 2007; Slubowska-Woldengen et al., 2005; Rasmussen et al., 2007; 2013; Slubowska-Woldengen et al., 2008; Müller et al., 2009; 2011; 2012; Werner et al., 2011; 2013; 2014; Aagaard-Sørensen et al., 2014a,b) However, complex water mass interactions complicate straightforward reconstructions of past water mass temperatures in the area. In this paper, Holocene planktic foraminifera from the eastern Fram Strait have been used to reconstruct past sea surface water temperatures using transfer functions (Husum and Hald, 2012). Modern planktic foraminifer assemblages are dominated by two major species Neogloboquadrina pachyderma and Turborotalita quinqueloba. N. pachyderma is associated with cold polar waters (e.g., Bé and Tolderlund, 1971; Volkman, 2000) while T. quinqueloba is commonly linked to temperature and saline Atlantic Water advection in the Fram Strait area (Vollmann, 2000), in addition to the Arctic Front where Atlantic and Arctic water masses encounter (Johannessen et al., 1994; Matthiessen et al., 2001). T. quinqueloba is a symbiont-bearing foraminifer that is therefore bound to the uppermost photic zone due to required light conditions for photosynthetic activity (Bé, 1977). Different from studies in the central Bothnian Sea where both species record temperatures around the same depth but during different periods of the spring/summer season (Jonkers et al., 2010), recent studies of living planktic foraminifera from the Fram Strait show no clear differences in seasonal flux patterns between N. pachyderma and T. quinqueloba. Depending on water mass and sea-ice conditions, both species inhabit rather similar summer depth habitats within the uppermost 200 m of the water column (Manno and Pavlov, 2014; Pados and Spielhagen, 2014). Previous studies from the Fram Strait and western Svalbard have shown heat advection to the Arctic Ocean was enhanced during the Early Holocene, driven by maximum summer insolation and wind force and/or thermohaline circulation (e.g., Köc et al., 1993; Hald et al., 2007; Rasmussen et al., 2007; 2013; Slubowska-Woldengen et al., 2008), Most studies indicate a cooling trend after ca 8 cal ka BP (Hald et al., 2004; Slubowska-Woldengen et al., 2007; Müller et al., 2012). A significantly warmer Mid-Holocene and probably increased heat flux to the Arctic Ocean was found by some studies in the Barents Sea/Svalbard area with ocean temperatures likely higher than for the remainder of the Holocene (e.g., Sarthein et al., 2003; Hald et al., 2007; Rasmussen et al., 2007). Consistent with a decreasing summer insolation during the late Holocene, compared to the preceding early and mid-Holocene intervals, reconstructed seawater temperatures were lower; and more stable conditions with extended sea-ice conditions prevailed in the area (Sarthein et al., 2003; Müller et al., 2012; Werner et al., 2013). However, many of the reconstructions using planktic foraminifera and transfer functions were based on transfer functions with only few polar and subpolar modern analogues (e.g., Kucera et al., 2005; Husum and Hald, 2012). Two high-resolution sediment cores comprising the last ca 12,000 years have been studied for planktic foraminiferal content in order to reconstruct past subsurface temperatures at 100 m water depth (SST100; cf. Husum and Hald, 2012). Proxy data such as the ice-rafted debris and biomarkers complement SST100 reconstructions for a comprehensive reconstruction of Holocene variability in this crucial area with regards to the changing modern Arctic system. The spring sea ice proxy IP25 (Belt et al., 2007; Brown et al., 2014), the ratio of IP25 to its structurally related C25 highly branched isoprenoid (C25-HBI)diene (DIP25 index; Fahl and Stein, 2012; Cabedo-Sanz et al., 2013), and biomarkers indicative of phytoplankton productivity allow estimates of sea-ice and primary-productivity changes associated with temperature variations at the core site. Previously, it has been assumed that the DIP25 index could refer to relative sea surface temperature (at ca 10 m water depth; SST10) changes (Xiao et al., 2013; Müller and Stein, 2014). Here, for the first time, DIP25 values are directly compared to reconstructed ocean temperature trends, which may give insight into the suitability of the DIP25 index as a potential SST10 proxy.

2. Hydrography

Northward flowing warm and saline Atlantic Water enters the Arctic Ocean as the West Spitsbergen Current (WSC) via the eastern Fram Strait (Fig. 1). The WSC delivers warm and saline (T up to 6 °C, S > 35, Figs. 1 and 2) AW at surface to subsurface water depths to the Arctic Ocean where its upper part becomes transformed into a less saline surface layer by ice melt and mixing with low-saline waters of mainly riverine origin.

The strength of northward flowing AW is affected by the variability of both the cyclonic subpolar gyre (SPG) and the anticyclonic subtropical gyre (STG) in the North Atlantic Ocean (Hätün et al., 2005). In particular, increasing salinities in the northeastern part of the North Atlantic have been found to correspond to decreasing circulation of the SPG (Häkkinen and Rhines, 2004) as well as to the shape of the gyre (Hätün et al., 2005).

In response to Arctic Ocean surface current patterns (Transpolar Drift and Beaufort gyre), sea ice as well as cold, fresh water masses exit along western Fram Strait as part of the East Greenland Current (EGC), which carries cold and fresh polar waters (T < 0 °C, S < 34.4; e.g., Rabe et al., 2009) towards the Greenland Sea. Accordingly, sea-ice extent in the Fram Strait is controlled by two opposite current systems keeping eastern Fram Strait at present ice-free year-round while western Fram Strait stays perennially sea-ice covered.

3. Material and methods

The 9.54 m long Kastenlot core MSM5/5-723-2 from the eastern Fram Strait (79° 9′ N, 5° 20′ E, 1350 m water depth) was retrieved in summer 2007 during cruise leg MSMS5/5 with RV Maria S. Merian. The uppermost section of the sediment core (4.75 m) was investigated at high resolution for proxy climate indicators such as biomarkers, foraminiferal assemblages and stable isotopes, and contents of ice-rafter debris. The studied section consists of dark olive to brownish black silty clays where from visual core inspection no evidence for abrupt changes in sedimentation rates such as hiatuses etc. could be indicated. Age control of the sediment core is based on six accelerator mass spectrometry (AMS) radiocarbon dates which had been published previously by Müller et al. (2012). To the existing age-depth-model we add two more dates (Table 1; Fig. 3). Analyses were conducted at the Leibniz Laboratory of Kiel University using ca 10 mg of CaCO3. All measurements were carried out on a single species N. pachyderma. Radiocarbon dates were converted to calendar years BP (present = 1950 AD) applying the calibration software Calib version 7.1 (Stuiver and Reimer, 1993) with the Marine13 calibration data set (Reimer et al., 2009), including a reservoir correction of ~400 years. Chronology is established using the calibrated calendar ages and assuming uniform sedimentation rates between them by linear interpolation.
In addition, previously published results of sediment core MSM5/5-712-2 from the eastern Fram Strait (Müller et al., 2012; Werner et al., 2013; Aagaard-Sørensen et al., 2014a) will be discussed and complement the new SST100 reconstruction.

3.1. Foraminiferal-derived climate proxy indicators

Sediment samples were freeze-dried, and dry bulk density was determined from defined 10 cm³ samples at 10–20 cm intervals. From samples washed through a 63 μm sieve prior to analysis, a representative split of at least 300 planktic foraminifer specimens was counted in size fraction 100–250 μm and identified to species level in order to apply the Husum and Hald (2012) transfer function and derive sea subsurface temperatures for the past 12,000 years (see 3.2).

In addition, fragmentation of planktic foraminifer tests has been monitored in size fraction 100–250 μm by counting planktic foraminifer test fragments that were no more determinable to species level.

Fig. 1. Map of the study area showing the investigated core location MSM5/5-723-2 (asterisk) as well as other sediment cores mentioned in the text (filled circles). Warm and saline Atlantic Water advection and cold polar surface waters of Arctic origin are indicated by red and blue arrows, respectively. The dashed lines show approximate recent locations of the maximum sea-ice margins after Vinje (2001). EGC East Greenland Current, ESC East Spitsbergen Current, NAC North Atlantic Current, NCC North Cape Current, NwCC Norwegian Coastal Current, RAC Return Atlantic Current, TPD Transpolar Drift, WSC West Spitsbergen Current. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.).

Please cite this article in press as: Werner, K., et al., Holocene sea subsurface and surface water masses in the Fram Strait – Comparisons of temperature and sea-ice reconstructions, Quaternary Science Reviews (2015), http://dx.doi.org/10.1016/j.quascirev.2015.09.007
Fragmentation (%) was calculated following the procedure by Pfuhl and Shackleton (2004) (Eq. (1)):

\[
\text{Fragmentation} (\%) = \frac{\# \text{planktic foraminiferal fragments/g}}{(\# \text{pl.foram.fragments/g})^3 + (\# \text{pl.foram.tests/g})} \times 100
\]  

(1)

Stable isotope analysis was carried out from picked planktic foraminiferal species *N. pachyderma* and *T. quinqueloba* (ca 25 specimens each, size fraction 100–250 \(\mu\)m) as well as the benthic foraminifer species *Cibicidoides wuellerstorfi* (ca 10 to 15 specimens, size fraction 250–500 \(\mu\)m). Tests were crushed and well-mixed prior to measurements using a Finnigan MAT 253 mass spectrometer system and a Kiel IV Carbonate Preparation Device. All measurements were calibrated to Pee Dee Belemnite (NBS 19). Analytical accuracy was <0.06\% for \(\delta^{18}O\) and <0.03\% for \(\delta^{13}C\).

### 3.2. Transfer functions

The transfer functions by Husum and Hald (2012) have been used to calculate sea subsurface temperatures (SST100) based on planktic foraminiferal fauna assemblages. This approach is based on a comparison between modern oceanographic data (summer SST at 100 m depth) and compositions of planktic foraminifer fauna in surface sediment samples from the northern North Atlantic and the Fram Strait. The method by Husum and Hald (2012) applies a size fraction of >100 \(\mu\)m. For the transfer function, the procedure of the weighted averaging partial least squares (WA-PLS) has been applied.
and sterol fractions from the total lipid extract was achieved via open column chromatography using hexane and methylacetate:n-hexane (20:80 v/v), respectively. Coupled GC–MS analyses (Agilent 6850 GC, Agilent 5975 C VL mass selective detector) were performed for the identification and quantification of highly branched isoprenoids and sterols. For further analytical details we refer to Müller et al. (2012) and Müller and Stein (2014). While the DIP25 index is calculated from IP25 and C25:0 diene GC–MS peak areas, PIP25 indices used here were calculated following Müller et al. (2012).

4. Results

4.1. Planktic foraminifer assemblages and fragmentation

N. pachyderma percentages are relatively high from ca 12 to 10.6 cal ka BP and strongly decrease thereafter while T. quinqueloba becomes dominant (>80%) at this time (Fig. 4a, b). Minimum abundance of N. pachyderma is noted at ca 10 cal ka BP, thereafter it returns stepwise. Local maxima of N. pachyderma between ca 10 and 5 cal ka BP occur at 9.4, 8.5–8.1, 6.9, 5–5.8, and at ca 5 cal ka BP. Accordingly, during other times, T. quinqueloba is dominant in the foraminiferal record. The significant increase of N. pachyderma between 5.5 and 5 cal ka BP is remarkable as thereafter, except for the uppermost sample, N. pachyderma stays at relatively high levels (75–50%). Slight increases of T. quinqueloba are noticeable during the later part of the record: at ca 2 and 1.5 cal ka BP (40%) and during the past decades where T. quinqueloba reaches >50% and N. pachyderma decreases to <30%.

Planktic foraminifer concentration and fluxes are low until 10.6 cal ka BP but higher and highly variable during the early Holocene part of the record until ca 5.2 cal ka BP (Fig. 4c, d). Thereafter, the concentration (ind./g dry sediment) slightly decreases and remains relatively stable except for single peak values at ca 2 cal ka BP and a low value at ca 1.5 cal ka BP. Fluxes slightly decrease after 5.2 cal ka BP and display the same trends as the concentrations. Planktic foraminifer fragmentation reaches highest values between 4.8 and 3.5 cal ka BP (Fig. 4e). Other shorter maxima are noticed after 10 cal ka BP, around 11 cal ka BP, before 8.5 cal ka BP, and between 7.5 and 6 cal ka BP. Fragmentation is low after 3.5 cal ka BP.

4.2. Sea-subsurface temperature reconstruction

Both ecological transfer functions discussed in this paper are based on comprehensive data sets of modern foraminiferal assemblages in the Fram Strait area and the northern North Atlantic (Pflaumann et al., 1996; 2003; Husum and Hald, 2012). Fig. 4f shows Holocene variability of subsurface water mass temperatures in sediment core 723-2. SST100 are rather low at 12 cal ka BP but gradually increase from 2 to 4°C until ca 10.7 cal ka BP. A further steep increase in SST100 is noted thereafter when temperatures rise to maximum values of 5 and 6°C around ca 10.2 cal ka BP. While gradually decreasing temperatures stay always above 5°C until ca 8.6 cal ka BP, from 8.5 to 8.1 cal ka BP a two-phase low in SST100 is observed with minimum values of less than 4.5°C at 8.5 and 8.1 cal ka BP and a slightly warmer phase of ca 4.9°C in between. After 8.1 cal ka BP, temperatures return to values between 5 and 5.5°C until ca 7.5 cal ka BP. Thereafter, a stepwise rather steep decrease in SST100 is found towards minimum temperatures of 3.5°C at 6.6 cal ka BP. Another warming follows, though with high fluctuations, until ca 6 cal ka BP (up to 5°C ). After 6 cal ka BP, temperatures drop stepwise to 3°C until ca 3 cal ka BP. Warmer temperatures are indicated thereafter between 2.5 and 1.0 cal ka BP (on average ca 3.6°C). Within the uppermost
sediments, a strong temperature increase is found (−5 °C) similar to previous findings by Spielhagen et al. (2011) in sediment core MSM5/5-712-1.

Likewise, SST100 were reconstructed from sediment core MSM5/5-712-2 (Fig. 4g) with the same approach indicating similar trends and temperatures, though slight differences are observed between the two sediment cores. Notably are maximum Holocene temperatures of 6.5 °C occurring around 10 cal ka BP which are slightly higher than in 723-2. Early Holocene short-term cooling episodes (between 2.5 and 3.5 °C) are more pronounced in 712-2 than in 723-2 occurring at 9.2, 8.1, 6.9 and 6.1 cal ka BP. After 5.2 cal ka BP a strong decrease in temperatures is noted from 5 to 3 °C. SST100 stay between 2.5 and 3.5 °C during the late Holocene with slightly warmer episodes around ca 3 cal ka BP and from 2 to 1 cal ka BP.

4.3. Foraminiferal stable isotopes

Foraminiferal stable isotope data are shown in Fig. 5. Stable oxygen isotope values of N. pachyderma are strongly fluctuating around 12 cal ka BP (2.2–4.6‰) (Fig. 5a). Until ca 11.0 cal ka BP, δ18O values are relatively high but drastically decrease thereafter. From ca 10.5 to 7.5 cal ka BP N. pachyderma δ18O values are generally lower with respect to the preceding and ensuing intervals, except for a short peak at ca 8.7 and 8.2 cal ka BP. After ca 7.5 cal ka BP, N. pachyderma δ18O values increase until ca 6.5 cal ka BP. A short minimum is noted thereafter while values in general increase again until ca 3.8 cal ka BP. Afterwards a slight but continuous decrease in N. pachyderma δ18O values is noted.

The δ18O record of T. quinqueloba starts at 10.5 cal ka BP (Fig. 5a). Values are highly fluctuating throughout the record. However, in
Fig. 5. a, b) Holocene stable isotope data from three foraminifer species in core M5/5-723-2. c) Ice-rafted debris and portion of rounded quartz grains (as calculated from total quartz grains). The orange bar highlights the 8.2 cal ka BP-event (see Section 5.3). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
general lighter values are noticeable during the early Holocene until ca 5.2 cal ka BP. Strong short-term increases during that period are found before and after ca 10 cal ka BP as well as at ca 6 cal ka BP. A general, less variable increase in *T. quinqueloba* δ^{18}O values is observed between 5 and 3.5 cal ka BP. Thereafter values decrease with higher variability.

Stable oxygen isotope values of the epibenthic species *C. wuellerstorfi* are highly fluctuating around 12 cal ka BP and are relatively high thereafter until ca 10.5 cal ka BP (Fig. 5a). A stepwise increase in δ^{18}O values is noted thereafter with higher fluctuations and an decrease after ca 9 cal ka BP. In the following, δ^{18}O values are fluctuating around an average of ca 4.1‰ with some increase between 5 and 4 cal ka BP and lighter values since ca 0.5 cal ka BP.

Stable carbon isotopes are shown in Fig. 5b. *N. pachyderma* δ^{13}C values stay rather low from 12.0 to 10.0 cal ka BP but slightly increase around 11.5 cal ka BP (Fig. 5b). After 10 cal ka BP, a gradual increase is noted lasting until ca 3.8 cal ka BP. A trend to lower values is found thereafter until ca 1 cal ka BP. During the last 1000 years, *N. pachyderma* δ^{13}C values are slightly increasing again. The youngest δ^{13}C values are strongly decreasing.

Carbon isotope values of *C. wuellerstorfi* are very low after 12 cal ka BP but strongly increase after 11 cal ka BP (Fig. 5b). Thereafter they stay on a high level with minor fluctuations (local minima found at 10 to 9 and 5.8 to 3.8 cal ka BP) with a trend to slightly higher values in the last 4 cal ka BP. Again, the youngest values are strongly decreasing.

*T. quinqueloba* δ^{13}C values are on average 1.5‰ lower and more variable than those of the two other species (Fig. 5b). They generally follow the trends of the *C. wuellerstorfi* record.

### 4.4. Ice-rafted debris

Lithic particles in the size fraction 150–250 μm reveal a negative trend around 11 cal ka BP (Fig. 5c). However, after ca 9 cal ka BP three significant IRD peaks are noted: after 9, after 8.5, and at ca 7.5 cal ka BP. After 7 cal ka BP, IRD increase stepwise but strongly until ca 4 cal ka BP. A minimum around 2 cal ka BP is followed by higher values in the last millennium.

Percentages of rounded quartz grains are highly variable throughout the record but show a positive trend from 11 to ca 3 cal ka BP (Fig. 5c). A minimum is noted between 7 and 6 cal ka BP. After 3 cal ka BP, values gradually decrease with a minor peak at ca 2 to 1 cal ka BP.

### 4.5. Biomarker data

TOC and CaCO₃ contents as well as concentrations of biomarkers and IP₂₅ and DIP₂₅ indices are shown in Fig. 6. A TOC minimum (0.7wt.%) is observed between 12 and 11.7 cal ka BP, which is followed by a significant increase in TOC from 11.5 cal ka BP reaching maximum values (1.8wt.%) at 11 cal ka BP (Fig. 6f). Thereafter variability is low throughout the remainder of the record with values in the range of 1–1.4wt.%. A gradual though slight increase is noted between 7.2 and 0.5 cal ka BP.

CaCO₃ content is low (3–5wt.%) until ca 10.5 cal ka BP but increases stepwise thereafter (Fig. 6g). It is highly variable between 8.4 and 8 cal ka BP with local maxima between 7.5 and 7 cal ka BP as well as 6 and 5.5 cal ka BP. Thereafter CaCO₃ slightly increases. Noticeable is the high variability with higher values from ca 2.5 cal ka BP.

Concentrations of the phytoplankton-derived biomarkers dinosterol and brassicasterol (Kanazawa et al., 1971; Boon et al., 1979; Volkman, 2006) are highly variable throughout the entire record (Fig. 6d, e). A general increase is observed for dinosterol during the late Holocene. In strong contrast to the phytoplankton biomarkers, IP₂₅ concentrations are less variable (Fig. 6c). A distinct decrease in IP₂₅ concentrations from maximum values of 0.6 μg/g to 0.1 μg/g OC occurred between 12 and 11.2 cal ka BP. Thereafter, concentrations remain low until ca. 7.2 cal ka BP, when a slight shift to higher values occurred and a successive increase in IP₂₅ concentrations is observed for the mid and late Holocene (Fig. 6c).

The DIP₂₅ or diene/IP₂₅ index which describes the C₂₅-HBI diene to IP₂₅ ratio exhibits a relatively high variability (Fig. 6b). Low values of 3 are noted until 11.4 cal ka BP. Between 11 and 6.5 cal ka BP, DIP₂₅ values oscillate significantly between minimum values of 3 and maximum values of 15. Less variable and generally decreasing DIP₂₅ ratios are observed after 7 cal ka BP (Fig. 6b).

Dinosterol and brassicasterol-based IP₂₅ indices (PdIP₂₅, PsIP₂₅; Müller et al., 2011) both show parallel trends of strongly decreasing values from 11.7 to 11 cal ka BP (Fig. 6a). Values stay low (0.3–0.4) with few fluctuations until ca 7 cal ka BP. Noticeable is a short-term minimum at ca 9 cal ka BP and a subsequent maximum between 9 and 8.5 cal ka BP. PsIP₂₅ and PdIP₂₅ indices both increase consistently from 7 to 0.5 cal ka BP, when they reach similar high values (ca. 0.7–0.8) as observed for 11.7 cal ka BP.

### 5. Discussion

#### 5.1. Sub-surface temperature reconstructions in the eastern Fram Strait area

The sediment record consists of two dominant planktic foraminiferal species that are negatively correlated (Fig. 4a, b): the polar planktic foraminifer species *N. pachyderma* (Sistmich et al., 2003) and the subpolar species *T. quinqueloba*, which is indicative for warm and saline Atlantic Water advection (Volkman, 2000). On average, other planktic foraminifer species only contribute by 4.8% of the entire fauna. The fauna clearly shows polar conditions from 12 to 10.6 cal ka BP when *T. quinqueloba* becomes dominant therefrom (Fig. 4b). Warmer conditions remain until ca 5 cal ka BP when relative abundances of *T. quinqueloba* and *N. pachyderma* become more similar.

We like to note here, that we are aware of the fact that foraminiferal transfer functions are only one of several possible approaches to estimate past seawater temperatures. For our interpretation, we use general trends indicated by temperatures records rather than absolute numbers. Due to the precision limits of these methods, we carefully abstain from over-interpreting small-scale (less than 0.5 °C) changes in temperatures.

The Holocene subsurface water temperature reconstruction of core 723-2 using the 100–250 μm size fraction of planktic foraminifer fauna (Husum and Hald, 2012) strongly aligns with the abundance of the planktic foraminifer species *T. quinqueloba* in the same size fraction (Figs. 4b, f and 7b). These parallel trends can also be observed for sediment core 712-2 (Figs. 4b and 8d). Hence it appears that the current reconstructed SST100 are driven mainly by the relative abundance of *T. quinqueloba*. Since presence of *T. quinqueloba* in the eastern Fram Strait area has been related to an increased advection of warmer Atlantic Water (AW) masses (e.g., Carstens et al., 1997; Volkman, 2000; Pados and Spielhagen, 2014), we conclude the reconstructed temperature variations are caused by the temperature fluctuations due to variable AW contribution rather than fluctuations of the Arctic Front with which *T. quinqueloba* has been associated within the Nordic Seas (Johannessen et al., 1994). It has to be noted that planktic foraminifer fragmentation is high between ca 4.8 and 3.5 cal ka BP (Fig. 4) indicating that less robust species as *T. quinqueloba* may be underrepresented during this interval (cf. Zamelczyk et al., 2012;
resulting in too low temperatures reconstructed during this time. However, the decrease both in *T. quinqueloba* percentages and carbonate concentration after 5 cal ka BP are parallel to an increase in $\delta^{18}O$ of *T. quinqueloba* (Figs. 4b, 6g and 7e). This supports a conclusion on generally colder surface water conditions not only during the time of the pronounced increase in planktic foraminifera fragmentation at 4.8–3.5 cal ka BP but also during short-term increases of fragmentation at 11.2–10.5, 8.6, and 7.5–6.6 cal ka BP when increased planktic $\delta^{18}O$ indicate contributions of cold waters (Fig. 7d, e and h). A selective removal of dissolution-prone species such as *T. quinqueloba* by CO$_2$-enriched, low-pH Arctic waters has been discussed by Zamelczyk et al. (2014). Likewise, we are able to use the fragmentation proxy here as an additional indicator of rather cold surface water conditions due to enhanced impact of cold, corrosive Arctic surface waters to the study area. Whether subpolar foraminifera were present in the area or not, AW advection was likely weakened at the core location during these intervals, and dominant Arctic surface waters might have caused partial dissolution of foraminifera.

A similar temperature reconstruction for the past ca 9000 years has been previously achieved by means of the SIMMAX transfer function for sediment core 712-2 (Fig. 8d; Werner et al., 2013) derived from a slightly coarser size fraction of counted planktic foraminifera (150–250 $\mu$m). Calculated water temperatures at 50 m
water depths (furthermore referred to as SST50) indicate the same feature being controlled by abundance of the two dominant species *T. quinqueloba* and *N. pachyderma*.

When comparing trends and absolute numbers of the past ca 9000 years of both temperature reconstructions, in general SST50 during the early Holocene do not reveal large differences from reconstructed SST100 (Fig. 8d). There are, however, differences noted during colder events when SST100 indicate slightly higher temperatures than SST50 (about 0.5–1 °C difference at ca 7.8, 7.5, 6.9, 6.1, 5.2, 5–4.5 and 4–3.5 cal ka BP). For one, this difference may

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**Fig. 7.** a) Insolation at 78°N for June, July, August, and September (Laskar et al., 2004), b) Reconstructed SST100 (light grey) including 3-point running means (dark grey; Husum and Hald, 2012), c) DIP25 index derived from biomarker data, d) *N. pachyderma* stable oxygen isotopes, e) *T. quinqueloba* stable oxygen isotopes incl. 3-point average mean, f) PBIP25 index derived from biomarker data and g) *N. pachyderma* percentages, h) Fragmentation of planktic foraminifer tests and i) ice-rafted debris. Grey bars refer to periods when elevated fragmentation has been found. The yellow and orange bars highlight data linked to the 8.2 cal ka BP-event (see Section 5.3) and the modern warming, respectively. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.).
be attributed to the different temperatures at the different water depths used in the reconstructions, and thus cold fresh Arctic waters may have a higher impact to the shallower SST50 reconstruction. Predominantly, this difference is likely related to the higher presence of small-sized *T. quinqueloba* specimens in the sediment fraction considered by the Husum and Hald (2012) reconstruction while SIMMAX estimates are based on a coarser size fraction. This is good in agreement with findings by Bauch (1994) that average *T. quinqueloba* size is reduced under lower temperatures. Higher abundance of small-sized specimens of *T. quinqueloba* during colder periods has also been noted by previous studies in the area (e.g., Bauch et al., 2001; Kandiano and Bauch, 2002; Nørgaard-Pedersen et al., 2007). We therefore strongly support the recommendation of earlier works (e.g., Kandiano and Bauch, 2002) to include smaller-sized planktic foraminifera in temperature reconstructions for areas with colder water masses.

In addition, occurrence of *T. quinqueloba* within the nutrient-rich marginal ice zone in the Fram Strait has also been discussed previously (e.g., Carstens et al., 1997; Pados and Spielhagen, 2014). Preference of depth habitats of *T. quinqueloba* but also of *N. pachyderma* were associated with depth of maximum food availability, in particular the position of the deep chlorophyll maximum (Kohlfeld et al., 1996; Pados and Spielhagen, 2014). Maximum abundance of *T. quinqueloba* may thus be used in...
addition to conclude on nutrient-rich subsurface waters especially during the early Holocene when *T. quinqueloba* reached abundances of >80% in eastern Fram Strait (Fig. 4b).

5.2. Early Holocene thermal maximum conditions in surface to subsurface water conditions

A retreat of spring sea ice and icebergs due to rising sea surface temperatures in the earliest phase of the Holocene is supported by decreasing IRD content and a significant drop in *P. pachyderma* and *P. pachyderma* values with coincidently rising DIP25 values, the latter indicative for surface water temperatures SST10 (Figs. 6 and 7). Elevated and even maximum concentrations of the phytoplankton-derived biomarkers brassicasterol and dinosterol and low concentrations of the sea ice biomarker IP25 between 11.0 to 10.5 and 9.5 to 9 cal ka BP (Fig. 6c–e) suggest a higher primary productivity in response to a reduced sea-ice cover. However, between 11 and 10.6 cal ka BP, subpolar planktic foraminifer abundance (and thus planktic foraminifer SST100, 2–4 °C) are still relatively low and strongly fluctuating, a pattern that can be similarly observed in the neighbor-bouring core MSMS5-712-2 (Aagaard-Sørensen et al., 2014a). The delayed record of subpolar planktic foraminifera may be related to the higher plakinatic foraminifer fragmentation during that time (Fig. 7b) and a preferential removal of the dissolution-prone *T. quinqueloba* when corrosive low-pH Arctic waters prevailed at the surface in proximity to the summer sea-ice margin. This conclusion may be further supported by low calcium carbonate concentration and low planktic foraminifer flux and abundances (Figs. 4c, d and 6g). A further reason for the disparities between the onset of warming at surface and subsurface waters may also be the difference in maximum Holocene insolation depending on the summer month. Maximum phytoplankton blooms have been observed to occur in spring (May–June) in the northern North Atlantic (Wassmann et al., 1991; Baurerfeind et al., 1994). Surface water proxies might have responded to the very high radiation maximum in June at ca 11 cal ka BP. As a consequence, by the end of the spring, surface waters might have been nutrient-depleted due to an intense spring phytoplankton bloom, which hence limited later occurring zooplankton blooms, in agreement with Jonkers et al. (2010) suggesting food availability as one of the major controls on planktic foraminifer fluxes. Maximum occurrence of planktic foraminifers may therefore be linked only to the time around 10 cal ka BP when maximum insolation occurred later during summer (Fig. 7) and more nutrients might have been still available in (sub-)surface waters.

Accordingly, SST100 only rapidly increased to maximum values at 10.6 ka, which was significantly later than the decrease in the IP25 indices, and clearly after the early peaks in Dinosterol, Brassiater at 10.6 ka, which was significantly later than the decrease in the IP25 indices, and clearly after the early peaks in Dinosterol, Brassiatercater, and DIP25. At the same time, 13C of *N. pachyderma* is rather low in both cores from the Fram Strait (this study; Aagaard-Sørensen et al., 2014a). Low 13C values may be related to weak ventilation of the water masses inhabited by *N. pachyderma* (Spielhagen and Erlenkeuser, 1994), Werner et al. (2013) suggested a deeper habitat depth of *N. pachyderma* in less ventilated water masses potentially due to the strong prevalence of fresh and cold sea-ice derived water masses at the surface, which might have been the case during the earlier part of the Holocene.

The drastic increase in *T. quinqueloba* and SST100 at 10.6 cal ka BP may primarily be explained by a shift in water mass contribution around 10.6 cal ka BP which only occurred within about 60 years, similar to the amplitude and rate of change in subsurface water temperatures in the eastern Fram Strait during the last ca 120 years (Spielhagen et al., 2011; see also Fig. 7b). From both decreased IRD contents and planktic foraminifer fragmentation since ca 10.6 cal ka BP we conclude on a retreat of the sea ice margin to a position north of the core location and a subsequent strong warming/prevalence of warm Atlantic Water at surface to subsurface depth from 10.6 cal ka BP onwards. To a large extent, this may be caused by a maximum of northward heat advection at ca. 10 cal ka BP as concluded by Risebrobakken et al. (2011) from seawater temperature reconstructions in the Nordic Seas and Barents Sea.

By far, the large amplitude of 3–10 °C in summer SST reconstructed by dinocysts (de Vernal et al., 2013) cannot be reproduced by foraminifer-derived reconstructions of maximum subsurface temperatures 5–6 °C between 10.6 and 9.0 cal ka BP. This difference in amplitude could be explained by e.g., a highly variable stratification with subsurface waters of more or less constant temperatures but highly variable surface water temperatures but also by differences between proxies in seasonality or their depth habitats. Low values and little variability seen in the sea-ice indicators shown here, however, suggest rather stable conditions with a the study sites. Stable conditions during the critical interval are as well reported by reconstructions from the Nordic Seas (Risebrobakken et al., 2011).

A gradual decrease in foraminiferal water temperatures can be noted soon after ca 10 cal ka BP. This might be partially related to decreasing July insolation and coherent cooling of surface waters as well as consequences of an expanded marginal ice zone (Fig. 7a, 1). However, strong Atlantic Water advection likely prevailed dominantly at subsurface depth which was also indicated further on the Western Svalbard slope (Rasmussen et al., 2013).

Between ca 10 to 9 cal ka BP, surface and subsurface water proxies, all indicate higher seawater temperatures (Fig. 8a–c). Regardless of the very high temperatures, the SST10 reconstruction based on dinoflagellate cysts in core 712-2 indicates higher, though strongly fluctuating summer month surface water temperatures at 10 to 9 cal ka BP (de Vernal et al., 2013, Fig. 8a). Thus, from similar temperature trends indicated by surface (biomarkers, dinocysts-based) and subsurface (foraminiferal-derived) proxies, we conclude on the presence of a thick and warm summer mixed layer at least until ca 9 cal ka BP with probably limited stratification within the uppermost water column. This thick layer of northward advected strong Atlantic Water likely occupied the uppermost ca 200 m including near-surface waters in the eastern Fram Strait area. Maximum insulation during early summer (June and July, Fig. 7a) probably played an important additional role.

From studies on the shelf and slope of the Western Svalbard margin, Rasmussen et al. (2013) attribute differences in surface to subsurface water temperatures to the variable influence of polar and melt waters derived from inland and east of Spitsbergen. They find strong stratification of the upper water column until ca 9.6 cal ka BP with Arctic and melt waters at surface (30–100 m depth) and warm Atlantic Water below. Apparently, water masses in the eastern Fram Strait further off the Western Svalbard shelf were not as much influenced from these cold water masses found at the surface on the shelf and slopes.

Dinocyst-derived summer surface temperatures suggest an abrupt shift from highly variable and generally warmer to less variable and overall cooler SST10 after ca 9 cal ka BP (Fig. 8a; de Vernal et al., 2013). Such a rapid sea surface cooling is, however, not reflected in the biomarker records. Slightly decreasing IP25 and PIP25 values even seem to refer to an ongoing retreat of the sea ice until 7 ka BP; when a distinct rise in PIP25 values and generally decreasing DIP25 ratios indicate a successive re-expansion of sea ice and decreasing SST10, respectively (Fig. 7c, f). The disparities between the different surface water proxies (biomarkers versus dinocysts) remain unclear at this point. To a certain extent they might not be equally sensitive to environmental changes. Dinocyst data reflect surface water changes in the summer season while IP25 reacts to ice coverage. Furthermore, the strong fluctuations in the dinocyst-
derived SST10 record may also be caused by changes in nutrient availability (Devillers and de Vernal, 2000).

From latitudinal and temporal discrepancies of alkenone versus foraminiferal-derived seawater temperature reconstruction in the Nordic Seas and Barents Sea, Risebrobakken et al. (2011) discussed the separate roles of maximum insolation and strong oceanic heat advection during the early Holocene with different impacts to surface and subsurface water layers. While maximum northward heat advection (as concluded from subsurface-inhabiting planktonic foraminiferal temperatures) occurred already at 10 cal ka BP, the summer mixed layer (indicated by alkenone-derived temperatures) in the Nordic Seas seems to be influenced by maximum insolation slightly later from ca. 9 to 6 cal ka BP (Risebrobakken et al., 2011). For their argumentation, the authors used August insolation with a Holocene maximum around ca 8 cal ka BP. Maximum June and July insolation, however, occurred earlier around 11 to 10 cal ka BP, respectively. In addition, June maximum radiation was higher compared to insolation maxima of later summer months (Fig. 7). To date there are only few studies available on the seasonal cycle of planktonic foraminifera from the northern part of the North Atlantic (e.g., Stangeew, 2001; Jonkers et al., 2010; 2013). From the Irminger Sea, Jonkers et al. (2010) report on a bimodal flux pattern of N. pachyderma and T. quinqueloba in the northwestern and eastern Fram Strait (Fig. 5). Foraminiferal-derived seawater temperatures during summer with a maximum in late September. This pattern may be different both for the Fram Strait and during the early Holocene. Thus, one reason why we do not see a clear separation of surface and subsurface maximum water temperatures during the early part of the early Holocene may be the difference in the seasonal cycle of planktonic foraminifera maximum fluxes at different latitudes.

5.3. Pronounced shifts in surface to deepwater proxy records during the ‘8.2 ka’ climate anomaly

A sudden negative trend, concurrent to the freshwater outburst of the glacial lake Agassiz (e.g., Barber et al., 1999), is obvious for few hundred years in foraminiferal temperatures at ca. 8.5 and 8.2 cal ka BP with subsurface water temperatures below 4.5 °C (Fig. 7b). This trend is accompanied by local maxima in δ13C of N. pachyderma and T. quinqueloba as well as in ice-rafted debris (Fig. 7d, e and h). Furthermore, a decrease in the dolostone concentration and in DIP2S at ca. 8.2 cal ka BP are noted, the latter suggesting decreased surface water temperatures (Fig. 6b, d). We also find slightly reduced amounts of rounded (versus angular) quartz grains during the time of the anomaly (Fig. 5c). This might indicate increased sediment supply of glacial/subglacial origin rather than of ocean-rafted material from beaches etc. to have contributed to ice-rafted debris during that time. We like to note that some of the proxies mentioned above indicate short-term shifts of the same or larger amplitude already between 8.8 and 8.7 cal ka BP (Fig. 5) which may be of rather local origin. As for the cooling around 8.2 cal ka BP, similar features have been found in the neighbouring core 712-2 (Werner et al., 2013) with cool surface and subsurface water conditions around 8.2 cal ka BP we suggest a rather strong impact of the North Atlantic Deep Water renewal further south (Kleiven et al., 2008). Interestingly, the low values in benthic foraminiferal carbon isotopes are accompanied by local increases of δ13C of N. pachyderma and T. quinqueloba. Since the signal appears in all foraminiferal δ13C records of cores MSM5/5-723-2 and MSM5/5-712-2 we assume a change in seawater origin had affected the entire water column in eastern Fram Strait during the 8.2 cal ka BP event.

5.4. Decoupling of surface from subsurface water conditions during the late Holocene

Decreasing insolation during the late Holocene led to changes in water mass properties in the eastern Fram Strait such as a surface cooling accompanied by extended sea-ice extent (Müller et al., 2012; Werner et al., 2013). Extended sea-ice extent in the Fram Strait during the so-called Neoglacial period might have been the result of, amongst other factors, a high production and subsequent export of Arctic sea-ice with the Transpolar Drift via Fram Strait into the Nordic Seas (Werner et al., 2013). Similar to results from sediment core MSMS/5-510-12 (Werner et al., 2013), a significant decrease in T. quinqueloba percentages is noted at ca 5 cal ka BP, which is resembled in estimated SST10 (Fig. 7b). Temperatures reach minimum values of ca 3.1 °C between 4.6 and 3 cal ka BP, similar to findings of lowest temperatures in the area from 5 to 2 cal ka BP by Rasmussen et al. (2013). In addition, increased fragmentation of planktonic foraminifer carbonate shells between ca 4.8 and 3.5 cal ka BP parallel to an increase in δ18O of T. quinqueloba (Fig. 7f and h) point to generally colder subsurface water conditions and weakened AW advection. Higher amounts of rounded quartz grains between 6 and 3 cal ka BP may indicate enhanced entrainment of sediments by sea-ice from the shallow Arctic shelves where river load sediments have been identified as one of the sources for sea-ice sediments (Nürnberg et al., 1994).

While the sea-ice increase during the late Holocene has been extensively discussed in earlier studies (e.g., Müller et al., 2012; Werner et al., 2011, 2013), we like to further focus on the concurrently noted development of subsurface (Atlantic) waters.

From ca 2.5 to 1.0 cal ka BP, slight increases in T. quinqueloba percentages and estimated SST100 are noted in both cores from the eastern Fram Strait (this study; Werner et al., 2013). Increasing subsurface water temperatures derived from Mg/Ca ratios of N. pachyderma since 3 cal ka BP with a maximum of 5 °C at ca 1.1 ka (Aagaard-Sørensen et al., 2014b, Fig. 8e) furthermore suggest slightly strengthened AW advection in the study area. Evidence for increased AW inflow after 3 ka has been found on the slope and shelf of the Western Svalbard margin (e.g., Rasmussen et al., 2013), in the Franz Victoria Trough (Lubinski et al., 2001), the western Barents Sea (Berben et al., 2014), and in the Nordic Seas (Hald and Aspeli, 1997; Hald et al., 2007; Risebrobakken et al., 2003; 2011; Thornalley et al., 2009). Using data from the western Barents Sea shelf, Sarnthein et al. (2003) attributed an increase in T. quinqueloba percentages after 3 ka to a slightly strengthening of the meridional overturning circulation during the late Holocene.

The warming in subsurface waters suggested by SST100 reconstructions apparently contrasts the simultaneously occurring denser sea-ice cover on the Yermak Plateau and over core sites 723-2 and 712-2, as reconstructed from sea-ice indicative biomarkers and increased IRD contents (Müller et al., 2009; 2012; this study). Also, SST10 based on dinocysts as well as DIP2S values indicate rather cool conditions at the surface likely caused by spreading sea-ice coverage and sea-ice derived freshwater contributions. The apparent contradiction in surface versus subsurface waters can only be explained by simultaneously existing cold conditions at surface waters and increasing temperatures of the Atlantic Water layer, likely separated by a strong pycnocline during the late
Holocene. The strong stratification of the upper water masses was likely caused by longer winters due to the reduced solar forcing (Fig. 7a) and thus increased seasonality (Berben et al., 2014). As one possible explanation, the noted increases of late Holocene Atlantic Water advection at various sites in the northern North Atlantic may be related to the variable strength of the North Atlantic Current that was previously explained by dynamic changes in the relative contributions of water masses of the North Atlantic subpolar gyre (SPG) and subtropical gyre (STG) during the past and present (Hätún et al., 2005; Thoralley et al., 2009; Born et al., 2011; Montero-Serrano et al., 2013). Freshwater inputs to the Labrador Sea can prevent deep convection and reduce the SPG circulation resulting in enhanced northward flow of North Atlantic Drift waters which have entrained warmer and saline STG waters (Hätún et al., 2005). Thoralley et al. (2009) found evidence for such linkage throughout the major part of the Holocene in the northeast North Atlantic. The authors identify a negative relationship between Holocene salinity variations south of Iceland and in the Labrador Sea where phases of strong inflow of warm and saline Atlantic Water coincided with a SPG weakened by freshening.

Likewise, the described process holds an explanation for the apparent stratification in the eastern Fram Strait with warmer Atlantic Water temperatures versus surface–water cooling during the late Holocene. Increased sea-ice presence likely caused an effective freshening of surface waters not only in the Fram Strait but also in the Nordic Seas and in the region of the subpolar gyre. We suppose that this freshening caused a late Holocene weakening of the SPG, thereby allowing for enhanced northward transport of heat and salt by the North Atlantic Current eventually affecting strong Atlantic Water advection through the eastern Fram Strait region from 2.5 to 1.0 cal ka BP.

6. Summary

Holocene subsurface water temperature reconstructions of two sediment cores in the eastern Fram Strait following the approach by Husum and Hald (2012) based on planktic foraminiferan fauna in size fraction 100–250 μm strongly align with the abundance of the planktic foraminifer species T. quinqueloba in the same size fraction. During warmer periods, sea subsurface temperatures estimated by the Husum and Hald (2012) approach are similar to previously published SST50 reconstructions estimated by the SIMMAX transfer function (Pflaumann et al., 1996; 2003). However, during colder periods and events, the Husum and Hald (2012) method indicates higher SST100 due to a higher presence of small-sized T. quinqueloba specimens in sediment samples. We therefore strongly agree with earlier recommendations to preferably include smaller-sized planktic foraminifera into temperature reconstructions in the Fram Strait area.

For the first time, DIP25 values have been directly compared to reconstructed ocean temperature trends. Similar temperature trends as revealed by surface (biomarkers, dinocysts-based) and subsurface (foraminiferal-derived) proxies indicate the presence of a deep and warm summer mixed layer influenced by northward advected strong Atlantic Water between ca 10 and 9 cal ka BP with limited stratification within the uppermost water column. This warm and supposedly nutrient-rich Atlantic layer likely occupied the uppermost ca 200 m including near-surface waters in the eastern Fram Strait area and was additionally influenced by maximum early summer (June–July) insolation as the sea-ice extent for the period between ca 11 to 8.5 cal ka BP was strongly limited.

During the late Holocene, upper water mass stratification significantly strengthened in the eastern Fram Strait with warmer Atlantic Water advection at the subsurface, cold and fresh polar waters at surface water depth, and a pronounced pycnocline prevailing in between. Increased sea-ice presence likely caused an effective freshening of surface waters not only in the Fram Strait but also in the Nordic Seas and in the region of the subpolar gyre. Increasing supply of sea-ice derived polar freshwaters might have led to a weakening of the subpolar gyre activity and allowed for enhanced northward transport of heat and salt by the North Atlantic Current. Eventually, the strengthened North Atlantic Current probably intensified northward Atlantic Water advection through the eastern Fram Strait region into the Arctic Ocean from 2.5 to 1.0 cal ka BP.

Acknowledgements

We thank the science party and crew onboard RV “Maria S. Merian” during the expedition MSMS5/5 for retrieving the sediment core. For laboratory assistance, we thank Marieke Göser, Antonia Hofmann, Jan Oesterwalbesloh, David Poggemann, and Torben Struve. We kindly acknowledge Luizim Haxhij for technical assistance on stable isotope measurements. H. Christian Hass provided one of the additional AMS 14C datings. The German Research Foundation (DFG) provided financial support of KW, JM, and RFS within the Priority Programme 1266 (INTERDYNAMIC, Project HOVAC). KW was furthermore funded through the Byrd Foundation and NSF P2C2 Grant (Award No. 1404370). The manuscript benefited from productive discussion with Steffen Aagaard-Sørensen. Constructive comments by two anonymous reviewers are gratefully appreciated.

Appendix A. Supplementary data

Supplementary data to this article can be found online at http://dx.doi.org/10.1594/PANGAEA.849325.

References


