Recent planktic foraminifera in the Fram Strait: ecology and biogeochemistry

Dissertation

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Kiel, 2014

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Dissertation

in fulfilment of the requirements for the degree
"Dr. rer. nat."

of the Faculty of Mathematics and Natural Sciences
at Kiel University

submitted by

Theodora Pados

Kiel, 2014

First referee: Prof. Dr. Martin Frank

Second referee: PD Dr. Mara Weinelt

Date of the oral examination: 25.09.2014

Approved for publication: 25.09.2014

Signed: Prof. Dr. Wolfgang J. Duschl, Dean

Affidavit

I hereby affirm that apart from the supervisor´s guidance the content and design of the essay is all my own work. The thesis has been prepared subject to the Rules of Good Scientific Practice of the German Research Foundation and its content has been submitted neither partially nor wholly as a part of a doctoral degree to another examining body. Parts of the essay have been published, submitted for publication or are in preparation for submission.

Kiel, 20.07.2014

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Abstract

Since the species distribution and biochemical composition of planktic foraminifera reflect properties of the environment they inhabit, fossil assemblages in sediments are often used to reconstruct palaeoenvironments. However, for a correct interpretation of fossil data it is important to improve our understanding of the relation between recent environmental variability in the ocean and the related distribution of living foraminifera and the biochemical composition of their shells. A particularly relevant issue is the inconsistency in the results of Mg/Ca thermometry in high latitudes that contributes to the uncertainty of palaeo-reconstructions. The objective of this study is to obtain new data on the vertical and horizontal distribution of planktic foraminifera species in the Fram Strait (Arctic Ocean) and to determine the factors controlling the stable isotopic signature and the Mg/Ca ratios of the tests in these high latitude waters.

The results of this report show that in the Fram Strait the distribution pattern of the two dominating species, *Neogloboquadrina pachyderma* (sinistral) and *Turborotalita quinqueloba* has a distinct relation to the complex hydrography. High abundances in the marginal ice zone are related to the increased primary production triggered by the strongly stratified water column at the ice margin. The depth habitats of *N. pachyderma* (sin.) and *T. quinqeloba* in the study area seem to be predominantly controlled by the availability of food. We could not verify the hypothesis that associates the vertical distribution of *N. pachyderma* (sin.) to a discrete isopycnal band. The foraminiferal fauna living in the water column is fairly well reflected on the sediment surface and thus, the sedimentary record in the studied region provides a reliable proxy for palaeo-water mass characterization.

The average depths of calcification of *N. pachyderma* (sin.) and *T. quinqueloba* seem to have a similar range, but the "vital effect" in both, oxygen and carbon isotope composition shows different values in the two species. The discrepancies with earlier published results in the magnitude of the "vital effect" suggest that oceanographic variability plays an important role in the incorporation of stable isotopes in the foraminiferal tests. The increased offset between $\delta^{13}C_{DIC}$ and the $\delta^{13}C$ of shells in the upper part of the water column indicates the influence of the "carbonate ion effect" on the calcification process of the tests. The offset between the stable isotope

composition of the net-sampled shells and those from the core tops suggest that the tests on the sediment surface recorded changes in environmental parameters in the past ~200-1000 years. The negative offset in $\delta^{18}O$ between the sediment and plankton samples may be explained by the combined effect of temperature rise and $\delta^{18}O_{water}$-change, while the offset in $\delta^{13}C$ may be due to the surface ocean Suess effect.

The Mg/Ca ratios in both species show a very weak correlation with *in situ* water temperatures and at the same time, significantly elevated values compared to calibration predictions. The results may be explained by the effect of carbonate chemistry on the shells that might mask the temperature-dependence in the studied region. The offset in Mg/Ca between *N. pachyderma* (sin.) and *T. quinqueloba* most likely results from differences in their calcification processes.

Kurzfassung

Da die Artenverteilung und biochemische Zusammensetzung der Gehäuse von Foraminiferen die Umgebung widerspiegeln in der sie leben, werden die fossile Exemplare aus Sedimenten oft für paläozeanographische Rekonstruktionen verwendet. Um die fossilen Informationen richtig interpretieren zu können, ist es jedoch wichtig, unser Wissen über die Korrelation zwischen der rezenten Variabilität in den Ozeanen und der damit im Zusammenhang stehenden Artenverteilung und biochemischen Schalenzusammensetzung der Foraminiferen zu erweitern. Ein besonders relevanter Aspekt ist die Widersprüchlichkeit der Ergebnisse der "Mg/Ca-Thermometrie" in hohen Breiten, die zu den Unsicherheiten von Paläo-Rekonstruktionen beiträgt. Das Ziel dieser Studie ist deshalb die Erhebung neuen Daten, einerseits über die vertikale und horizontale Verteilung von planktischen Foraminiferen-Arten in der Framstraße (Arktischer Ozean), anderseits über die Faktoren, die die Isotopenzusammensetzung und den Mg/Ca-Gehalt der kalkschaligen Mikroorganismen in dieser Region beeinflussen.

Die Ergebnisse dieser Arbeit zeigen, dass die Verteilungen die zwei in der Framstraße dominierende Arten *Negloboquadrina pachyderma* (sinistral) and *Turborotalita quinqueloba* einen eindeutigen Zusammenhang mit der komplexen Hydrographie zeigen. An der Eisgrenze spiegelt sich die vom stark stratifizierten Wasser begünstigte hohe primäre Produktivität in höheren Häufigkeiten wider. Die Tiefenhabitate von *N. pachyderma* (sin.) und *T. quinqueloba* scheinen vorwiegend von der Nahrungsverfügbarkeit beeinflusst zu sein. Die Hypothese einer Korrelation der vertikalen Verteilung von *N. pachyderma* (sin.) mit bestimmten Schichten gleicher Dichte konnte nicht bestätigt werden. Die Foraminiferen-Fauna in der Wassersäule wird in sehr ähnlicher Artenverteilung auch in den Sedimentoberflächeproben gefunden, dieser Proxy ist also gut geeignet für Wassermassen-Rekonstruktionen.

Die durchschnittliche Kalzifikationstiefen von *N. pachyderma* (sin.) und von *T. quinqueloba* weisen ähnliche Werte auf, doch die in der Sauerstoff- und Kohlenstoffisotopenzusammensetzung erkennbaren "Vitaleffekte" sind in den zwei Arten unterschiedlich. Von den gemessenen Werten abweichende, früher publizierte "Vitaleffekte" weisen darauf hin, dass die ozeanographische Variabilität eine wichtige

Rolle bei der Einlagerung von stabilen Isotopen in die Schalen spielt. Der erhöhte Versatz zwischen den $\delta^{13}C$-Werten der Gehäuse und den $\delta^{13}C_{DIC}$-Werten im obersten Teil der Wassersäule lässt auf den Einfluss des "Karbonationeneffekts" auf die Kalzifikation schließen. Die Differenz in der Isotopenzusammensetzung zwischen den in der Wassersäule und von der Sedimentoberfläche gewonnenen Schalen deutet darauf hin, dass die Foraminiferengehäusen aus den Sedimenten auch die Umweltparameter-Veränderungen der letzten ~200-1000 Jahre reflektieren. Der negative Versatz in den $\delta^{18}O$-Werten zwischen den Sediment- und Planktonproben deutet auf den kombinierten Effekt von Temperaturanstieg und $\delta^{18}O_{Wasser}$-Veränderungen hin. Der negative Versatz in den $\delta^{13}C$-Werten hingegen lässt sich mit dem "Suess-Effekt" im oberflächennahen Ozean erklären.

Die Mg/Ca-Verhältnisse beider Arten zeigen eine sehr schwache Korrelation mit *in situ*-Wassertemperaturen; außerdem sind die Werte wesentlich erhöht, verglichen mit der Werten, die mit existierenden Kalibrationsgleichungen berechnet wurden. Eine Erklärung dafür könnte der Einfluss der Karbonat-Chemie des Wassers auf die Mg/Ca-Verhältnisse sein, der in dieser Region die Temperaturabhängigkeit abschwächt. Die Differenz in den Mg/Ca-Werten zwischen *N. pachyderma* (sin.) und *T. quinqueloba* resultiert wahrscheinlich aus Unterschieden im jeweiligen Kalzifikationsvorgang.

1. Introduction

1. 1 Extant planktic foraminifera

Foraminifera are testate marine protozoa with a variety of test composition and morphology (Fig. 1.1). The first planktic forms appeared in the mid-Jurassic. These first groups were most probably meroplanktic and have evolved from a group of benthic calcareous, hyaline foraminifera. In the early Cretaceous planktic foraminifera had become holoplanktic and widespread over the worlds ocean (Culver, 1993). All extant planktic foraminifera belong to the suborder Globigerinina (Kingdom: Protista, Subkingdom: Protozoa, Phylum: Sarcomastigophora, Subphylum: Sarcodina, Class: Granuloreticulosea, Order: Foraminiferida).

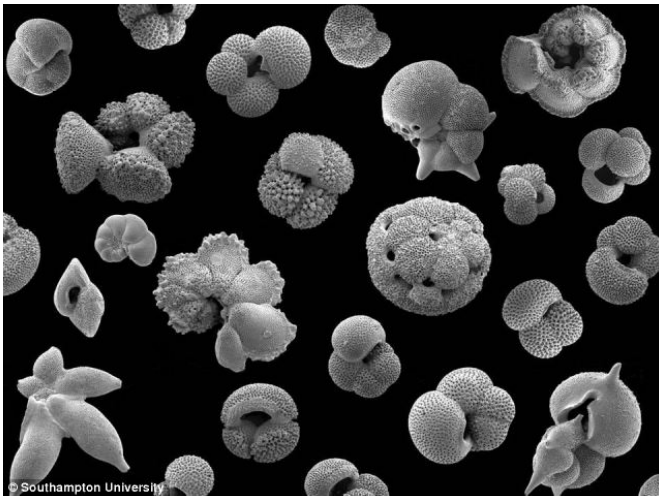

Figure 1.1. SEM image of a variety of planktic foraminiferal tests (Source: http://www.dailymail.co.uk/sciencetech).

Planktic foraminifera consist of single eukaryotic cells and secreted calcareous shells. The calcite wall is formed by the primary organic membrane (POM) that deposits calcite layers at the inner and the outer side of the POM (bilamellar wall). A monolamellar wall has been observed only in the genus *Hastigerina*. In particular areas of the early test-wall "normal" or "microperforate" pores and spines may be formed. Accordingly, on the lowest taxonomic level three informal morhpogroups are differentiated by the special characteristics of the calcite wall: spinose (all species from the superfamily Globogerinoidea), non-spinose (all species from the superfamily

Globorotaloidea) and non-spinose microperforate (species from the superfamily Heterohelocoidea) (Schiebel & Hemleben, 2005).

Planktic foraminifera inhabit species-specific water depths that range between the sea surface and more than 400 m. The dwelling depths are determined by ecologic and autobiologic factors (e.g., food availability, symbionts) and shift during ontogeny. Spinose species prefer animal prey while non-spinose species have been reported to be mainly herbivorous (Anderson et al., 1979; Caron & Be, 1984; Spindler et al., 1984). Many species living in the euphotic zone harbour dinoflagellate or chrisophyte symbionts. Algal symbiosis may provide energy from the photosynthesis, may enhance the calcification and may play a role in the removal of the metabolic products of the hosts (Hemleben et al., 1989; Hallock, 2003). In planktic species only sexual reproduction has been observed. Shallow-dwelling foraminifera are known to reproduce on a synodic lunar cycle once or twice a month (e.g., *Globigerina bulloides* and *Globigerinoides ruber*, respectively), possibly at the greatest depth level they enter during their ontogeny (Schiebel & Hemleben, 2005). Intermediate and deep-dwelling species (e.g., *Globorotalia truncatulinoides*) are thought to reproduce much less often and they ascend close to the sea surface to release gametes (Berberich, 1996; Schiebel & Hemleben, 2005).

Most of the modern species are ubiquitous. The assemblages can be attributed to five major faunal provinces: polar, subpolar, temperate, subtropical and tropical (Be, 1977). In the high and mid latitudes a seasonal distribution pattern is pronounced, following the primary production. In the Nordic Seas, for instance, the production maximum of planktic foraminifera occurs during summer (Kohfeld et al., 1996; Jonkers et al., 2010), with almost zero production during other seasons. In the subtopical-tropical oceans foraminiferal production is more balanced and consequently, the sedimentation of empty tests is less seasonally pulsed (Schiebel & Hemleben, 2005).

1.2 Foraminifera in palaeoceanography

Planktic foraminifera are proxies extensively used in palaeoceanography. Their constant presence through geological times since the mid-Jurassic, their ubiquitous prevalence, and the (often) good preservation of tests in sediments together provide a powerful tool to reconstruct past marine environments. Fossil specimens in sediment cores offer many possibilities to reveal information about ancient marine systems. The ratio between the abundances of different planktic foraminifera species – first used as a proxy by Schott (1935) – refer to the hydrographic regime, e.g., the distribution of water masses, water temperatures, and the position of the summer sea ice margin. Moreover, planktic foraminifera form their calcium carbonate tests from the surrounding seawater; the biochemical composition of the shells therefore reflects the properties of the ambient water. Oxygen isotopic composition ($^{18}O/^{16}O$, $\delta^{18}O$) of the tests gives information about variations in sea surface temperatures (first recognized by Emiliani (1954)) and salinities (e.g., Duplessy et al., 1992) in the past. Other widely used measures of palaeotemperature are the Mg/Ca ratio (e.g., Cronblad & Malmgren, 1981; Nürnberg et al., 1996) and the Ca isotope composition (Nägler et al., 2000) of the calcite tests. Planktic carbon isotope records ($^{13}C/^{12}C$, $\delta^{13}C$) are generally assumed to reflect changes in palaeoproductivity and ventilation of surface and near-surface waters (Duplessy, 1978), just like the Cd isotope composition (e.g., Boyle et al., 1976). The B isotope composition of the tests is related to the seawater pH (Sanyal et al., 1996). Additionally, there are several other trace elements in foraminiferal shells that serve as proxies for various marine biochemical processes.

Despite much effort and many calibration studies using modern foraminifera and water mass data, there are still uncertainties about the correct interpretation of the fossil data. Several studies revealed that foraminifera do not calcify their shells in equilibrium with the seawater (Shackleton et al., 1973; Vergnaud Grazzini, 1976; Kahn, 1979; Kohfeld, et al., 1996; Bauch et al., 1997; Volkmann & Mensch, 2001) and that the above-mentioned proxies (always or under certain conditions) do not have a linear correlation with the parameters they are thought to reflect. Moreover, in particular cases the degree of disequilibrium seems to vary regionally (e.g., in $\delta^{18}O$: Bauch et al., 1997; Volkmann & Mensch, 2001; Simstich et al.; 2003, Jonkers et al., 2010). Another particularly relevant issue that contributes to the uncertainty of palaeo-reconstructions are the inconsistent results of Mg/Ca thermometry in high latitudes (Nürnberg, 1995; Meland et al., 2006; Kozdon et al., 2009; Jonkers et al., 2013). For the reasons mentioned above, it is important to examine which biological, chemical or regional factors affect the incorporation of elements into the tests during calcification, and to improve our understanding of the correlation between recent oceanic variability and the distribution/biochemical composition of shells of living foraminifera.

1.3 The Arctic Ocean and the Fram Strait

The Arctic Ocean is an enclosed ocean with a total area of approximately 9.6 x 10^6 km^2. It is connected to the Pacific Ocean through the Bering Strait, and to the Atlantic Ocean through Fram Strait, the Barents Sea and the Canadian Arctic Archipelago via Baffin Bay (Fig. 1.2). The Arctic Ocean water mass structure can be characterized by the Polar Mixed Layer (PML) at the surface, a cold and saltier halocline between 50-200 m water depth, a warm Atlantic Layer (AL; 200-500 m water depth), the Upper Polar Deep Water (UPDW) extending down to a depth of 1700 m, deep water (down to 2500 m), and bottom water beneath that (Jones, 2001).

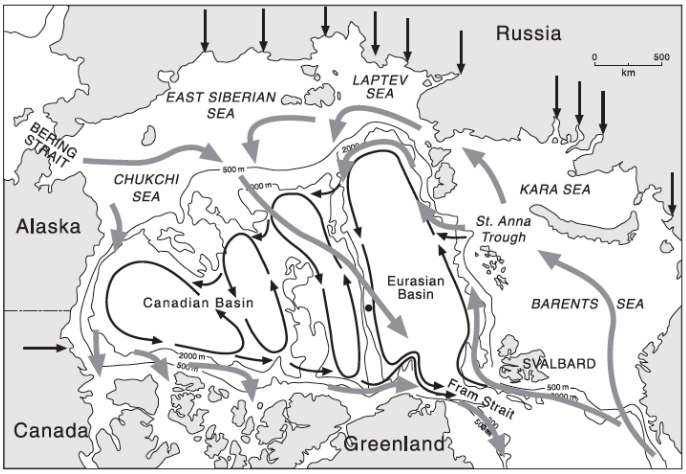

Figure 1.2. Schematic surface (grey arrows) and subsurface (black arrows) circulation of the Arctic Ocean. The strait arrows represent the mouths of major rivers (from Jones, 2001).

The Arctic Ocean is particularly sensitive with respect to ongoing climate changes. The reported ice extent-decrease (e.g., Stroeve et al., 2012), as a consequence of rising surface air temperatures (e.g., Rigor et al., 2000; Johanessen et al., 2004) causes a decrease in ice albedo and a stronger heat absorption of the ocean. This, in turn increases the ice-melt (Arctic amplification; Manabe & Stouffer, 1980; Serreze & Francis, 2006; Serreze et al., 2007). The Fram Strait plays a key role in these processes, which makes the region of special interest. It is the only deep passage between the Arctic Ocean and the rest of the world ocean, supplying the Arctic Ocean – apart from the Barents Sea – with North Atlantic-derived waters. It is the most important pathway for the oceanic transport of sensible heat to and the main exit for Polar Water and sea ice from the Central Arctic (Rudels et al., 1999; Jones, 2001). Hence, along its approximately 500 km cross section highly variable oceanographic conditions can be found. Two major surface current systems dominate the area: the West Spitsbergen Current (WSC) in the eastern part that transports warm and saline water masses northward and the southward flowing cold and fresh East Greenland Current (EGC) in the west (Fig. 1.3). The upper 500 m of the WSC are dominated by Atlantic Water (AW). This submerges north of 78°N beneath a cool and fresh upper mixed layer of Arctic origin and continues as a subsurface current into the Arctic Ocean (Johannessen, 1986).

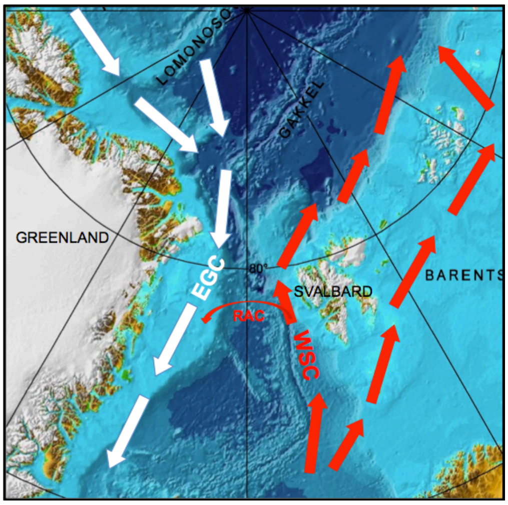

Figure 1.3. Schematic surface circulation of the Fram Strait. Red arrows indicate the West Spitzbergen Current (WSC) and the Return Atlantic Current (RAC), white arrows indicate the East Greenland Current (EGC). Map source: the International Bathymetric Chart of the Southern Ocean (Jakobsson et al., 2012).

In the northern part of the Fram Stait also a strong recirculation of AW takes place. Here Atlantic Water is carried by meso-scale eddies westward in the strait whereby it submerges below the EGC and forms the subsurface Return Atlantic Current (RAC) (Johanessen et al., 1987) that contributes to the deepwater formation in the Nordic Seas (Gascard et al., 1988; Marnela et al., 2008). Above it, in the upper 200 m the EGC carries cold, low-saline Arctic outflow waters along the Greenland continental slope into the Nordic Seas.The sea-ice extent in the strait is controlled by the interaction between the two strongly contrasting surface/subsurface currents and is highly variable (Vinje, 1977, 2001): because of the melting effects of the warm and saline AW, usually only the shelf of East Greenland and the northern Fram Strait stay covered by ice in the summer, but in the case of extremely cold winter/spring periods the ice can even intrude into the eastern part of the strait (Dickson et al., 2000).

1.4 Objectives and outline of the study

Despite of its importance in understanding the recent global warming, the Arctic realm still represents a challenging field of study for palaeoceanography. Knowledge about past climate variability is necessary for evaluating recent changes and predicting future climate trends. For this reason, several palaeoceanographic studies have been conducted in the area (for recent reviews see Stein, 2008 and Polyak et al., 2009). However, the interpretation of proxy data from high-latitude oceans raises some difficulties. The Arctic region represents an extreme environment to planktic life with very low sea surface temperatures (< −1.5°C) and partly permanent ice cover, which leads to limitations in the applicability of certain palaeoceanographic methods. Low diversity of planktic foraminifera species in the area results in too high temperature estimates with modern analog techniques (Pflaumann et al., 1996; Husum & Hald, 2013), while cold surface waters seem to cause inconsistency in Mg/Ca thermometry (Nürnberg, 1995; Meland et al., 2006; Kozdon et al., 2009; Jonkers et al., 2013). Moreover, local environmental parameters like ice coverage and surface hydrography might influence abundances, depth habitat and stable isotope incorporation of planktic foraminifera (e.g., Carstens et al., 1997; Volkmann, 2000; Bauch et al., 1997). Thus, in order to correctly interpret the palaeo-record, it is inevitable to examine the modern assemblage, its biochemistry and the relations to environmental parameters. However, compared to tropical regions, rather few studies on recent planktic foraminifera concentrated on the Fram Strait and generally on the (sub)Arctic. Modern distributions and/or stable isotopic compositions of foraminiferal shells were analysed in Freemansundet, Svalbard (Hansen & Knundsen, 1995), in the

Nansen Basin (Carstens & Wefer, 1992; Bauch et al., 1997), in the North-east Water Polynya (Kohfeld et al., 1996) and in the Nordic Seas (Simstich et al., 2003). Samples collected close to our study site in the Fram Strait were investigated by Carstens et al. (1997), Volkmann (2000) and Volkmann & Mensch (2001). These studies revealed that the modern planktic foraminiferal assemblage in the Fram Strait is dominated by the polar, non-spinose species *Neogloboquadrina pachyderma* (sinistral) (Ehrenberg, 1861) and the subpolar, spinose, symbiont-bearing species *Turborotalita quinqueloba* (Natland, 1938). *N. pachyderma* (sin.) is often used in palaeoceanographic reconstructions in the high latitudes (e.g., Devereux et al., 1970; Kellogg, 1980; Spielhagen et al., 2004) because of its almost continuous presence during glacial and interglacial times. *T. quinqueloba* is a reliable proxy for the warmer Atlantic Waters entering the Arctic Ocean (Carstens & Wefer, 1992; Bauch, 1994) and it is dominating the Holocene sediments in the Nordic Seas (Bauch, 1993). Nevertheless, the ecology of these two species is not well known. In the Fram Strait the ratio between the abundances of *N. pachyderma* (sin.) and *T. quinqueloba* seems to vary with the hydrographical regime (Carstens et al., 1997; Volkmann, 2000). Findings about the depth distribution of both species are contradicting: Carstens et al. (1997) concluded that *N. pachyderma* (sin.) and *T. quinqueloba* follow the path of the Atlantic Water submerging below Polar waters, while Volkmann (2000) assumed that *N. pachyderma* (sin.) prefers shallower water depths under permanent ice coverage. However, both studies were conducted in late summer and did not take in account that – just like in the North Atlantic (Schiebel & Hemleben, 2000) – seasonality might influence abundances and depth habitat as well. Moreover, the degree of disequilibrium in the stable isotope incorporation of these microorganisms seems to vary regionally (e.g., in $\delta^{18}O$: Bauch et al., 1997; Volkmann & Mensch, 2001; Simstich et al., 2003; Jonkers et al., 2010), while the Mg/Ca ratio of tests collected in high-latitude oceans do not always have the expected exponential correlation with water temperatures. Therefore, to improve the reliability of palaeoceanographic reconstructions, the factors influencing the biochemical composition of the tests still need further attention. For this purpose the upper 500 m of the water column and the sediment surface were sampled for planktic foraminifera with a multinet and a multicorer, along a transect across the Fram Strait (78°50'N, 5°W-8°E). Additionally, at the same locations temperature and salinity of the water column were measured by a Conductivity-Temperature-Depth (CTD) profiler and water samples were taken with a rosette sampler. The sampled transect comprises strongly contrasting water masses (see above), which makes it possible to investigate the effects of distinctly different water temperatures/salinities on the species assemblage and on the biochemical composition of the tests.

The major objectives of this study are:

(1) To describe the horizontal and vertical distribution of the planktic foraminiferal fauna in the Fram Strait and to determine the factor(s) influencing the depth habitat of the two dominating species in this area.

(2) To compare the species assemblage found on the sediment surface to that sampled from the water column.

(3) To compare the stable isotope composition of shells collected from the water column and from the sediment surface to that of the ambient water and to investigate possible reasons for discrepancies.

(4) To test the validity of Mg/Ca thermometry in these high latitudes on *N. pachyderma* (sin.) and *T. quinqueloba* collected in the water column and from the sediment surface.

This thesis comprises three manuscripts (chapter 2-4) that are published (chapter 2 and 3) and to be submitted soon (chapter 4). Plankton, sediment surface and water samples analysed in the study were collected during RV Polarstern expedition XXVI/1 in the Fram Strait in June/July 2011 and were identical in all three manuscripts. Chapter 2 (Pados & Spielhagen, 2014) focuses on the ecology and depth distribution of living planktic foraminifera in the study area. Chapter 3 (Pados et al., 2014) discusses the correlation between stable isotope composition of living planktic foraminifera and *in situ* measured water column properties. In chapter 4 we test the temperature dependence of Mg/Ca in planktic foraminifera sampled from the water column and from the sediment surface, and describe the Mg/Ca distribution in the water column in the Fram Strait (Pados et al., to be submitted).

Declaration of my contribution to the manuscripts:

Pados, T. & Spielhagen, R.F. (2014). Species distribution and depth habitat of recent planktic foraminifera in the Fram Strait (Arctic Ocean). *Polar Research, 33*, 22483, http://dx.doi.org/10.34027/polar.v33.22483.

Statement: I planned the study together with Robert F. Spielhagen. Sampling and sample analysis were done by myself. I wrote the manuscript, Robert F. Spielhagen helped improving and revising the text.

Pados, T., Spielhagen R. F., Bauch, D., Meyer, H., & Segl, M. (2014). Oxygen and carbon isotope composition of modern planktic foraminifera and near-surface waters in the Fram Strait (Arctic Ocean) – A case-study. *Biogeosciences Discussions, 11*, 1-38.

Statement: I designed the study together with Robert F. Spielhagen. The collection of samples was done by myself. Hanno Meyer and Monika Segl performed the stable isotope analysis of the water samples. I analysed the data set and wrote the first draft of the manuscript. Robert F. Spielhagen and Dorothea Bauch helped to improve and revise the manuscript.

Pados, T., Regenberg, M., Spielhagen, R.F., & Garbe-Schönberg, D.. Mg/Ca in water samples and in planktic foraminifera from plankton tows and sediment surface in the Fram Strait (to be submitted).

Statement: I planned the study and collected the samples by myself. I performed the sample analysis together with Marcus Regenberg and Dieter Garbe-Schönberg. I interpreted the data set and wrote the manuscript. Robert F. Spielhagen and Marcus Regenberg helped to improve and revise the manuscript

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2. Species distribution and depth habitat of recent planktic foraminifera in the Fram Strait (Arctic Ocean)

From {Pados, T. & Spielhagen, R. F. (2014). Species distribution and depth habitat of recent planktic foraminifera in the Fram Strait (Arctic Ocean). *Polar Research, 33*, 22483, http://dx.doi.org/10.3402/polar.v33.22483.}. Reprinted under Creative Commons License.

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RESEARCH/REVIEW ARTICLE

Species distribution and depth habitat of recent planktic foraminifera in Fram Strait, Arctic Ocean

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Keywords
Planktic foraminifera; Fram Strait; Arctic Ocean; depth habitat; *N. pachyderma* (sin.); *T. quinqueloba*.

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Abstract

To describe the horizontal and vertical distribution of recent planktic foraminifera in Fram Strait (Arctic), plankton samples were collected in the early summer of 2011 using a MultiNet sampler (> 63 µm) at 10 stations along a west–east transect at 78°50′N. Five depth intervals were sampled from the sea surface down to 500 m. Additionally, sediment surface samples from the same locations were analysed. The ratio between absolute abundances of planktic foraminifera in the open ocean, at the ice margin and in the ice-covered ocean was found to be approximately 2:4:1. The assemblage was dominated by the polar *Neogloboquadrina pachyderma* (sin.) and the subpolar *Turborotalita quinqueloba*, which accounted for 76 and 15% of all tests in the warm, saline Atlantic waters and 90 and 5% in the cold and fresh Polar waters, respectively. Both species had maximum absolute abundances between 0 and 100 m water depth, however, they apparently lived shallower under the ice cover than under ice-free conditions. This indicates that the depth habitat of planktic foraminifera in the study area is predominantly controlled by food availability and not by temperature. The distribution pattern obtained by plankton tows was clearly reflected on the sediment surface and we conclude that the assemblage on the sediment surface can be used as an indicator for modern planktic foraminiferal fauna.

To access the supplementary material for this article, please see Supplementary files under Article Tools online.

Planktic foraminifera are protozoa that inhabit the upper part of the water column in the world oceans. As their species distribution in high latitude oceans reflects properties of their habitat (i.e., water temperatures and the position of the summer sea-ice margin), fossil assemblages in sediments are often used to reconstruct palaeoenvironments. However, for a correct interpretation of fossil data, it is important to improve our understanding of the correlation between environmental variability in the ocean and the related distribution of living foraminifera. It is particularly essential to reveal the actual depth habitat of recent specimens, as the water properties reconstructed from their fossil calcite shells must be attributed to correct water depths.

Many studies on recent planktic foraminiferal distributions were conducted in tropical/subtropical areas (e.g., Bé & Hutson 1977; Andrijanic 1988; Naomi et al. 1990; Kuroyanagi & Kawahata 2004), while rather few studies concentrated on (sub-)Arctic regions, mainly on Fram Strait. Hansen & Knudsen (1995) published data from Freemansundet, Svalbard, Carstens & Wefer (1992) presented results on recent foraminifera in the Nansen Basin, and Kohfeld et al. (1996) analysed samples from the Northeast Water Polynya. Close to our work area, studies have been conducted by Carstens et al. (1997) and Volkmann (2000), who showed that planktic foraminiferal species composition in Fram Strait is dominated by polar *Neogloboquadrina pachyderma* (sinistral) (Ehrenberg 1861)

Citation: *Polar Research* 2014, *33*, 22483, http://dx.doi.org/10.3402/polar.v33.22483

and subpolar *Turborotalita quinqueloba* (Natland 1938). The abundance ratio of these two species varies with the hydrographical regime but they both occur in the upper 500 m of the water column, with highest abundances at the sea-ice margin. In this paper, we present new data on the vertical and horizontal distribution of planktic foraminifera species, collected during the *Polarstern* ARK-XXVI/1 cruise, in 2011, at 10 stations along a transect at 78°50′N across Fram Strait (Fig. 1). The samples were taken using a MultiNet sampler with nets of 63 µm mesh size at five depth intervals. Additionally, the sediment surface at the same locations was sampled. In this area, the production maximum of planktic foraminifera usually occurs in mid-summer (Kohfeld et al. 1996; Jonkers et al. 2010). Foraminifer assemblages in core-top and down core samples from Fram Strait are therefore thought to represent modern and past summer conditions, respectively. In order to examine whether processes during deposition (e.g., lateral transport, selective dissolution) significantly change the distribution of planktic foraminiferal tests in sediments, our MultiNet sampling results are also compared to the planktic foraminifer assemblages found on the sea floor.

Oceanography

Fram Strait is characterized by high oceanographic variability. Two major surface current systems dominate the area: the West Spitsbergen Current (WSC) in the eastern part that transports water masses northward and the southward flowing East Greenland Current (EGC) in the west (Johannessen 1986). The upper 500 m of the WSC is dominated by Atlantic Water with a thin (<5 m) mixed layer on top in the summer, resulting from ice melting. In winter, the Polar Surface Water takes over its place and forms a thicker wedge to the west (Rudels et al. 1999). The Atlantic Layer is characterized by salinities above 34.5 and temperatures higher than 0.5°C. It ranges between 200 and 500 m. The Polar Mixed Layer (PML) is colder, with a low salinity (<34) and about 50 m thickness. The two layers are isolated by a cold halocline (Jones 2001). The EGC carries cold, low-saline Arctic outflow waters in the upper 200 m. As a result of ice melt, surface waters at the East Greenland continental slope are characterized by extremely low salinities in summer (below 32). The PML here is located above a well-developed halocline (Rudels et al. 2000). Underneath the halocline, warmer and saline waters of Atlantic origin can be found. The sea-ice extent in the strait is controlled by the interaction between Polar and Atlantic waters and is highly variable (Vinje 1977, 2001): usually only the shelf of East Greenland and the northern Fram Strait stay covered by ice in the summer, but in the case of extremely cold winter/spring periods the ice can even intrude into the eastern part of the strait (Dickson et al. 2000).

Oceanographic measurements during ARK-XXVI/1 cruise (Beszczynska-Möller & Wisotzki 2012) provided

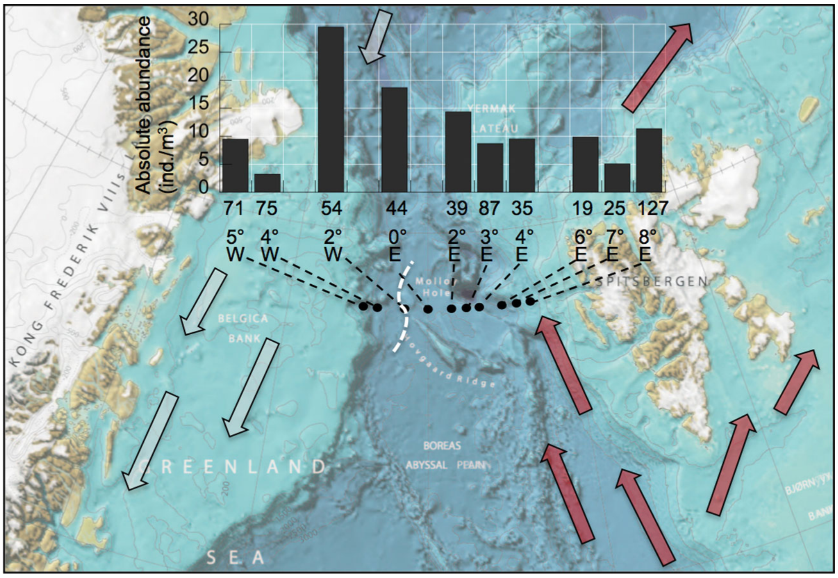

Fig. 1 Absolute abundances of planktic foraminifera at 10 stations along a transect at 78°50′N across Fram Strait. Arrows—red for the West Spitsbergen Current and white for the East Greenland Current—indicate the ocean circulation in the study area. The white dashed line shows the position of the sea-ice margin in the transect during the sampling period (shipboard observation and satellite data from the US National Oceanic and Atmospheric Administration). Map source: the International Bathymetric Chart of the Arctic Ocean (Jakobsson et al. 2012).

Citation: *Polar Research* 2014, *33*, 22483, http://dx.doi.org/10.3402/polar.v33.22483

a detailed profile across Fram Strait at the time of sampling (Fig. 2). The Atlantic Layer in the east was characterized by salinities near 35 and temperatures between 7 and 2°C, while the Polar Layer in the western part of the strait yielded salinities around 33 and a mean temperature of −1.5°C. Here, on the surface Beszczynska-Möller & Wisotzki (2012) found the above-mentioned low-salinity layer with lowest values of 30.6. Underneath the Arctic-derived and mainly ice-covered water mass, submerged warm and salty Atlantic waters can be found. The sea-ice margin in the transect was located at ca. 2°W during the sampling period.

Material and methods

Samples used in this study were obtained during the ARK-XXVI/1 expedition with the research vessel *Polarstern* in late June/early July 2011 in Fram Strait. Plankton samples were collected by a MultiNet sampler (net opening 0.5 m^2; Hydro-Bios, Kiel, Germany) at 10 stations along a transect at 78°50′N across Fram Strait (Fig. 1, Table 1). The nets of 63 µm mesh size were towed vertically on regular depth intervals (500–300 m, 300–200 m, 200–100 m, 100–50 m, 50–0 m) with a maximum winch speed of 0.3 m/s. During the sampling, the volume of filtered water was measured with a flow metre attached to the MultiNet frame. The flow metre was discovered to be defective and could not be repaired during the cruise. For some MultiNet casts and depth intervals, the instrument gave no or obviously unreliable results. We have therefore decided to omit the flow metre data for this study. The filtered volume was calculated from the mouth opening of the MultiNet and the vertical length of the towed interval for each sample. The plankton samples were sieved with a sieve of 500 µm mesh size, fixed with ethanol and stored at 4°C on board. Before picking the foraminifera by a pipette, all samples were rinsed with distilled water, treated with hydrogen peroxide to remove organic material and preserved in seawater. The picked specimens were rinsed with distilled water, dried on air and ashed in a low temperature vacuum asher for 90 min.

Sediment surface samples were obtained from multicorer deployments at the same stations and preserved in ethanol. Samples were freeze-dried, wet-sieved with distilled water through a 63 µm mesh, dried at 40°C and split into several fractions.

All foraminifera from a split of the 100–250 µm fraction of the plankton samples (half or quarter) and from a split of the 100–250 µm fraction of the sediment surface samples (containing >300 individuals) were identified using the taxonomy of Hemleben et al. (1989) and counted under dry conditions. Temperature and salinity of the water column were measured by a

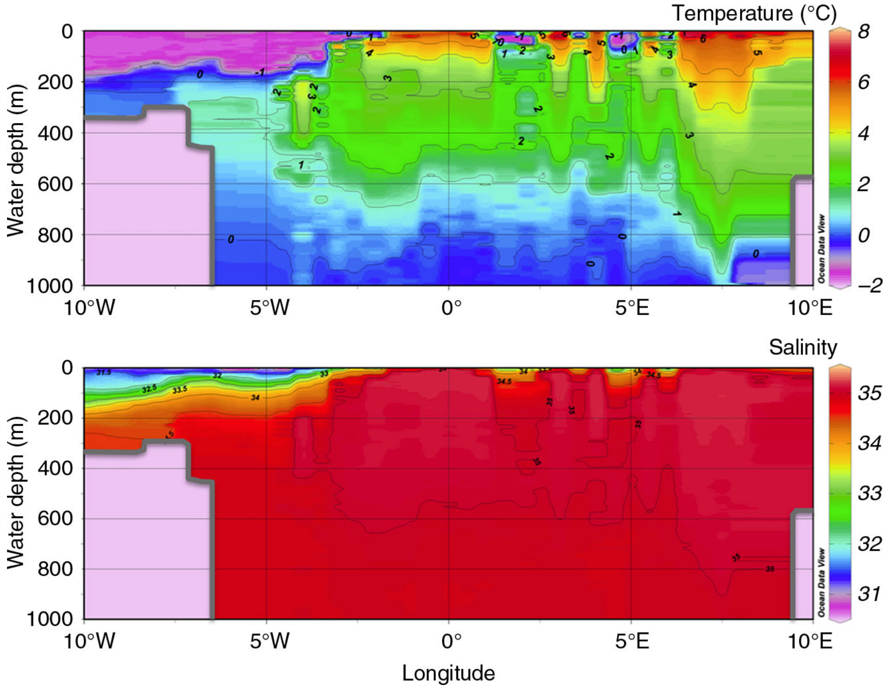

Fig. 2 Temperature and salinity of the water column in the upper 1000 m along a transect at 78°50′N across Fram Strait. Data obtained by conductivity–temperature–density measurements during the ARK XXVI/1 expedition (Beszczynska-Möller & Wisotzki 2012).

Table 1 List of the stations sampled during the ARKXXVI/1 cruise with the *Polarstern* in June/July 2011.

Station no.	Latitude	Longitude	Water depth	Date	Ice cover
PS78-19	78°49.84′N	6°0.69′E	2464 m	25.06.11	No
PS78-25	78°49.962′N	7°0.077′E	1465 m	26.06.11	No
PS78-35	78°49.772′N	3°58.380′E	2335 m	28.06.11	No
PS78-39	78°50.09′N	1°54.56′E	2554 m	28.06.11	No
PS78-44	78°49.972′N	0°4.630′E	2636 m	29.06.11	No
PS78-54	78°50.02′N	2°0.21′W	2714 m	01.07.11	Ice margin
PS78-71	78°49.66′N	5°20.99′W	684 m	04.07.11	Ice covered
PS78-75	78°49.74′N	3°55.44′W	1978 m	04.07.11	Ice covered
PS78-87	78°50.44′N	3°0.19′E	2454 m	06.07.11	No
PS78-127	78°49.84′N	8°1.33′E	1019 m	10.07.11	No

conductivity–temperature–depth (CTD) profiler (Beszczynska-Möller & Wisotzki 2012) immediately before the plankton tows. Data of ice coverage were obtained by shipboard observations.

Results

Planktic foraminifera in the plankton samples

The absolute abundances of planktic foraminifera of the 100–250 µm size class in the upper 500 m of the water column show a general increase from ca. 10 individuals per cubic metre (ind./m^3) at the eastern stations to 29 ind./m^3 at the sea-ice margin (station 54). Further westward we can see a drastic decrease (3 and 9 ind./m^3, stations 75 and 71, respectively; Fig. 1).

The species assemblage is dominated by polar *Neogloboquadrina pachyderma* (sin.) and subpolar *Turborotalita quinqueloba*, with proportions of around 76 and 15% in the east and 90 and 5% in the west, respectively. Other planktic foraminifera species found in the samples (*N. pachyderma* [dex.] and *Globigerina bulloides*) contribute less than 9% to the faunal composition. The relative abundance of *N. pachyderma* (sin.) varies between 73 and 92% with an increasing trend towards the west, while *T. quinqueloba* shows a reverse tendency with the highest value of 23% in the eastern part of the strait and the lowest (5%) at the Greenland slope. Figure 3 shows the species composition at the stations in preselected depth intervals. *N. pachyderma* (sin.) contributes less than 70% to the total planktic foraminiferal fauna at two stations (at 8°E and 6°E), between 0 and 50 m water depth. The highest proportion of this species was found at station 75 (4°W).

The two dominant species, *N. pachyderma* (sin.) and *T. quinqueloba*, show similar trends in the depth distribution: both have maximum absolute abundances (ind./m^3) between 0 and 100 m water depths (Fig. 4). *N. pachyderma* (sin.) has highest abundances near the surface between 0 and 50 m at stations 35 and 87 in the middle of the strait (4°E and 3°E, respectively) and under the ice cover at station 75. Highest amounts of *T. quinqueloba* in the upper 50 m were found at stations 127, 35 and 44 (8°E, 4°E and 0°E, respectively); at the remaining stations, the depth of 50–100 m yielded maximum abundances of this species.

Looking at the relative abundances (percentage of total population of the given species at the given station per depth interval; Fig. 4), both species show a trend to increasing percentages from the surface to the interval 50–100 m, followed by a decrease to 300–500 m. Only at the easternmost station (127), the population maximum of *N. pachyderma* (sin.) descends below 200 m, while the highest relative abundance of *T. quinqueloba* at all stations lies above 200 m.

Sediment samples

The samples do not show any evidence of selective carbonate dissolution. Only few broken shells were found, and the two studied species do not show any significant difference in the degree of fragmentation.

In the absolute abundances of planktic foraminifera, we see a trend similar to that of the plankton samples: the station close to Svalbard (25) has a low value ($<$400 ind./g sediment), followed by a minor peak at 4°E (ca. 6500 ind./g). After a steady interval (ca. 3000 ind./g), a significant peak can be observed at the sea-ice margin (station 54, ca. 26 000 ind./g), followed by a drastic decrease to ca. 50 ind./g below the sea ice (Fig. 5).

Neogloboquadrina pachyderma (sin.) has highest proportions in the assemblage at the sea-ice margin (91%) and lowest (73%) in the east (7°E, station 25). Between these two stations, a steady increase can be detected. *T. quinqueloba* shows an exactly reverse trend with abundances between 8 and 24% (Fig. 6). Comparing the relative abundances of *N. pachyderma* (sin.) and *T. quinqueloba* on

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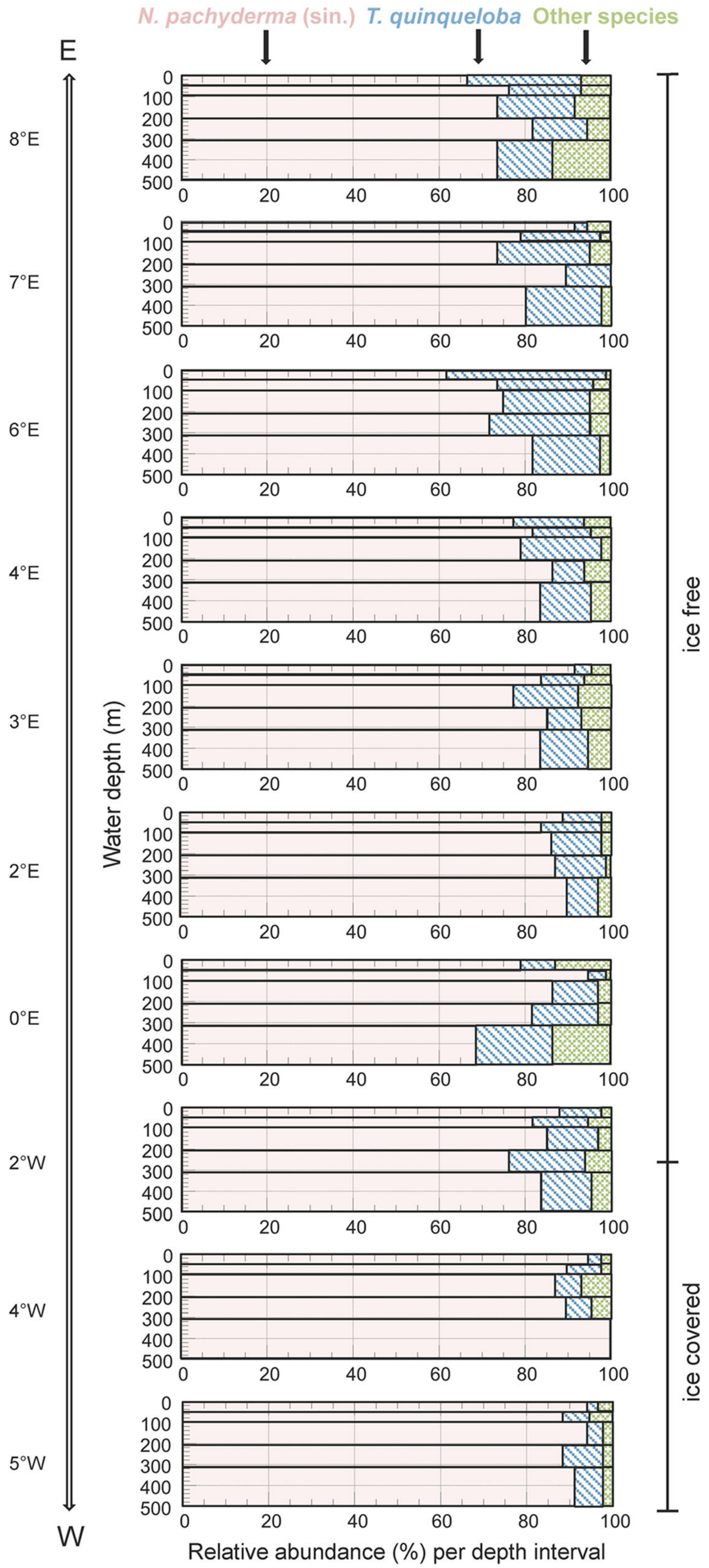

Fig. 3 Faunal composition of samples taken from the upper 500 m of the water column along a transect at 78°50′N in Fram Strait.

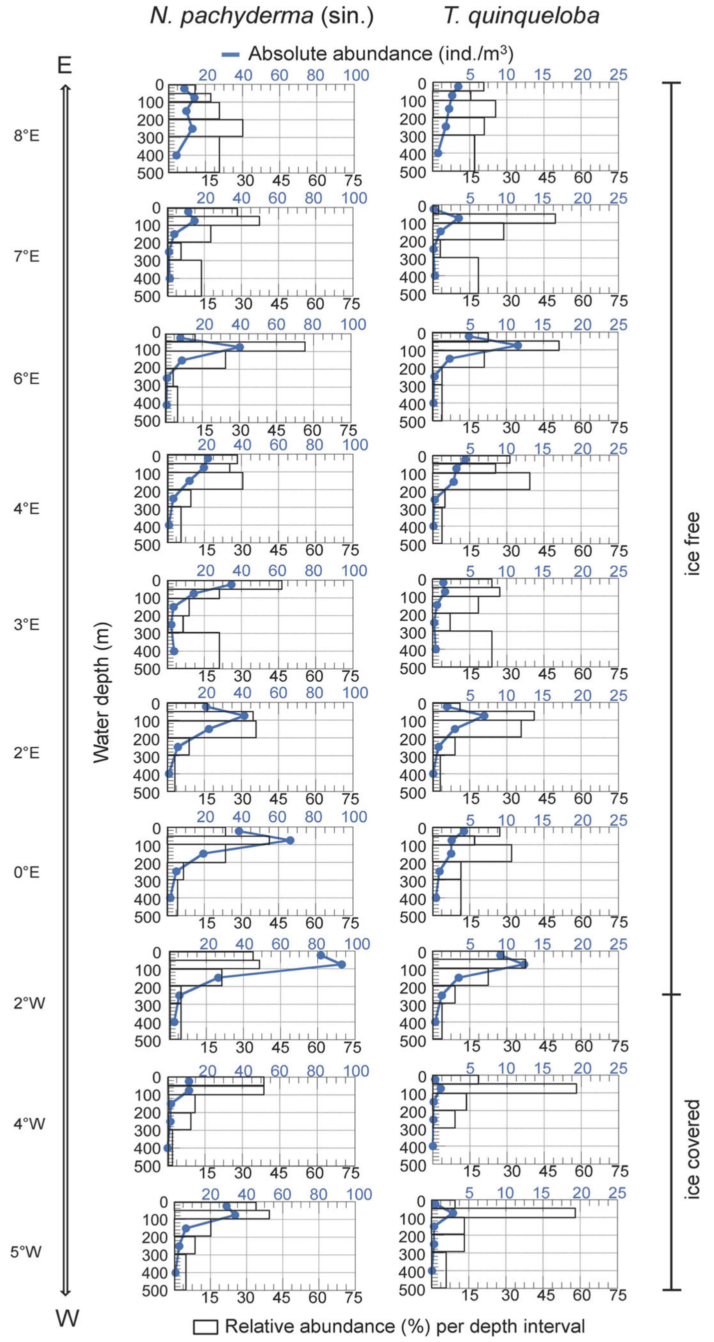

Fig. 4 Absolute (blue lines) and relative (black bars) abundances (percentage of total population of the given species at the given station per depth interval) of *Neogloboquadrina pachyderma* (sin.) and *Turborotalita quinqueloba* in the upper 500 m along a transect at 78°50′N in Fram Strait.

the sediment surface to those counted in the upper 100 m of the water column, the species composition in the depth interval of 50–100 m follows the trend of the core-top samples more precisely than the depth interval between 0 and 50 m mirrors the sediment surface (Fig. 6). All count data can be found in Supplementary Tables S1–S3.

Discussion

Species assemblage

The planktic foraminiferal species composition of the study area resembles the typical foraminiferal fauna previously reported from Arctic/sub-Arctic regions with

Citation: *Polar Research* 2014, *33*, 22483, http://dx.doi.org/10.3402/polar.v33.22483

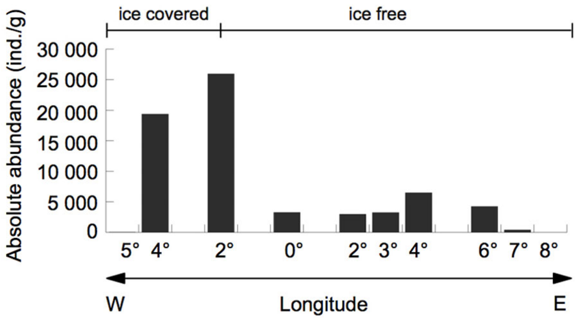

Fig. 5 Absolute abundances of planktic foraminifera on the sediment surface at 10 stations along a transect at 78°50′N across Fram Strait. The sample from station PS78-127 (8°E) did not hold enough individuals for counting.

the clear dominance of *Neogloboquadrina pachyderma* (sin.) and *Turborotalita quinqueloba*. A similar species distribution has been already shown for the Arctic Ocean (Bé 1960; Carstens & Wefer 1992) and for Fram Strait (Carstens et al. 1997; Volkmann 2000). Furthermore, despite the warming of Atlantic waters passing through the strait (Spielhagen et al. 2011), the polar species *N. pachyderma* (sin.) still clearly prevails in our samples, compared to the subpolar *T. quinqueloba*. Interestingly, at the sampled stations, *T. quinqueloba* showed relatively low values compared to the results of the above-mentioned studies (Fig. 3). Carstens et al. (1997) observed a range from 15 to 64% for this species, while the proportion of *T. quinqueloba* in certain samples collected by Volkmann (2000) in warm, Atlantic Water-dominated regimes reached even 93%. In our plankton samples, the relative abundance of *T. quinqueloba* varied between 5 and 23% and the sediment surface samples showed a very similar trend. These discrepancies point to a variable foraminiferal production found due to different sampling periods. Carstens et al. (1997) and Volkmann (2000) took their samples in August, while our samples were collected in late June/early July and therefore probably reflect an early summer plankton bloom. In the eastern North Atlantic Ocean, Schiebel & Hemleben (2000) found distinctly different planktic foraminiferal faunas in June and in August. In contrast to early summer, the late summer assemblage was dominated by species preferring warmer waters than other species present, just like the subpolar *T. quinqueloba* in the planktic foraminiferal fauna of Fram Strait. It has been shown for the western North Atlantic that, while the production of *N. pachyderma* (sin.) has two peaks during the season and one already in early summer, the production of *T. quinqueloba* reaches its maximum only in early autumn, after the second bloom of *N. pachyderma* (sin.) (Jonkers et al. 2010). Although sediment trap studies in the Northeast Water Polynya revealed a unimodal pattern of planktic foraminiferal production (Kohfeld et al. 1996), our data indicate that a temporal offset in the production maxima of the two species exists also at these northern latitudes. This could lead to the relatively low numbers of *T. quinqueloba* in our samples, collected in June/July compared to the studies conducted in August.

Horizontal distribution

The variable species composition along the transect reflects the complex hydrography of Fram Strait. In the east, the WSC carries warm, saline Atlantic Water northwards. In this area, the subpolar *T. quinqueloba* had higher abundances and the polar *N. pachyderma* (sin.) had lower abundances than in the western part where the upper 200 m of the EGC consist of cold, low-saline Arctic outflow waters. Here, *N. pachyderma* (sin.) had a proportion of around 90%. This indicates that the horizontal distribution of the two species is primarily controlled by the distribution of different water masses.

The ratio between absolute abundances of planktic foraminifera in the open ocean, at the ice margin and in the ice-covered ocean was found to be ca. 2:4:1 (Fig. 1). Our data show a similar trend to that in the study of Carstens et al. (1997). The ice margin offers increased food supply for the foraminifera with higher primary production caused by diatoms that are the major food

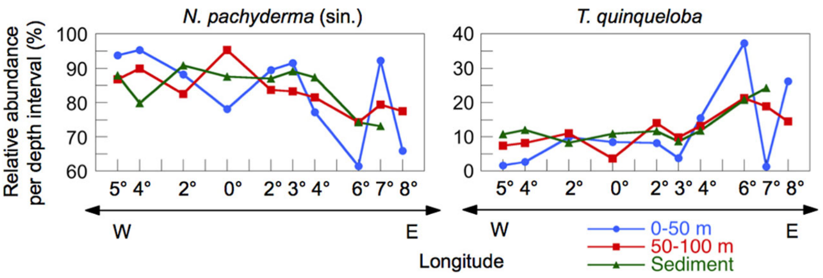

Fig. 6 Comparison of relative abundances of *Neogloboquadrina pachyderma* (sin.) and *Turborotalita quinqueloba* in the depth intervals 0–50 m and 50–100 m as well as on the sediment surface at 10 stations along a transect at 78°50′N in Fram Strait.

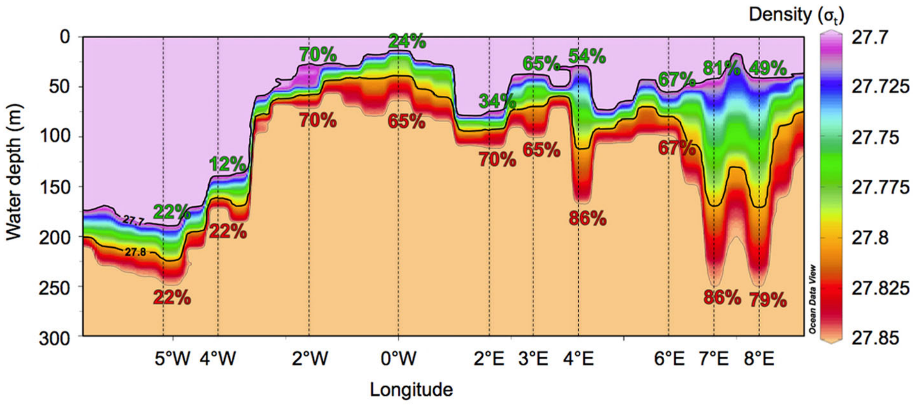

Fig. 7 The position of the isopycnal layer of 27.7–27.85 along the transect at 78°50′N across Fram Strait and relative abundance of *Neogloboquadrina pachyderma* (sin.) found in the depth intervals including the density range of 27.7–27.8 (green numbers at the top) and the range of 27.7–27.85 (red numbers at the bottom) at the 10 stations. Water density data obtained by conductivity–temperature–density measurements during the ARK XXVI/1 expedition (Beszczynska-Möller & Wisotzki 2012).

source for *N. pachyderma* (sin.) (Hemleben et al. 1989). Diatoms typically do not develop well under permanent ice cover due to the light limitation, but they prefer seasonally stratified water conditions that occur in the marginal ice zone (Smith et al. 1987; Williams 1993). Ice melting increases stratification and consequently the stability of the water column, which triggers phytoplankton blooms (Alexander 1980), and the enhanced phytoplankton biomass along the ice margin may have led to a major increase in absolute abundances of planktic foraminifera.

Vertical distribution

Both *N. pachyderma* (sin.) and *T. quinqueloba* had maximum absolute abundances between 0 and 100 m water depth. However, the relative abundances per depth interval (Fig. 4) showed that in the western Fram Strait, at the ice-covered stations more specimens of *N. pachyderma* (sin.) lived close to the surface (above 100 m), while in the eastern part of the strait the majority of the population was found deeper, between 50 and 200 m. Simstich et al. (2003) analysed plankton and several core-top samples from the Nordic seas, taken mainly south of our transect (80°N–62°N). They reported on a shallow apparent calcification depth in *T. quinqueloba* over most of the studied profile. If we assume that foraminifera calcify in the water depth where they are most abundant, these results are consistent with our findings. In their study, going from west to the east, the habitat of *N. pachyderma* (sin.) seems to sink deeper in the Atlantic Water off Norway, similar to what we observe from our samples taken from the warm and saline WSC. These results contrast with the findings of Carstens et al. (1997), who concluded that *N. pachyderma* (sin.) and *T. quinqueloba* follow the path of the Atlantic Water submerging below Polar Water and therefore live deeper in the western part of the strait. During our sampling period, in the western Fram Strait, the surface Polar water masses extended down to 200 m. As none of the studies on planktic foraminifera conducted in the same area (including the study of Carstens et al. 1997) revealed an average depth habitat below 200 m, the scenario described by Carstens et al. (1997) seems rather unlikely here. Furthermore, previous investigations revealed that food availability strongly affects the distribution of planktic foraminifera (Schiebel & Hemleben 2005). Thus, the depth habitat of these protozoa might be controlled by the position of the chlorophyll *a* maximum (Fairbanks & Wiebe 1980; Kohfeld et al. 1996) rather than by the water mass distribution. From the summer months in the Northeast Water Polynya, with oceanographic conditions similar to the western Fram Strait, Kohfeld et al. (1996) reported maximum chlorophyll *a* concentrations and consequently highest abundances of *N. pachyderma* (sin.) between 20 and 80 m water depth, which accords with the results presented here. Hirche et al. (1991) also found, for the same period of the year in Fram Strait, chlorophyll *a* maxima in the open ocean at greater depth than under the ice coverage. Assuming that the depth distribution of *N. pachyderma* (sin.) is indeed predominantly controlled by food availability in this area, the deeper depth habitat reported by Carstens et al. (1997) could be explained by different sampling periods. It is possible that during the above-mentioned experiments in late summer, the ice coverage was not as intact as in June/July and consequently, due to the higher translucency, the deep chlorophyll maximum might have descended/extended to deeper layers. Planktic foraminifera respond to the redistribution of chlorophyll within several days (Schiebel et al. 2001) and may therefore have followed the descent of the chlorophyll maximum. Another possible explanation for the different findings might be the sampling in different ontogenetic stages. *N. pachyderma* (sin.) and *T. quinqueloba* are known to reproduce on a synodic lunar cycle (Bijma et al. 1990; Schiebel & Hemleben 2005) and for reproduction they simultaneously descend from their average depth habitat, possibly to the greatest depth they inhabit during their ontogenetic cycle (Berberich 1996; Schiebel & Hemleben 2005). Volkmann (2000) also assumed that *N. pachyderma* (sin.) prefers shallower water depths under permanent ice coverage. In her study, this species showed maximum abundances in the narrow temperature layer between −1.5 and −1.8°C, while *T. quinqueloba* did not show such preference to a particular temperature range. In our study, the maximum abundances of *N. pachyderma* (sin.) and *T. quinqueloba* were found between 0 and 100 m water depth at every station and therefore in a very wide temperature range (from −1.7 to 7°C; Fig. 2). We therefore assume that the depth habitat of planktic foraminifera in Fram Strait during June/July is more influenced by the distribution of the food source than by temperature.

Kozdon et al. (2009) proposed that the average depth habitat of *N. pachyderma* (sin.) in the Nordic seas is controlled by water density and that this species is bound to a layer between 27.7 and 27.8 (potential density, σ_t). Along our studied transect during the sampling period, this isopycnal layer was found mainly between 0 and 100 m water depth. There are two exceptions. The first are the stations at 8°E, 7°E and 4°E, where this water mass reaches deeper. The second is the western part of the strait (west of ca. 3°W), on the Greenland shelf, where the whole layer submerged below 150 m. Regarding the relative abundances of *N. pachyderma* (sin.), our data show that in the eastern and central Fram Strait (2°W–8°E) during the sampling period a broad range from 34 to 81% of the population lived in the depth intervals including the specific density range. Extending the range to 27.85, we found a better correlation with values between 65 and 86% (Fig. 7). In contrast, in the west, in the cold and fresh water masses of the EGC, the depth habitat of *N. pachyderma* (sin.) was found in significantly shallower water depths than the layer of $\sigma_t = 27.7–27.8$. This discrepancy of findings may be related to the nature of the samples and data used. While the water mass and foraminifer data of our study are from just a short time interval and may be representative of only one early summer plankton bloom, they are nonetheless consistent in being obtained at the same time and place (i.e., station and water depth). Kozdon et al. (2009) calculated the calcification depth of *N. pachyderma* (sin.) using a long-term oceanographic database and temperatures obtained from paired Mg/Ca and $\delta^{44/40}$Ca measurements of shells picked from sediment surface samples. Such samples integrate foraminifer assemblages over the entire plankton growth seasons in a certain number of years, depending on sedimentation rates and bioturbation intensity. Furthermore, variable ages of the surface samples analysed (modern to more than 1000 yr ago; see Simstich et al. 2003) may play a more significant role than anticipated by Kozdon et al. (2009), who claimed that the samples "represent modern oceanographic conditions." Recent work in the Norwegian Sea and Fram Strait (Sejrup et al. 2010; Spielhagen et al. 2011; Werner et al. 2011) has demonstrated that temperature variations of 2°C and more occurred in the Atlantic Water layer during the last one to two millennia. Particularly in areas with low sedimentation rates (i.e., in ice-covered regions), sediment surface samples may thus rather represent earlier and colder periods (e.g., the Little Ice Age) than just modern, warmer conditions of the comparatively short Industrial Period. For this reason, the isopycnals calculated by Kozdon et al. (2009) on the basis of data extracted from a hydrographic database may not always correspond to the average conditions during deposition of the related sediment surface samples. Our comparison of *in situ* sampled foraminifera and measurements avoid such potential bias and indicates a wider density range for the habitat of *N. pachyderma* (sin.).

Comparison with the sedimentary records

Comparison of plankton tow results with those from the sediment surface has important implications for palaeoceanographic reconstructions because in such studies core-top samples are usually assumed to represent the recent foraminiferal fauna and therefore modern conditions. In our study area, the planktic foraminiferal species composition in the overlying water column is fairly well reflected in the sediment surface samples: the polar *N. pachyderma* (sin.) clearly dominates all the samples, with relative abundances of the subpolar *T. quinqueloba* varying between 8 and 24%. Again, the differences between plankton tow and sediment data can most likely be ascribed to the early summer sampling of the water column. Lateral advection of foraminifera is not considered a major factor because mean transport distances in Fram Strait are only 25–50 km for *N. pachyderma* (sin.) and 50–100 km for *T. quinqueloba*, as determined for a sediment trap at 1125 m water depth (von Gyldenfeldt et al. 2000). In the east and the west of the study area (where differences between plankton tow and sediment data are largest), the flow direction of warm and cold water masses is largely normal to the sampling transect and transport effects of subpolar and polar species on the species assemblages on the sea floor can be expected to be minor. Furthermore, the increase in abundances in the water column at the sea-ice margin compared to the open ocean and under the ice cover is also seen on the sediment surface (Fig. 5). Comparison of the relative abundances of *N. pachyderma* (sin.) and *T. quinqueloba* in the depth intervals of 0–50 m and 50–100 m to those from the sediment surface (Fig. 6) suggests that the sediment surface samples relate best to the fauna living between 50 and 100 m water depth. This hypothesis is supported by the fact that in the water column at almost all stations both species were most abundant in this depth interval. The congruence of the fauna found in the water column and on the sediment surface indicates that in this area processes that can influence the deposition of shells (e.g., lateral transport, selective dissolution) do not change significantly the distribution during settling. Since Fram Strait is relatively narrow and interannual west–east variability in the position of the average summer sea-ice margin can be high, effects of fluctuations related to centennial-scale climate changes may be more subdued in the sediment surface samples than in the Nordic seas. As a consequence, the large-scale oceanic regimes are more obvious from the faunal compositions. This suggests that planktic foraminifera in Fram Strait sediments provide a fairly reliable proxy for palaeo-water mass characterization.

Conclusions

In Fram Strait, the distribution pattern of the two dominating species, *Neogloboquadrina pachyderma* (sin.) and *Turborotalita quinqueloba* shows a distinct relation to the complex hydrography. Warm and saline Atlantic waters yield higher abundances of the subpolar *T. quinqueloba* than the cold and fresh waters of the EGC, while the polar *N. pachyderma* (sin.) shows an exactly reverse tendency.

The ratio between absolute abundances of planktic foraminifera in the open ocean, at the ice margin and in the ice-covered ocean is ca. 2:4:1. High abundances in the marginal ice zone are related to the high primary production triggered by the strongly stratified water column at the ice margin.

The depth habitats of *N. pachyderma* (sin.) and *T. quinqueloba* in the area seem predominantly controlled by the availability of food. Both species dwell shallower under the ice cover than under ice-free conditions, possibly following the position of the deep chlorophyll maximum. The effect of other water column properties, like temperature, on the depth habitat seems to be masked. We could not verify the hypothesis of Kozdon et al. (2009), which associates the vertical distribution of *N. pachyderma* (sin.) to a discrete isopycnal band. This discrepancy might be due to various factors, for example, differences in applied methods and palaeoceanographic variability in the study area in the recent past.

Sediment surface samples, often used for palaeoceanographic reconstructions, fairly well reflect the composition of the foraminiferal fauna living in the overlying water column. In Fram Strait, the species composition and relative abundance in the sediments correlate best with that of the fauna living between 50 and 100 m water depth. In this area—and probably beyond—planktic foraminifera may therefore be used as proxies for subsurface water conditions.

Acknowledgements

The authors wish to thank the captain and crew of the RV *Polarstern* and the Alfred Wegener Institute, Helmholtz Centre for Polar and Marine Research in Bremerhaven for support during ARK XXVI/1 cruise. We are indebted to Dr Angelina Kraft for assisting in the deployments of the MultiNet sampler and to Maciej Telesinski for collecting the sediment surface samples for this study. Dr Agnieszka Beszczynska-Möller and Andreas Wisotzki provided the CTD data for this study. Thanks are due to Dr Hartmut Schulz for kindly providing the low temperature vacuum asher at the University of Tübingen and for his hospitality and technical assistance during the process. Thanks also go to Dr Dorothea Bauch, who helped during the preparation phase for the cruise and for her valuable comments and suggestions throughout the study, and to Dr Daniele de Corte for reviewing the manuscript. We thank Thomas M. Cronin and one anonymous reviewer for their constructive feedback and insightful comments. This research was conducted with funding of the European Community's 7th Framework Programme FP7 2007/2013, Marie-Curie Actions, under Grant Agreement no. 238111–CASE ITN. RFS thanks the Academy of Sciences, Humanities and Literature Mainz for financial support through the Akademienprogramm.

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Citation: *Polar Research* 2014, *33*, 22483, http://dx.doi.org/10.3402/polar.v33.22483

Supplementary file for: Pados T. & Spielhagen R.F. 2014. Species distribution and depth habitat of recent planktic foraminifera in Fram Strait, Arctic Ocean. *Polar Research 33*. Correspondence:

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Supplementary Table S1. Number of counted foraminifera individuals in the 100-250 μm size fraction from the water column. (Table continues next page.)

Station	Depth (m)	Split counted	Count			
			N. pachyderma (sin.)	*T. quinqueloba*	*N. pachyderma* (dex.)	*G. bulloides*
PS78-127	0-50	1/2	108	43	11	2
	50-100	1/2	176	33	17	1
	100-200	1/2	241	56	30	0
	200-300	1/2	326	45	25	0
	300-500	1/2	222	37	44	0
PS78-25	0-50	1/4	71	1	5	0
	50-100	1/2	185	44	3	1
	100-200	1/2	95	26	7	0
	200-300	1/2	24	3	0	0
	300-500	1/2	72	16	2	0
PS78-19	0-50	1/2	102	62	1	1
	50-100	1/4	251	72	15	0
	100-200	1/2	226	59	15	0
	200-300	1/2	27	8	2	0
	300-500	1/2	44	9	1	0
PS78-35	0-50	1/2	274	55	24	2
	50-100	1/2	246	40	16	0
	100-200	1/2	300	71	7	1
	200-300	1/2	83	7	7	0
	300-500	1/2	52	5	3	0

PS78-87	0-50	1/4	216	9	11	0
	50-100	1/2	179	21	14	1
	100-200	1/2	85	15	10	0
	200-300	1/2	56	5	5	0
	300-500	1/2	186	20	14	0
PS78-39	0-50	1/2	263	24	7	0
	50-100	1/2	520	87	12	2
	100-200	1/2	561	75	14	1
	200-300	1/2	145	20	2	0
	300-500	1/2	54	4	2	0
PS78-44	0-50	1/2	484	53	83	0
	50-100	1/2	827	32	9	0
	100-200	1/2	489	62	14	1
	200-300	1/2	124	23	5	0
	300-500	1/2	88	23	18	0
PS78-54	0-50	1/4	510	57	10	1
	50-100	1/4	580	77	46	0
	100-200	1/4	331	44	13	0
	200-300	1/2	138	31	13	0
	300-500	1/2	138	18	8	0
PS78-75	0-50	1/2	143	4	3	0
	50-100	1/2	143	13	3	0
	100-200	1/2	46	3	4	0
	200-300	1/2	38	2	2	0
	300-500	1/2	4	0	0	0
PS78-71	0-50	1/2	361	6	18	0
	50-100	1/2	420	36	28	0
	100-200	1/2	160	8	3	0
	200-300	1/2	70	8	2	0
	300-500	1/2	39	3	1	0

Supplementary Table S2. Abundances of foraminifera (ind./m^3) calculated for total samples from the water column. (Table continues next page.)

Station	Depth (m)	Split counted	Abundance calculated for total sample (ind./m^3)			
			N. pachyderma (sin.)	*T. quinqueloba*	*N. pachyderma* (dex.)	*G. bulloides*
PS78-127	0-50	1/2	8.64	3.44	0.88	0.16
	50-100	1/2	14.08	2.64	1.36	0.08
	100-200	1/2	9.64	2.24	1.2	0
	200-300	1/2	13.04	1.8	1	0
	300-500	1/2	4.44	0.74	0.88	0
PS78-25	0-50	1/4	11.36	0.16	0.8	0
	50-100	1/2	14.8	3.52	0.24	0.08
	100-200	1/2	3.8	1.04	0.28	0
	200-300	1/2	0.96	0.12	0	0
	300-500	1/2	1.44	0.32	0.04	0
PS78-19	0-50	1/2	8.16	4.96	0.08	0.08
	50-100	1/4	40.16	11.52	2.4	0
	100-200	1/2	9.04	2.36	0.6	0
	200-300	1/2	1.08	0.32	0.08	0
	300-500	1/2	0.88	0.18	0.02	0
PS78-35	0-50	1/2	21.92	4.4	1.92	0.16
	50-100	1/2	19.68	3.2	1.28	0
	100-200	1/2	12	2.84	0.28	0.04
	200-300	1/2	3.32	0.28	0.28	0
	300-500	1/2	1.04	0.1	0.06	0
PS78-87	0-50	1/4	34.56	1.44	1.76	0
	50-100	1/2	14.32	1.68	1.12	0.08
	100-200	1/2	3.4	0.6	0.4	0
	200-300	1/2	2.24	0.2	0.2	0
	300-500	1/2	3.72	0.4	0.28	0
PS78-39	0-50	1/2	21.04	1.92	0.56	0

	50-100	1/2	41.6	6.96	0.96	0.16
	100-200	1/2	22.44	3	0.56	0.04
	200-300	1/2	5.8	0.8	0.08	0
	300-500	1/2	1.08	0.08	0.04	0
PS78-44	0-50	1/2	38.72	4.24	6.64	0
	50-100	1/2	66.16	2.56	0.72	0
	100-200	1/2	19.56	2.48	0.56	0.04
	200-300	1/2	4.96	0.92	0.2	0
	300-500	1/2	1.76	0.46	0.36	0
PS78-54	0-50	1/4	81.6	9.12	1.6	0.16
	50-100	1/4	92.8	12.32	7.36	0
	100-200	1/4	26.48	3.52	1.04	0
	200-300	1/2	5.52	1.24	0.52	0
	300-500	1/2	2.76	0.36	0.16	0
PS78-75	0-50	1/2	11.44	0.32	0.24	0
	50-100	1/2	11.44	1.04	0.24	0
	100-200	1/2	1.84	0.12	0.16	0
	200-300	1/2	1.52	0.08	0.08	0
	300-500	1/2	0.08	0	0	0
PS78-71	0-50	1/2	28.88	0.48	1.44	0
	50-100	1/2	33.6	2.88	2.24	0
	100-200	1/2	6.4	0.32	0.12	0
	200-300	1/2	2.8	0.32	0.08	0
	300-500	1/2	0.78	0.06	0.02	0

Supplementary Table S3. Number of counted foraminifera individuals in the 100-250 μm size fraction from sediment surface samples and their abundances (ind./g) calculated for total samples. The sample from station PS78-127 did not contain enough individuals for counting.

Station	Split counted	Count				Ind./g sediment
		N. pachyderma (sin.)	*T. quinqueloba*	*N. pachyderma* (dex.)	*G. bulloides*	
PS78-127	–	–	–	–	–	–
PS78-25	1/8	508	167	18	0	399.2
PS78-19	1/128	318	89	20	1	4248.0
PS78-35	1/256	534	72	5	0	6495.5
PS78-87	1/128	329	32	8	0	3240.7
PS78-39	1/128	322	43	5	0	2982.9
PS78-44	1/128	387	48	7	0	3280.2
PS78-54	1/1024	483	44	5	0	25948.2
PS78-75	1/1024	292	44	30	0	19367.7
PS78-71	1/2	316	40	5	0	48.3

3. Oxygen and carbon isotope composition of modern planktic foraminifera and near-surface waters in the Fram Strait (Arctic Ocean) – A case-study

From {Pados, T., Spielhagen R. F., Bauch, D., Meyer, H., & Segl, M. (2014). Oxygen and carbon isotope composition of modern planktic foraminifera and near-surface waters in the Fram Strait (Arctic Ocean) – A case-study. *Biogeosciences Discussions, 11*, 1-38.}. Reprinted under Creative Commons License.

Original paper available online at:
http://www.biogeosciences-discuss.net/11/8635/2014/bgd-11-8635-2014.html

Biogeosciences Discuss., 11, 8635–8672, 2014
www.biogeosciences-discuss.net/11/8635/2014/
doi:10.5194/bgd-11-8635-2014

Biogeosciences
Discussions

Oxygen and carbon isotope composition of modern planktic foraminifera and near-surface waters in the Fram Strait (Arctic Ocean) – a case-study

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Received: 14 May 2014 – Accepted: 19 May 2014 – Published: 11 June 2014

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Published by Copernicus Publications on behalf of the European Geosciences Union.

Abstract

The upper 500 m of the water column and the sediment surface along an E–W transect in the Fram Strait were sampled for recent planktic foraminifera. The $\delta^{18}O$ and $\delta^{13}C$ values of the tests are compared to the stable isotope composition of water samples taken from the same depths, and related to the characteristics of the water column. The polar species *Neogloboquadrina pachyderma* (sin.) clearly dominates the species assemblage in the Fram Strait in the early summer, while the subpolar *Turborotalita quinqueloba* accounts only for 5–23 %. In this area the average depth of calcification of *N. pachyderma* (sin.) lies between 70–150 m water depth, *T. quinqueloba* shows a similar range with 50–120 m water depth. The $\delta^{18}O$ values of *N. pachyderma* (sin.) show an average vital effect of about −1.5 ‰ compared to calculated equilibrium calcite values. Except for the upper ~ 75 m, the vertical profiles of $\delta^{13}C$ of the net-sampled shells are nearly parallel to the values measured in the water column with an average offset of −1.6 ‰ and −3.6 ‰ for *N. pachyderma* (sin.) and *T. quinqueloba*, respectively. The discrepancy found in the upper ~ 75 m might indicate the influence of the "carbonate ion effect" on the carbon isotope incorporation in the tests. Oxygen and carbon isotopes from the sediment surface yield higher values than those from the water column for both species. This may be because specimens from the water column reflect a modern snapshot only, while tests from surface sediments record environmental parameters from the past ~ 1000 years.

1 Introduction

The stable isotope composition of foraminiferal shells from sediment cores is often used in palaeoceanographic studies to reconstruct past oceanic environments. Planktic foraminifera precipitate their calcite tests from the surrounding seawater. Thus, variations in the stable oxygen ($^{18}O/^{16}O$, $\delta^{18}O$) and stable carbon ($^{13}C/^{12}C$, $\delta^{13}C$) isotope ratios of shell carbonate should reflect variations of the stable isotope composition of the ambient water. This, on the other hand, is determined by the interplay of various factors (e.g., evaporation, sea-ice formation, productivity). Therefore, δ^{18}O values of fossil planktic foraminiferal shells have been widely used to estimate sea surface temperatures (e.g., Kellogg et al., 1978; Erez and Luz, 1983), salinity (e.g., Duplessy et al., 1992), and ocean stratification (Simstich et al., 2003). Planktic δ^{13}C records are generally assumed to reflect changes in paleoproductivity and ventilation of surface and near-surface waters (Duplessy, 1978). Several studies of living planktic foraminifera showed that these protozoa do not calcify in equilibrium with the ambient sea water (Shackleton et al., 1973; Vergnaud Grazzini, 1976; Kahn, 1979; Kohfeld et al., 1996; Bauch et al., 1997; Volkmann and Mensch, 2001) and that the stable isotope composition of their tests is influenced not only by regional effects but also by numerous other chemical (e.g., pH, carbonate ion concentration) and biological (e.g., symbiont photosynthesis, metabolic activity, test weight) factors. Increasing pH and carbonate ion concentrations are known to decrease stable isotope ratios (Spero et al., 1997; Bauch et al., 2002), while the presence of symbionts causes depletion in ^{18}O and an increase in δ^{13}C values of the shells (Spero and Deniro, 1987). A higher metabolic rate, characteristic for earlier ontogenetic stages (Hemleben et al., 1989), is associated with a discrimination against heavy isotopes (McConnaughey, 1989). Furthermore, planktic foraminifera are assumed to migrate in the water column during ontogeny (Berberich, 1996; Schiebel and Hemleben, 2005) and form their secondary calcite crust while descending into deeper layers (Simstich et al., 2003). Heavier, encrusted individuals are reported to have higher δ^{18}O and lower δ^{13}C values (Kohfeld et al., 1996; Bauch et al., 1997; Volkmann and Mensch, 2001; Simstich et al., 2003). Combinations of these and other factors make the fossil record difficult to interpret. For example, while both increasing temperature and symbiont activity tend to decrease the δ^{18}O value, at the same time both of them can increase the test weight (Hecht, 1976; Spero, 1992), and thereby also indirectly increase the ^{18}O / ^{16}O ratio. Thus, it is essential for an improved interpretation of isotopic records from sediment cores to have detailed knowledge on how modern environmental parameters influence the stable iso-

tope composition of living foraminifera and to which degree the isotopic composition of their shells reflects the composition of the ambient sea water.

Here we report on results from a multidisciplinary approach to investigate the effects of environmental factors on the isotopic composition of polar planktic foraminifera. Our study was performed in the Fram Strait that connects the Arctic and North Atlantic Oceans and comprises strong oceanographic contrasts (Fig. 1). On an E–W transect between the Svalbard and Greenland margins, the upper water column and sediment surface were sampled for planktic foraminifera species *Neogloboquadrina pachyderma* (sinistral) (Ehrenberg, 1861) and *Turburotalita quinqueloba* (Natland, 1983). The stable oxygen and carbon isotope compositions of net-sampled tests are compared to the isotope data of ambient seawater and to that of tests from sediment surface samples. Possible reasons for the revealed discrepancies are discussed. Findings about the species assemblage and depth distribution of foraminifera from the same samples are discussed in Pados and Spielhagen (2014).

2 Hydrographical setting

The Fram Strait is the only deep passage between the Arctic Ocean and the rest of the world ocean, supplying the Arctic Ocean with waters of North Atlantic origin. The oceanography is dominated by two major surface current systems: the northward flowing West Spitsbergen Current (WSC) carries warm and saline water masses, while in the East Greenland Current (EGC) cold and fresh Polar waters exit the Arctic Ocean (Johannessen, 1986). The upper 500 m of the WSC are dominated by Atlantic Water (Rudels et al., 1999), characterized by temperatures higher than 0.5 °C and salinities above 34.5 (Jones, 2001). In the western part of the strait the upper 200 m are characterized by temperatures below 0 °C and salinities between 32 and 34. In the summer, as a result of ice melt, the surface waters have even lower salinities (below 32). A well-developed halocline separates the Polar waters from the Atlantic Layer found between 200 and 500 m (Rudels et al., 2000). The interaction between these two water masses controls the sea-ice coverage in the strait. In summer the ice cover usually extends from Greenland to the East Greenland shelf edge but it can intrude also into the eastern part of the strait in the case of extremely cold winter/spring conditions (Dickson et al., 2000).

The oceanographic system is highly variable and the surface distribution of Polar waters can change on a daily timescale (e.g. Beszczynska-Möller et al., 2011). Oceanographic measurements during cruise ARK XXVI/1 (Beszczynska-Möller and Wisotzki, 2012) provide a detailed profile across the Fram Strait at the time of sampling (Fig. 2). The upper 500 m of the WSC were characterized by salinities near 35 and temperatures between 4 °C and 2 °C. At the surface an approx. 100 m thick, warm layer was observed, with temperatures up to 7–8 °C. During the time of sampling, at stations 35 and 39, cold and fresh water masses intruded into the upper 50–100 m of this layer. After one week these water masses were not present anymore at the same locations (see water column properties at 3° E (station 87) that was sampled 8 days later), in agreement with high variability of oceanic fluxes in this region (Beszczynska-Möller et al., 2011). In the western part of the strait and in the EGC, the Polar waters extended down to ~ 200 m water depth and had salinities around 33 and an average temperature of −1.5 °C at the time of the sampling (Fig. 2). Here, on the very surface low salinities (~ 30) were found that is probably caused by melt water. The Polar waters were mainly ice-covered. Below ~ 200 m water depth submerged warmer and salty Atlantic waters were found. The sea-ice margin over the position of the transect was located at ~ 2° W during the sampling period.

3 Material and methods

Plankton tow, sediment surface and water samples used in this study were obtained during expedition ARK XXVI/1 with research vessel *Polarstern* in June/July 2011 in the Fram Strait. Samples were collected at 10 stations along a transect at 78°50′ N across the Fram Strait (Fig. 1, Table 1). Plankton samples were collected by a MultiNet sampler

(net opening 0.5 m^2, Hydro-Bios, Kiel, Germany); the nets of 63 µm mesh size were towed vertically on regular depth intervals (500–300 m, 300–200 m, 200–100 m, 100–50 m, 50–0 m). Sediment surface samples were obtained from multicorer deployments at the same stations. Sample collection and handling procedures are described in detail by Pados and Spielhagen (2014).

For stable isotope analysis 10–25 specimens of *N. pachyderma* (sin.) and 10 specimens of *T. quinqueloba* were picked from the > 100 µm fraction (plankton samples) and from the 100–250 µm fraction (sediment samples). The number of picked tests was restricted by the number of available, same-sized and clearly identifiable shells. The oxygen and carbon isotope analysis was performed at the GEOMAR Stable Isotope Lab using a Finnigan-MAT 253 mass spectrometer system connected to a Kiel IV Carbonate Preparation Device.

Temperature and salinity of the water column were measured by a conductivity-temperature-depth (CTD) profiler (Beszczynska-Möller and Wisotzki, 2012), immediately before the deployment of plankton tows. Data of ice coverage were obtained by shipboard observations. Water samples for stable isotope analyses were taken from the rosette sampler, immediately after arriving on the deck to minimize the exchange of contained CO_2 with the atmosphere. Water samples for $\delta^{13}C$ analysis (100 mL) were drafted carefully into glass bottles without sputtering and thus avoiding bubbles. Samples were immediately poisoned with $HgCl_2$ (0.2 mL) to stop biochemical reactions, which may alter the carbon isotopic composition of CO_2, and bottles were then sealed by aluminium caps with high-density butyl-rubber plugs. Water samples for $\delta^{18}O$ analysis (50 mL) were filled into glass bottles and sealed by plastic screw-on caps. The oxygen and carbon isotope mass ratios of the water samples were measured in the stable isotope laboratories of AWI Potsdam with a Finnigan-MAT Delta-S mass spectrometer with two coupled equilibration units ($\delta^{18}O$) and of MARUM (Bremen) using a Gasbench coupled to a MAT 252 mass spectrometer ($\delta^{13}C_{DIC}$), with a precision and accuracy of at least ±0.1 ‰ and ±0.15 ‰ for $\delta^{18}O$ (Meyer et al., 2000) and $\delta^{13}C_{DIC}$, respectively.

"Equilibrium calcite values" (δ_c) were calculated for the water samples between sea surface and 500 m water depth from actual δ^{18}O (δ_w) and temperature (T) measurements according to the so called "palaeotemperature equation" of O'Neil et al. (1969):

$$T = 16.9 - 4.38(\delta_c - \delta_w) + 0.1(\delta_c - \delta_w)^2 \quad (1)$$

where δ_c and δ_w are the stable oxygen isotope values of the calcite and the water on the PDB scale, respectively. Isotope measurements of water are presented using SMOW as a standard. δ^{18}O (SMOW) was converted to δ^{18}O (PDB) for the palaeotemperature equation according to Bemis et al. (1998):

$$\delta^{18}\mathrm{O(PDB)} = 0.9998\delta^{18}\mathrm{O(SMOW)} - 0.2‰. \quad (2)$$

Weighted averages of the stable isotope results for each station were calculated using the standing stock (sum of isotope values for each interval multiplied with the concentration of foraminifera in depth interval/sum concentrations in all intervals; see Bauch et al., 1997).

4 Results

4.1 Water column properties

4.1.1 Stable isotopes in the upper water column

In general, oxygen isotope composition of water is roughly linearly correlated to salinity (Fig. 3). Both change along the transect with lowest values at the surface and are continuously increasing with depth. Surface δ^{18}O and salinity are lowest in the west and highest in the east (Fig. 4). The vertical δ^{18}O profiles of the two westernmost stations show strong similarities with the vertical profiles of stations 35 and 39 (4° E

and 2° E, respectively) where during the time of sampling cold and fresh water masses intruded into the warm surface waters. At these four stations extremely low salinity and δ^{18}O values mark Polar waters at the surface that are reaching from the surface down to ~ 200 m water depth at stations 75 and 71 and to ~ 100 m water depth at stations 39 and 35. Below the surface layers the Atlantic waters are characterized by relatively high and constant δ^{18}O values of about +0.3 ‰ (Fig. 4). The remaining six stations reveal more scatter: we found a slight increase from the surface to 25 m water depth, then a decrease to 75–100 m depth, followed by homogeneous δ^{18}O values of the Atlantic Layer (Table 2). When evaluating average values over the upper 500 m of water column, the two westernmost stations (75 and 71) yield the lowest average values (−0.41 ‰, ±0.94 and −0.84 ‰, ±1.16, respectively) while the highest average value (+0.34 ‰, ±0.03) is observed at 3° E at station 87.

The vertical variation of the carbon isotope composition of the dissolved inorganic carbon (DIC) in the upper 500 m of the water column is rather small. The $\delta^{13}C_{DIC}$ values have an average value of about +1.04 ‰ with a standard deviation of 0.24. Surface waters always yield heavier values, decreasing down to ~ 100 m water depth followed by relatively constant values down to 500 m water depth (Fig. 4, Table 2).

4.1.2 Equilibrium calcite values

The equilibrium calcite values calculated from δ^{18}O and temperature of seawater increase constantly with depth in the eastern part of the strait, while at stations 71 and 75, in the western part of the strait, a stronger increase is observed from the surface down to 150 m water depth, followed by a slight decrease. A similar pattern, though in a narrower range, can be found at stations 35 and 39, in the middle of the strait, where near the surface unusually low temperatures of the water were recorded during the sampling period (Fig. 7, Table 2).

4.2 Species distribution of planktic foraminifera

In the Fram Strait the planktic foraminiferal fauna is dominated by polar species *Neogloboquadrina pachyderma* (sinistral coiling) and subpolar symbiont bearing *Turborotalita quinqueloba* (Volkmann, 2000; Simstich et al., 2003; Pados and Spielhagen, 2014). *N. pachyderma* (sin.) clearly prevails in our samples contributing to the total assemblages with 73–92 % compared to the the subpolar *T. quinqueloba*, contributing with 23–5 %. The proportion of *N. pachyderma* (sin.) shows an increasing trend towards the west while *T. quinqueloba* follows an exactly reverse tendency. Both species have maximum absolute abundances between 0–100 m water depth with a shallower maximum under the ice cover than in the open ocean (Figs. 5 and 6). The species assemblage found in the sediment surface was found to correlate best with the living fauna caught between 50–100 m water depth (Pados and Spielhagen, 2014).

4.3 Stable isotope composition of foraminifera

4.3.1 *Neogloboquadrina pachyderma* (sin.)

The oxygen isotope composition of *N. pachyderma* (sin.) from sediment surface samples shows lowest values at the three easternmost stations (stations 127, 25, 19; at 8–6° E). The highest value is found in the western part at station 75, at 4° W. The sediment surface samples have an average δ^{18}O value of +3.2 ‰ (±0.3) (Table 3). The plankton samples generally have a lower oxygen isotope composition than the sediment surface samples. The weighted average δ^{18}O values at each station are ranging from −0.7 ‰ to +3.3 ‰ over the whole transect. There is no clear east-west difference and the stations in the middle of the strait (stations 35, 39 and 54) yield the highest values. The vertical distribution of δ^{18}O does not show a well-pronounced trend. The only clearly recognizable pattern is that values in the depth interval of 0–50 m are usually lower than those from the interval 50–100 m (an exception is station 54 where the surface waters yielded slightly heavier δ^{18}O values) (Fig. 7, Table 4).

The δ^{13}C values of *N. pachyderma* (sin.) from sediment surface samples are relatively constant and have an average value of +0.5 ‰ (±0.2). The lateral distribution of the δ^{13}C values from the sediments shows a trend similar to the δ^{18}O results: the easternmost stations (stations 127, 25, 19; at 8–6° E) have the lowest values while the highest are found in the western part of the section, at the sea ice margin at 2° W (Table 3). The carbon isotope composition of *N. pachyderma* (sin.) from plankton tows shows lower values at each station than the δ^{13}C values from sediments. The δ^{13}C values of plankton samples have a mean of −0.8 ‰ (±0.7) of weighted averages over the whole transect. The plankton samples do not follow the E–W trend found in the core top samples and both the lowest and the highest values are found at stations in the middle of the transect (stations 87 and 39, respectively). Vertical δ^{13}C profiles of the plankton samples show increasing values from the sea surface down to the depth interval of 100–200 m (exceptions are stations 127, 87 and 71 where maximum values lie in the depth interval of 50–100 m), followed by a slight decrease to 300–500 m (Fig. 7, Table 4).

4.3.2 *Turborotalita quinqueloba*

Both oxygen and carbon isotope compositions of *T. quinqueloba* from sediment surface samples show increasing values from east to west; averaging over the transect −0.1 ‰ (±0.2) and +2.7 ‰ (±0.3), respectively (Table 3).

The weighted average of the δ^{18}O and δ^{13}C values of net sampled specimens scatter along the stations and do not show any pronounced east-west difference. The vertical profiles of both parameters show extremely low values near the surface compared to all other depth intervals. The vertical variation of the δ^{18}O and δ^{13}C values in *T. quinqueloba* is much bigger than the variation found in *N. pachyderma* (sin.) (Fig. 8, Table 4). However, it has to be taken into account that the number of specimens available for isotope analysis (on average ten tests) was rather low and possibly not sufficient to yield reliable results for this thin-walled species. We cannot exclude that the differences in variation are due to a non-systematic error in the isotope measurements of *T. quinqueloba*.

5 Discussion

5.1 Oxygen isotope values of water samples and foraminifera

The calculated equilibrium calcite isotope composition represents the theoretical δ^{18}O value of inorganic calcite precipitated in isotopic equilibrium with the surrounding seawater. The offset found between the equilibrium calcite value and the measured δ^{18}O value of foraminiferal tests is commonly described as "vital effect" related to differential isotopic uptake in carbonate organisms compared to equilibrium conditions. The δ^{18}O values of living foraminifera in our work area during early summer were consistently lower than the calculated equilibrium calcite values. The mean offsets were −1.5 ‰ in *N. pachyderma* (sin.) and −3.7 ‰ in *T. quinqueloba*. Based on previously published results, the magnitude of the vital effect in *N. pachyderma* (sin.) appears to vary regionally. Bauch et al. (1997) reported a consistent offset of −1.0 ‰ between equilibrium calcite values and δ^{18}O data of net-sampled *N. pachyderma* (sin.) in the Nansen Basin. Volkmann and Mensch (2001) found an average vital effect of −1.3 ‰ in the Laptev Sea for *N. pachyderma* (sin.) and −1.6 ‰ and −1.3 ‰ in the Fram Strait for *N. pachyderma* (sin.) and for *T. quinqueloba*, respectively. Plankton tows from various hydrographic regimes in the Nordic Seas revealed vital offsets of −1.0 ‰ and −1.1 ‰ for *N. pachyderma* (sin.) and *T. quinqueloba*, respectively (Simstich et al., 2003). Significantly smaller offsets were reported from the western subpolar North Atlantic, calculated from shells collected with sediment traps (Jonkers et al., 2010).

In our study *T. quinqueloba* shows larger offsets between the equilibrium calcite values and the measured δ^{18}O values than *N. pachyderma* (sin.). Earlier works (e.g., Fairbanks et al., 1980; Lončarić et al., 2006) also recorded a larger negative offset in spinose species compared to nonspinose species. Moreover, symbiont-containing

species, like *T. quinqueloba*, are known to be more depleted in ^{18}O as a consequence of higher CO_2 fixation caused by photosynthesis (Bijma et al., 1990; Spero et al., 1997). In *N. pachyderma* (sin.) we found a clear east-west difference in the magnitude of the vital effect along the transect, similar to observations by Volkmann and Mensch (2001). In their study the eastern and western part of the strait yielded significantly different offsets, with highest deviations from the equilibrium calcite values in the west. They concluded that ice coverage increases the magnitude of the vital effect. In our samples in *N. pachyderma* (sin.) the strongest disequilibrium was indeed found at the two ice-covered stations (−4.0 ‰ and −1.8 ‰, at 4° W and 5° W, respectively) and at station 87 (−3.9 ‰, at 4° E). These results are also in line with observations of Bauch et al. (1997) who found slightly increasing isotopic differences between water and plankton samples with decreasing salinity and temperature. Volkmann and Mensch (2001) explained the greater vital offset in the cold and less saline waters of the western Fram Strait with faster growth and consequently higher calcification rates caused by unfavourable conditions. An increased calcification rate decreases the δ^{18}O of tests (McConnaughey, 1989) and may thus increase the disequilibrium. While this hypothesis can explain high offsets at increased calcification rates, the validity of the hypothesis seems rather unlikely as unfavourable conditions generally lead to lower metabolism and thus, decreased calcification rates. Moreover, lower temperatures decrease metabolic rates in all organisms (Hemmingsen, 1960; Gillooly et al., 2001). The abrupt increase in the offset close to the sea-ice margin may rather be explained by increased primary production, associated with the ice margin. During biological production dissolved inorganic carbon is consumed and this considerably increases pH and consequently the carbonate ion concentration ($[CO_3^{2-}]$) of the water (Chierchi and Franson, 2009). Spero et al. (1997) showed that increasing seawater $[CO_3^{2-}]$ decreases the $^{18}O/^{16}O$ ratios of the shells of foraminifera and may thus simultaneously increase the vital effect. Still, the effect of carbonate ion concentrations alone cannot explain the high deviation from equilibrium calcite found at station 87 where no increased primary production is expected. A possible reason for the increased vital effect at stations 71, 75 and 87 might also be a sampling during different ontogenetic stages. *N. pachyderma* (sin.) is known to reproduce on a synodic lunar cycle (Bijma et al., 1990; Schiebel and Hemleben, 2005) and as these three stations were sampled in sequence in the second half of the cruise, it is possible that in the respective samples there were more specimens in early life stages compared to the stations sampled 7–10 days before. Early ontogenetic stages are associated with higher respiration and calcification rates (Hemleben et al., 1989). Rapidly growing skeletons tend to show depletion in both ^{13}C and ^{18}O (McConnaughey, 1989), which could account for the increased vital effect observed at the respective stations.

In contrast to *N. pachyderma* (sin.), the offsets found between equilibrium calcite values and the δ^{18}O values of *T. quinqueloba* do not follow a clear trend along the transect and show great scatter (Fig. 8). However, the low numbers of specimens found in the samples at most of the stations did not allow us to determine δ^{18}O over the whole water column sampled. Moreover, as a consequence, lower numbers of tests (on average ten) were used for stable isotope analysis than in *N. pachyderma* (sin.) (25), which might also account for the scatter in both δ^{13}C and δ^{18}O values in *T. quinqueloba*. We therefore refrain from discussing the vital effect in *T. quinqueloba*.

Our analysis shows that recent specimens of planktic foraminifera from the water column have a lower oxygen isotopic value than fossils on the sediment surface (Figs. 7 and 8). This is in agreement with a number of studies conducted in different regions of the world (e.g., Duplessy et al., 1981; Schmidt and Mulitza, 2002). Berger (1970) suggested in his hypothesis on intraspecific selective dissolution that within one species preferentially the thin-shelled individuals are dissolved during deposition. These tests are secreted during the warmest period of the year and thus, their dissolution increases the average δ^{18}O value of the species in the core top samples. In the Nordic Seas the production maximum of planktic foraminifera occurs during summer (Kohfeld et al., 1996; Jonkers et al., 2010), with almost zero production during other seasons. This means that the majority of the specimens calcifies the shells under similar conditions. Accordingly, differences in the thickness of tests are not to be expected. Therefore the

hypothesis of Berger (1970) cannot explain the isotopic differences between plankton and sediment surface samples in this region. Lateral transport of the shells during deposition is another effect that could explain the discrepancies. However, mean transport distances in the Fram Strait are only 25–50 km for *N. pachyderma* (sin.) and 50–100 km for *T. quinqueloba* (von Gyldenfeldt et al., 2000). These distances are too short to transport isotopic signatures from water masses with significantly different temperature/salinity signatures into the sediments. The offset found in the δ^{18}O values between plankton and sediment surface samples can be attributed rather to the age difference between living plankton and sediment surface samples. Core top samples are assumed to represent modern conditions in palaeoceanographic reconstructions. Nevertheless, depending on sedimentation rates and bioturbation intensity, their average age can vary in a great range (in the Fram Strait a few decades to 3 ky, on average 1 ky, see Simstich et al., 2003) while net-sampled foraminifera reflect a snapshot of actual modern conditions. Discrepancies found between isotopic composition of shells collected on the sediment surface and in the water column may therefore be related to changes in the oceanographic parameters between the early summer of 2011 and average conditions during the period represented by sediment surface samples. To explain the lower modern δ^{18}O values, the water mass in the calcification depth interval of the foraminifera must have become warmer and/or the $\delta^{18}\text{O}_{\text{water}}$ must have decreased and thus, the salinity signature must have changed significantly. Both rising temperatures and increasing river discharges have been documented for the last decades in the Arctic (e.g., Zhang et al., 1998; Serreze et al., 2000; Peterson et al., 2002; Spielhagen et al., 2011). The mean offset found between the δ^{18}O values of net-sampled foraminifera and the tests from the sediment surface along the transect is ~ 1.3 ‰. Assuming that the oxygen isotope composition of the water remained constant over the time, this difference would correspond to a change in water temperature of about 5 °C. Neglecting the two extremely high offsets found at stations 75 and 87, the mean offset would decrease to ~ 0.6 ‰, corresponding to a temperature change of ~ 2.4 °C. A temperature change of 2.4 °C is similar to the reconstructed temperature increase of Atlantic Water during the last 200 years (Spielhagen et al., 2011). However, a temperature change of 5 °C during the last millennia over the whole Fram Strait area seems much too large and clearly, water temperature changes may not solely account for the differences found in the isotopic composition between modern and fossil foraminifera. The results nevertheless suggest the combined effect of temperature rise and $\delta^{18}O_{water}$-change, possible dissolution and transport effects during the last ~ 1000 years.

5.2 Calcification depth

With currently available methods we cannot directly determine the actual calcification depth of planktic foraminifera in the water column. Therefore we assume that planktic foraminifera build their shells at the depth where they are most abundant. The average depth of calcification (calculated from the standing stock) of *N. pachyderma* (sin.) in the Fram Strait lies between 70–150 m water depth. *T. quinqueloba* shows a similar calcification range at 50–120 m water depth (Figs. 5 and 6). Both species show deepest average calcification depth at the easternmost station. Our results are in accordance with Simstich et al. (2003) who calculated an apparent calcification depth for *N. pachyderma* (sin.) of 70–130 m and 70–250 m in the EGC and off Norway, respectively. From the Nansen Basin (eastern Arctic Ocean), Bauch et al. (1997) reported a deeper average calcification depth for *N. pachyderma* (sin.). However, in the northern regime of the Nansen Basin, where the water column properties are similar to those in the western Fram Strait, *N. pachyderma* (sin.) prefers shallower waters than in the southern Nansen Basin where the water column is strongly influenced by the subsurface inflow of Atlantic Waters (Bauch et al., 1997). This trend observed by Bauch et al. (1997) coincides with our results. The difference found in calcification depths in the Nansen Basin and in the Fram Strait might be caused by the different habitats that these locations represent. The northern Nansen Basin is covered by sea ice throughout the year and thus represents a different habitat for planktic foraminifera compared to the narrow Fram Strait. Here, the interannual W–E variability in the position of the average summer

sea ice margin is high and the ice-covered stations sampled in this study might therefore be ice-free in another summer. It has been shown that the depth habitat of planktic foraminifera in the Fram Strait in the early summer is predominantly controlled by the position of the deep chlorophyll maximum (Pados and Spielhagen, 2014). The permanent ice cover in the Nansen Basin may alter the factors controlling the depth habitat of foraminifera and may consequently cause a different depth habitat (and calcification depth) than in the Fram Strait.

Calculating the vital effect from differences between water and plankton samples at each depth level assumes that foraminifera calcified their tests at the depth interval where they were caught. This might not be true, as foraminifera are known to migrate in the water column during their life cycle. Alternatively we may assume that the main encrustation process of foraminifera indeed happened solely at the average calcification depth that is derived in our study from the standing stock. When calculating the average offset between water and foraminifera for the calcification depth only, a vital effect of −0.9 ‰ in *N. pachyderma* (sin.) and −3.1 ‰ in *T. quinqueloba* is determined. These vital effects are significantly smaller than those determined over the whole water column, which are −1.5 ‰ and −3.7 ‰ for *N. pachyderma* (sin.) and *T. quinqueloba*, respectively. In general, we have to take into account that both calcification-scenarios represent extreme cases and the actual vital effect may be between these two extremes.

5.3 Carbon isotope values of DIC and foraminifera

The interpretation of the carbon isotope composition of foraminiferal shells is quite complicated as several factors can influence the carbon isotope incorporation. The gas exchange between sea and atmosphere, the biological production, the community respiration and species dependent incorporations of carbon isotopes are the main processes that can affect the $^{13}C/^{12}C$ ratio in calcite tests. A number of studies reported on a consistent offset between $\delta^{13}C$ of calcite shells and the $\delta^{13}C_{DIC}$ measured within the water column (e.g., Bauch et al., 2000; Volkmann and Mensch, 2001). According to

Romanek et al. (1992) δ^{13}C of inorganic calcite that precipitates in equilibrium with seawater is 1 ‰ higher than $\delta^{13}C_{DIC}$. In our study area below 75 m water depth the δ^{13}C values of *N. pachyderma* (sin.) run relatively parallel to the $\delta^{13}C_{DIC}$, but with an average offset of −1.6 ‰. This reveals a vital effect of about −2.6 ‰. Kohfeld et al. (1996) reported from the Northeast Water Polynya on the Greenland shelf a vital effect of −1 ‰ while another study in the Nansin Basin (Bauch et al., 2000) revealed a vital effect of −2 ‰. The discrepancies found here may suggest the influence of oceanographic variability on the vital effect in δ^{13}C of *N. pachyderma* (sin.). The δ^{13}C of *T. quinqueloba* shows a stronger vertical scatter with an average vital effect of −4.6 ‰. However, in the upper 75–100 m of the water column for both species the $\delta^{13}C_{DIC}$ and the δ^{13}C of shells show an exactly reverse tendency (Figs. 7 and 8). The high $\delta^{13}C_{DIC}$ values found close to the sea surface are assumed to be caused by high primary production, resulting in enrichment in ^{13}C (Fogel and Cifuentes, 1993): as ^{12}C is taken for photosynthesis, the water becomes enriched in ^{13}C. However, if no other processes would affect the incorporation of carbon into the calcite shells, the tests should also show the enrichment in ^{13}C. One possible explanation for the deviation might be an increased carbonate ion concentration ([CO_3^{2-}]) as a consequence of strong biological production in the upper water column (Chierchi and Franson, 2009). Both culturing (Spero et al., 1997) and field experiments (Bauch et al., 2002) have shown that the carbon isotope composition of foraminifera is correlated to the carbonate ion concentration of the water. The "carbonate ion effect" (CIE) describes that increasing seawater [CO_3^{2-}] causes depletion in ^{13}C of the foraminiferal tests. The CIE could therefore explain our observed low δ^{13}C values of shells living in ^{13}C-enriched waters. However, an in-depth interpretation of this effect is not possible as during cruise ARKXXVI/1 the parameters needed to calculate [CO_3^{2-}] (pH and total alkalinity of the water samples) were not determined.

As also discussed with respect to the offset in δ^{18}O between coretop and living foraminifera, the age of core top samples can vary in a great range (between modern to 3 ky, with an average of ~ 1 kyr, Simstich et al., 2003). As a result, they may re-

flect significantly older environments than the plankton samples. The negative offset in δ^{13}C between the sediment and plankton samples may thus be explained by the surface ocean Suess effect: during the last 100 years the carbon isotope composition of the atmosphere has changed due to the increased anthropogenic combustion of fossil carbon which is extremely negative in δ^{13}C. The δ^{13}C values of the atmospheric CO_2 have decreased by about 1.4 ‰ (Friedli et al., 1986; Francey et al., 1999) and the concurrent shift in the stable carbon isotope composition of ocean surface water is reflected in the decrease of δ^{13}C of recent foraminiferal shells (Bauch et al., 2000). The offset of roughly −1 ‰ in δ^{13}C between the sediment and plankton samples observed both in this study and in that of Bauch et al. (2000) may therefore be explained by the different ages of the carbonate in both sample sets.

6 Conclusions

1. The polar species *Neogloboquadrina pachyderma* (sin.) clearly dominates the foraminiferal species assemblage in the Fram Strait in the early summer. Subpolar *Turborotalita quinqueloba* accounts for only 5–25 %.

2. In the study area both species dwell shallower under the ice cover than in the open ocean. The average depth of calcification of *N. pachyderma* (sin.) lies between 70–150 m water depth, *T. quinqueloba* shows a similar range with 50–120 m water depth.

3. When calculating the average vital effect in the oxygen isotope composition for the whole sampled water column, *N. pachyderma* (sin.) and *T. quinqueloba* show an average offset of about −1.5 ‰ and −3.7 ‰ (respectively) compared to calculated equilibrium calcite values. These vital effects are higher than those determined at the calcification depth only where it is −0.9 ‰ for *N. pachyderma* (sin.) and −3.1 ‰ for *T. quinqueloba*.

4. The $\delta^{13}C_{DIC}$ and the $\delta^{13}C$ values of net-sampled shells show an average offset of −1.6 ‰ and −3.6 ‰ for *N. pachyderma* (sin.) and *T. quinqueloba*, respectively. The discrepancies with earlier published results may suggest the influence of regional variability on the vital effect in $\delta^{13}C$.
5. In the upper ~ 75 m of the water column the $\delta^{13}C_{DIC}$ and the $\delta^{13}C$ of shells of both species show an exactly reverse tendency that might relate to the influence of the "carbonate ion effect" on the carbon isotope incorporation in the tests.
6. The shells of both species collected from the water column yield lower $\delta^{18}O$ and $\delta^{13}C$ values than those from the sediment surface, suggesting that the tests recorded changes in environmental parameters influencing the stable isotope compositions in the past ~ 200–1000 years. The negative offset in $\delta^{18}O$ between the sediment and plankton samples suggests a combined effect of temperature rise and $\delta^{18}O_{water}$-change, while the offset in $\delta^{13}C$ may be explained by the surface ocean Suess effect.

Acknowledgements. The authors wish to thank the captain and crew of RV *Polarstern* and the Alfred Wegener Institute, Helmholtz Centre for Polar and Marine Research in Bremerhaven for support during the cruise ARK XXVI/1. We are indebted to Angelina Kraft for assisting in the deployments of the MultiNet sampler and to Maciej Telesinski for collecting the sediment surface samples for this study. Agnieszka Beszczynska-Möller and Andreas Wisotzki provided the CTD-data for the study. Thanks also go to Hartmut Schulz for kindly providing the low temperature vacuum asher at the University of Tübingen and for his hospitality and technical assistance during the process. This research was conducted with funding of the European Community's 7th Framework Programme FP7 2007/2013, Marie-Curie Actions, under Grant Agreement No. 238111 – CASE ITN. R. F. Spielhagen acknowledges the financial support through the "Akademienprogramm" by the Academy of Sciences, Humanities, and Literature Mainz.

The service charges for this open access publication
have been covered by a Research Centre of the
Helmholtz Association.

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Table 1. List of the stations sampled during cruise ARKXXVI/1 in June/July 2011.

Station	Latitude	Longitude	Water depth	Date	Ice cover
PS78-19	78°49.84′ N	6°0.69′ E	2464 m	25.06.11	no
PS78-25	78°49.962′ N	7°0.077′ E	1465 m	26.06.11	no
PS78-35	78°49.772′ N	3°58.380′ E	2335 m	28.06.11	no
PS78-39	78°50.09′ N	1°54.56′ E	2554 m	28.06.11	no
PS78-44	78°49.972′ N	0°4.630′ E	2636 m	29.06.11	no
PS78-54	78°50.02′ N	2°0.21′ W	2714 m	01.07.11	ice margin
PS78-71	78°49.66′ N	5°20.99′ W	684 m	04.07.11	ice covered
PS78-75	78°49.74′ N	3°55.44′ W	1978 m	04.07.11	ice covered
PS78-87	78°50.44′ N	3°0.19′ E	2454 m	06.07.11	no
PS78-127	78°49.84′ N	8°1.33′ E	1019 m	10.07.11	no

Table 2. δ^{18}O, δ^{13}C and equilibrium calcite values in the upper 500 m of the water column.

Station	Depth (m)	δ^{18}O (‰ SMOW)	δ^{13}C (‰ PDB)	Equ. calc.
PS78-19	0	0.29	1.63	2.90
	25	0.38	1.71	3.01
	50	0.4	1.39	3.32
	75	0.34	0.92	3.44
	100	0.21	0.66	3.33
	125		0.95	
	150	0.28	1.06	3.49
	200	0.31	0.84	3.55
	250	0.21		3.50
	300	0.31	0.95	3.66
	400	0.23	1.10	3.77
	500	0.27	1.06	3.91
PS78-25	0	0.32	0.89	2.70
	25	0.37	0.92	2.73
	50	0.33	0.96	3.24
	75	0.3	0.98	3.29
	100	0.31	0.86	3.36
	125	0.3	0.83	3.44
	150	0.28	0.97	3.47
	200	0.29	0.90	3.50
	250			
	300	0.26	0.95	3.53
	400	0.25	0.99	3.70
	500	0.25	0.94	3.94
PS78-35	0	−0.98	1.30	3.06
	25	−0.46	1.30	3.99
	50	0.09	1.03	4.26
	75	0.23	0.96	3.62
	100	0.29	0.70	3.53
	125	0.37	0.73	3.65
	150	0.29	0.99	3.63
	200	0.33	1.05	3.82
	250			
	300	0.33	1.01	3.81
	400	0.23	1.07	3.82
	500	0.29	1.01	4.00
PS78-39	0	−0.29	1.50	3.92
	25	−0.25	1.31	4.27
	50	0.06	1.21	4.64
	75	0.14	1.11	3.99
	100	0.17	0.97	4.02
	125	0.22	1.05	3.82
	150	0.25	1.13	3.67
	200	0.36	1.01	3.87
	250			
	300	0.28	0.82	3.76
	400	0.29	0.79	3.71
	500	0.28	1.09	3.80

Table 2. Continued.

Station	Depth (m)	δ^{18}O (‰ SMOW)	δ^{13}C (‰ PDB)	Equ. calc.
PS78-44	0	0.25	1.39	2.73
	25	0.38	1.16	3.03
	50	0.38	1.01	3.14
	75	0.3	0.88	3.19
	100	0.37	0.89	3.40
	125	0.38	0.98	3.48
	150	0.31	0.96	3.45
	200	0.29	1.19	3.47
	250			
	300	0.28	0.88	3.60
	400	0.35	0.98	3.79
	500	0.28	1.14	3.89
PS78-54	0	0.33	1.68	4.41
	25	0.41	1.61	4.69
	50	0.37	0.94	3.06
	75	0.32	0.95	3.21
	100	0.37	1.06	3.38
	125	0.36	0.94	3.44
	150	0.36	1.10	3.49
	200	0.25	1.09	3.46
	250			
	300	0.21	0.90	3.49
	400	0.36	1.04	3.72
	500	0.3	0.87	3.77
PS78-71	0	−2.63	1.49	1.67
	25	−2.4	1.50	2.15
	50	−2.18	1.23	2.38
	75	−1.41	1.12	3.13
	100	−1	1.11	3.49
	125	−0.52	1.02	3.98
	150	−0.25	1.09	4.24
	200	0.04	1.13	4.15
	250			
	300	0.36	0.94	4.13
	400	0.37	0.95	4.07
	500	0.4	1.18	4.34
PS78-75	0	−2.2	1.34	2.31
	25	−1.85	0.96	2.70
	50	−1.2	1.15	3.34
	75	−0.51	1.10	3.94
	100	−0.27	1.11	4.05
	125	−0.09	0.91	4.16
	150	0.21	0.86	4.24
	200	0.36	0.91	3.34
	250			
	300	0.37	1.01	3.46
	400	0.34	0.87	3.56
	500	0.36	0.99	4.19

Table 2. Continued.

Station	Depth (m)	δ^{18}O (‰ SMOW)	δ^{13}C (‰ PDB)	Equ. calc.
PS78-87	0	0.31	1.54	2.58
	25	0.38	1.20	2.51
	50	0.35	1.08	2.94
	75	0.3	1.02	3.02
	100	0.36	0.64	3.19
	125	0.35	0.70	3.27
	150	0.37	0.69	3.38
	200	0.29	0.86	3.46
	250			
	300	0.36	0.79	3.67
	400	0.33	0.64	3.74
	500	0.37	0.87	3.87
PS78-127	0	0.34	1.64	2.60
	25	0.49	1.31	2.64
	50	0.4	0.95	2.91
	75	0.21	0.85	2.76
	100	0.38	0.87	2.96
	125	0.32	0.93	2.94
	150	0.36	0.88	3.03
	200	0.34	0.94	3.07
	250			
	300	0.28	0.96	3.22
	400	0.32	1.00	3.40
	500	0.3	1.02	3.49

Table 3. δ^{18}O and δ^{13}C values of *N. pachyderma* (sin.) (*N.p.* (sin.)) and *T. quinqueloba* (*T.q.*) from sediment surface samples.

	δ^{18}O (‰ PDB)		δ^{13}C (‰ PDB)	
Station	*N. p.* (sin.)	*T. q.*	*N. p.* (sin.)	*T. q.*
PS78-19	3.17	2.56	0.44	−1.10
PS78-25	2.84	2.05	0.15	−1.28
PS78-35	3.36	2.70	0.66	−1.05
PS78-39	3.31	2.53	0.61	−1.20
PS78-44	3.35	2.98	0.49	−1.09
PS78-54	3.44	3.08	0.75	−0.59
PS78-71	3.32		0.50	
PS78-75	3.57	3.00	0.72	−0.67
PS78-87	3.36	2.75	0.58	−1.02
PS78-127	2.71		0.09	

Table 4. δ^{18}O and δ^{13}C values of *N. pachyderma* (sin.) (*N. p.* (sin.)) and *T. quinqueloba* (*T. q.*) from plankton tows.

		δ^{18}O (‰ PDB)		δ^{13}C (‰ PDB)	
Station	Depth (m)	*N. p.* (sin.)	*T. q.*	*N. p.* (sin.)	*T. q.*
PS78-19	0–50	2.38		−0.99	
	50–100	2.69	1.96	−0.40	−1.64
	100–200	2.82	1.81	−0.38	−1.88
	200–300	2.36		−0.74	
	300–500	2.67		−0.78	
PS78-25	0–50	2.40		−0.78	
	50–100	2.55	−2.85	−0.37	−4.25
	100–200	2.51	1.64	−0.10	−1.72
	200–300	1.98		−1.40	
	300–500	2.56	0.31	−0.43	−2.50
PS78-35	0–50	2.32	−5.54	−0.89	−6.13
	50–100	3.08	−5.57	−0.28	−6.03
	100–200	3.00	2.25	−0.15	−1.68
	200–300	2.95		−0.16	
	300–500	2.88		−0.44	
PS78-39	0–50	3.25	−4.99	−0.19	−5.79
	50–100	3.25	−1.36	−0.13	−3.61
	100–200	3.34	1.57	0.08	−1.99
	200–300	3.34		0.03	
	300–500	2.32		−0.75	
PS78-44	0–50	2.58	−7.35	−0.60	−7.86
	50–100	2.64		−0.44	
	100–200	3.01	2.13	−0.10	−1.44
	200–300	2.86	0.41	−0.25	−2.46
	300–500	2.65	−0.51	−0.83	−5.78
PS78-54	0–50	2.77	−0.86	−0.62	−5.96
	50–100	2.75	2.46	−0.21	−1.22
	100–200	2.83	2.20	−0.21	−1.56
	200–300	2.69	1.59	−0.24	−2.57
	300–500	2.61	2.18	−0.28	−1.52
PS78-71	0–50	0.99		−0.63	
	50–100	1.86	1.72	−0.14	−1.78
	100–200	1.80	−1.93	−1.04	−3.84
	200–300	2.89	0.85	−0.15	−2.30
	300–500	2.12		−0.50	
PS78-75	0–50	−2.94		−4.14	
	50–100	1.36	−2.72	−0.79	−4.00
	100–200	0.93		−0.69	
	200–300	0.56		−1.38	
	300–500				
PS78-87	0–50	−1.54		−2.14	
	50–100	1.87	0.25	−0.43	−3.54
	100–200	1.89	0.25	−1.37	−2.22
	200–300	0.23		−1.88	
	300–500	−2.56		−4.38	
PS78-127	0–50	1.07	−5.01	−2.59	−7.79
	50–100	2.80	1.99	−0.48	−1.72
	100–200	2.06	2.39	−0.98	−1.51
	200–300	2.08	2.42	−1.02	−1.54
	300–500	2.16	1.90	−1.06	−1.76

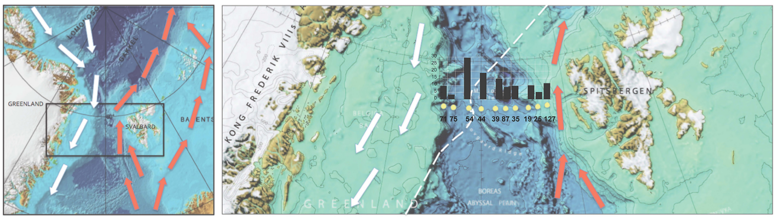

Figure 1. Map and schematic surface ocean circulation of the study area (red arrows: West Spitsbergen Current, white arrows: East Greenland Current). The enlarged part shows the sampled stations (yellow dots). Bars represent absolute abundances (ind./m^3 water) of planktic foraminifera in the upper 500 m of the water column along the transect at 78°50′ N across the Fram Strait. The white dashed line shows the position of the sea ice margin during the sampling period (shipboard observation and satellite data from the US National Oceanic and Atmospheric Administration). Map source: the International Bathymetric Chart of the Southern Ocean (Jakobsson et al., 2012).

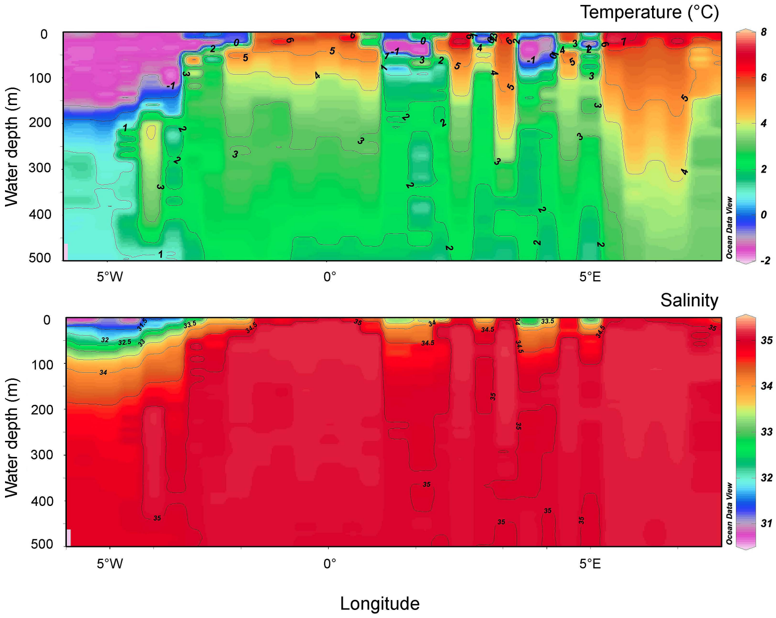

Figure 2. Temperature and salinity of the water column in the upper 500 m along a transect at 78°50′ N across the Fram Strait. Data obtained by conductivity–temperature–density (CTD) measurements during the expedition ARK XXVI/1 (Beszczynska-Möller and Wisotzki, 2012). The figure comprises data obtained from 30 CTD stations, equally distributed along the transect.

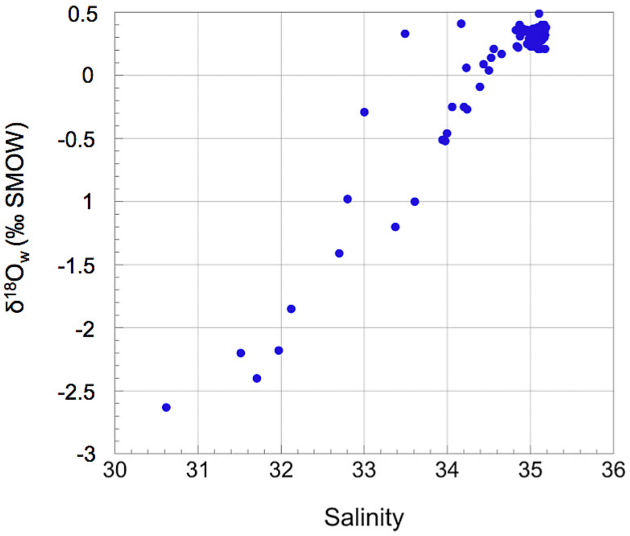

Figure 3. δ^{18}O (‰ SMOW) values of the upper 500 m of the water column vs. salinities (PSU) for the sampled transect at 78°50′ N across the Fram Strait.

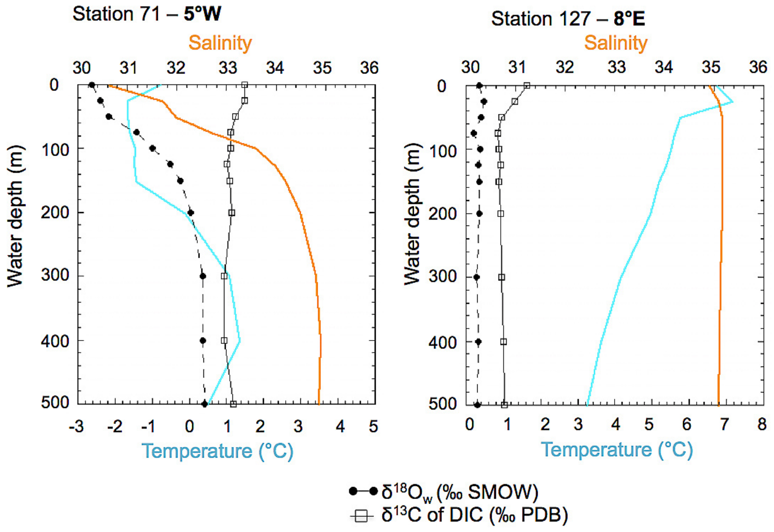

Figure 4. Salinity, temperature, δ^{18}O (‰ SMOW) and $\delta^{13}C_{DIC}$ (‰ PDB) profiles of the upper 500 m of the water column from the westernmost and easternmost stations sampled along a transect at 78°50′ N across the Fram Strait.

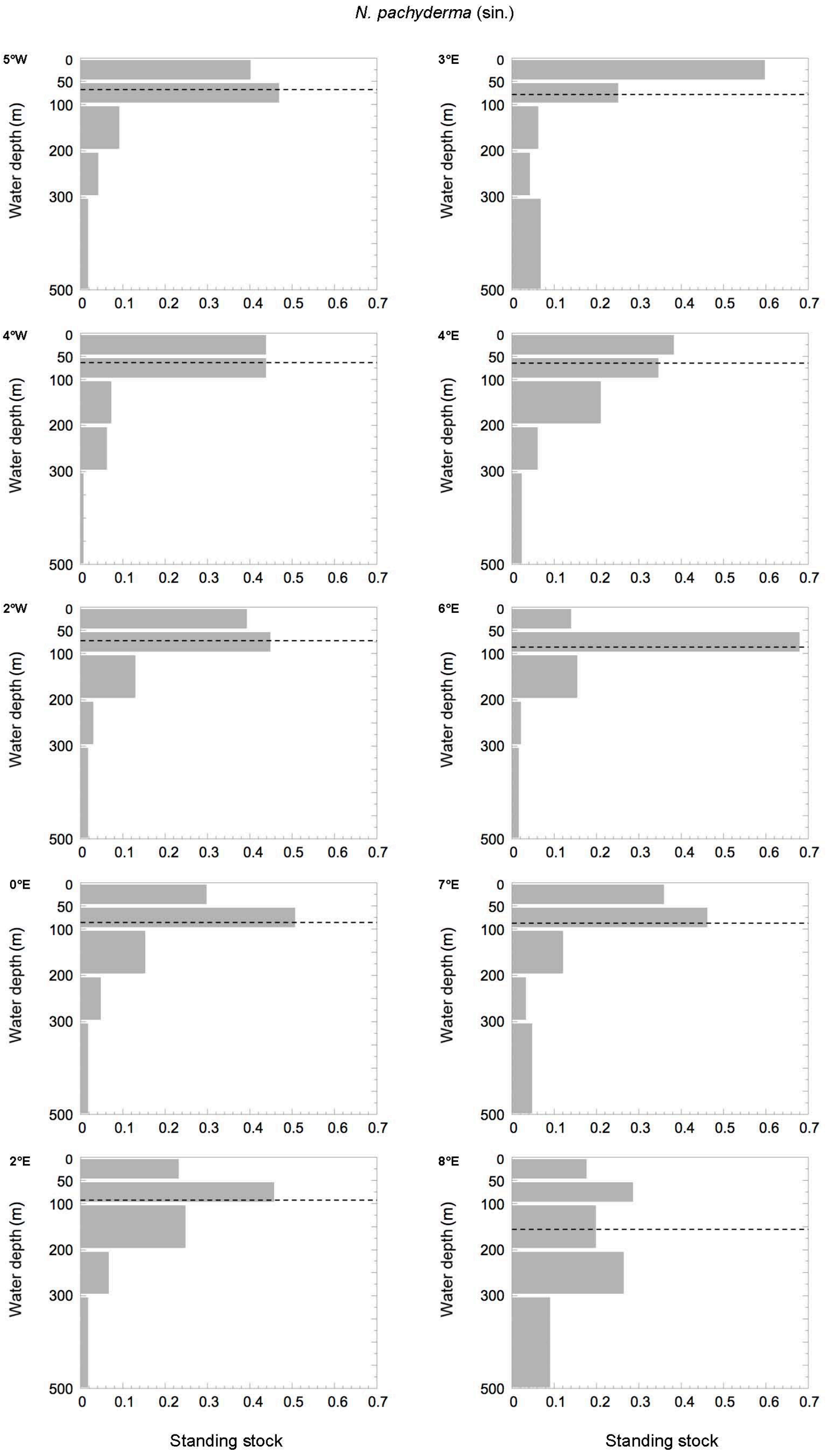

Figure 5. Standing stock of *N. pacyhderma* (sin.) for each sampling interval. The black dashed line indicates the average depth of calcification (calculated from the standing stock).

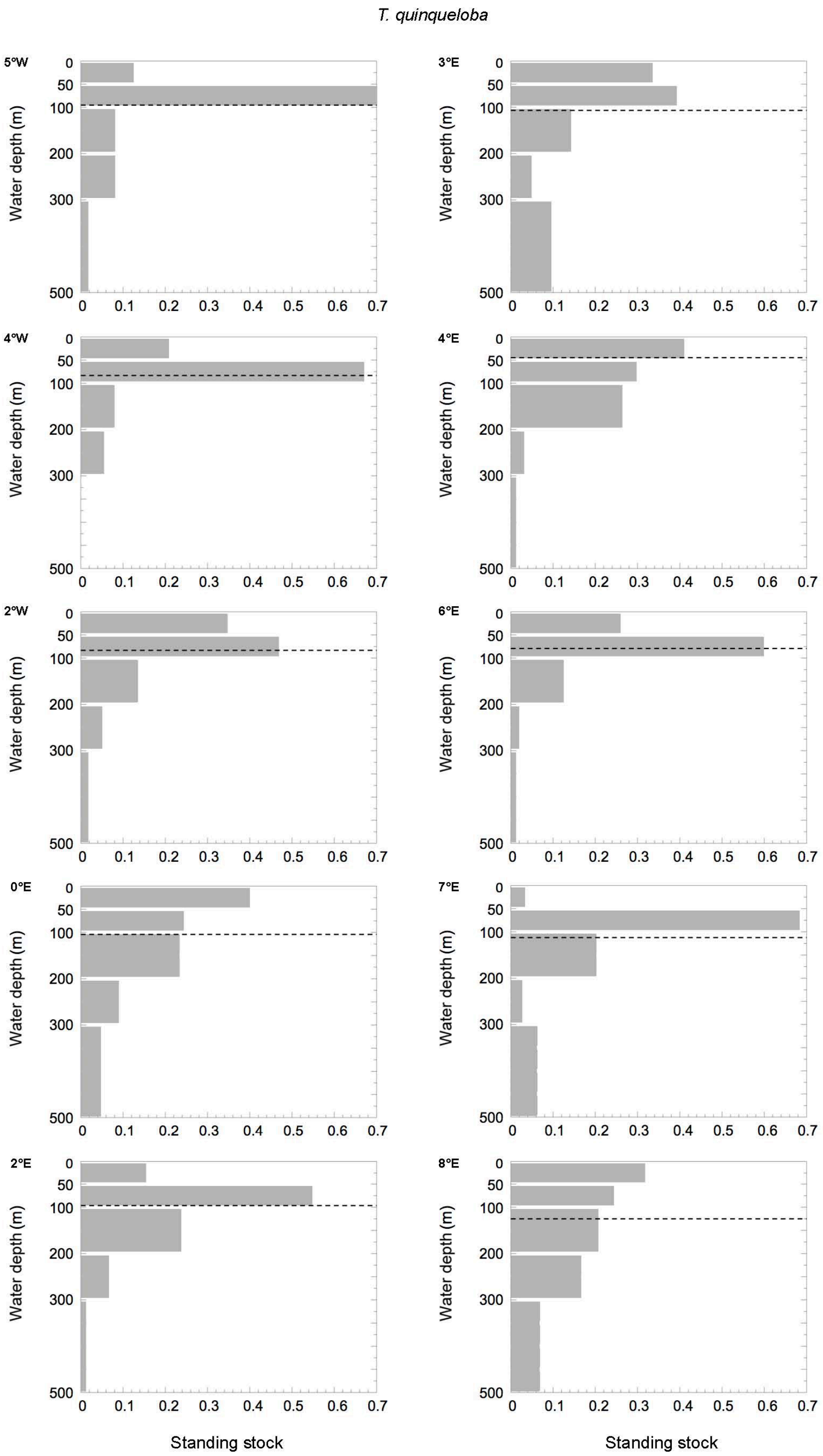

Figure 6. Standing stock of *T. quinqueloba* for each sampling interval. The black dashed line indicates the average depth of calcification (calculated from the standing stock).

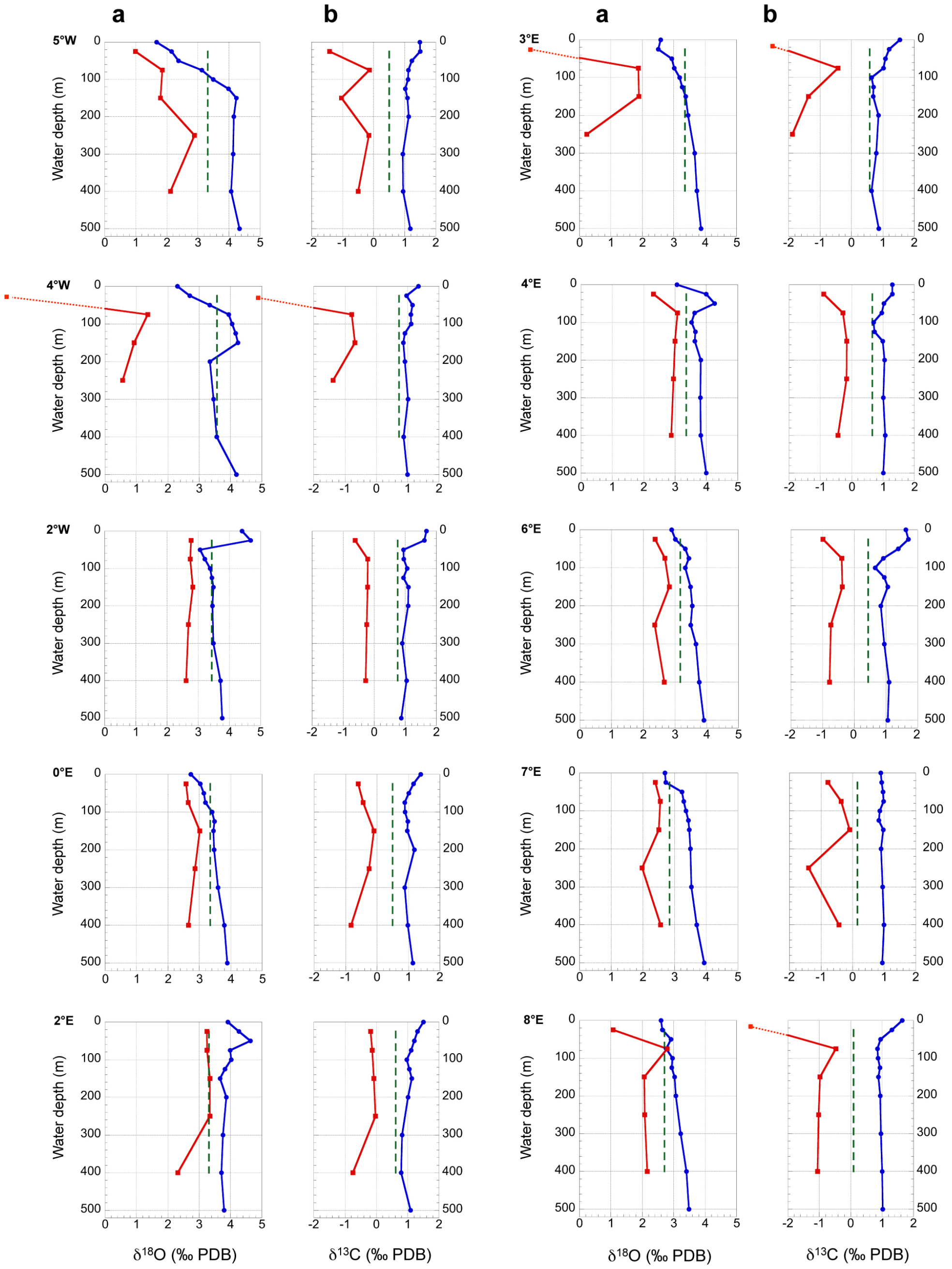

Figure 7. (a) δ^{18}O and **(b)** δ^{13}C values of *N. pachyderma* (sin.) from the water column (red squares) and from the sediment surface (green dashed line). The blue dots indicate **(a)** the equilibrium calcite and **(b)** the $\delta^{13}C_{DIC}$ profile of the water column.

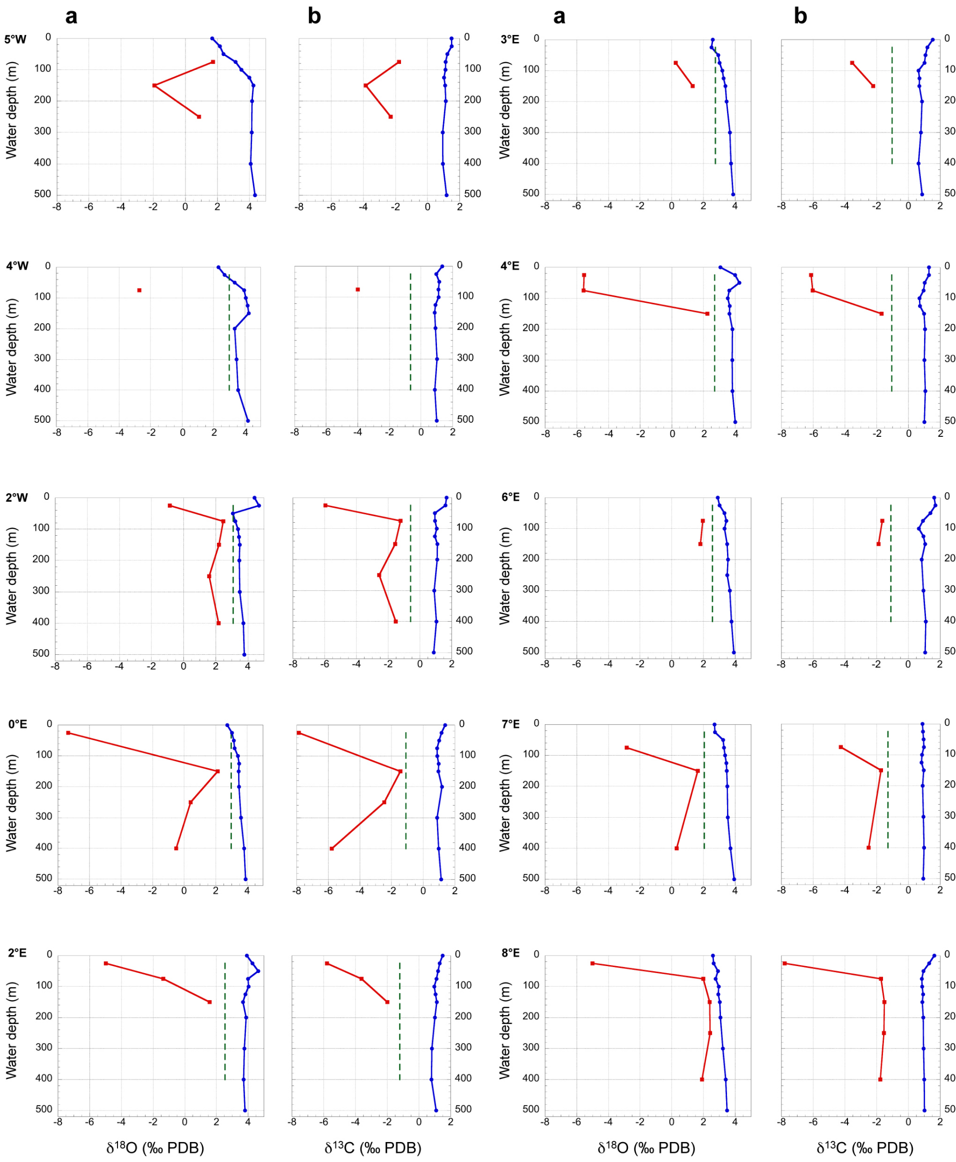

Figure 8. **(a)** $\delta^{18}O$ and **(b)** $\delta^{13}C$ values of *T. quinqueloba* from the water column (red squares) and from the sediment surface (green dashed line). The blue dots indicate **(a)** the equilibrium calcite and **(b)** the $\delta^{13}C_{DIC}$ profile of the water column.

4. Mg/Ca in water samples and in planktic foraminifera from plankton tows and sediment surface in the Fram Strait

Mg/Ca in water samples and in planktic foraminifera from plankton tows and sediment surface in the Fram Strait

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Abstract

To resolve the uncertainties of palaeotemperature reconstructions based on Mg/Ca thermometry in high latitude-regions, recent planktic foraminifera were collected from the water column and from the sediment surface in the Fram Strait. The Mg/Ca ratios measured in the foraminiferal tests were compared to *in situ* water column properties, like temperature, salinity and Mg/Ca ratios of the seawater. The two dominating species, *N. pachyderma* (sin.) and *T. quinqueloba* show a clear offset in Mg/Ca ratios, which might be due to the effect of secondary calcification and/or different calcification depths. The observed elevated shell Mg/Ca ratios and their weak correlation with *in situ* water temperatures cannot be explained with the influence of lateral transport, incompletely calcified shells or the Mg/Ca ratios of the water column, but the observed anomalous results might be related to the seawater carbonate chemistry in the area.

4. 1 Introduction

Mg/Ca thermometry in planktic foraminiferal calcite has become a widely used proxy method for reconstructing past sea surface temperatures. While another commonly used foraminiferal palaeotemperature proxy, the oxygen isotope composition ($\delta^{18}O$) is influenced not only by temperature but also by the oxygen isotope composition of the surrounding seawater ($\delta^{18}O_w$), the Mg/Ca ratio of the tests has been shown to have an exponential relationship with the calcification temperature (e.g., Nürnberg et al., 1995; Lea et al., 1999; Dekens et al., 2002; Anand et al., 2003; Regenberg et al., 2009). Thus, combination of Mg/Ca and $\delta^{18}O$ measurements in the same samples theoretically allows to reconstruct both temperature and $\delta^{18}O_w$ changes distinctly. Moreover, inter-species differences in Mg/Ca due to different depth habitats provide a tool to reconstruct changes in the water column structure from a single sediment core (Elderfield & Ganssen, 2000). Several studies have been conducted to calibrate Mg/Ca versus temperature for different planktic foraminifera species using core-top, culture and sediment-trap samples. Calibrations with core-top samples (Nürnberg et al., 1995; Elderfield & Ganssen, 2000; Dekens et al., 2002; Kozdon et al., 2009; Regenberg et al., 2009) have the advantage that the studied material will eventually form the palaeoceanographic record and that the analysed tests most probably have gone through a complete life cycle, potentially including secondary calcification (Barker et al., 2005). The disadvantage of this method is that corrections for post-depositional alterations, like partial dissolution might be needed (Dekens et al., 2002; Regenberg et al., 2006; Regenberg et al., 2014). Further, the calcification temperature must be derived from $\delta^{18}O$ measurements or from hydrographical databases and thus, it becomes a dependent variable that can introduce an error into the calibration (Anand et al., 2003). In culturing studies (Nürnberg et al., 1996; Lea et al., 1999; Maschiotta et al., 1999) foraminifera calcify under defined parameters in a controlled environment, but laboratory conditions may not fully reproduce natural processes and consequently, natural chamber growth. Moreover, the number of species that can be cultured is limited. Sediment-trap material, analysed by Anand et al. (2003) and Jonkers et al. (2013), represents closely the planktic assemblage in the sedimentary record, thereby avoiding post-depositional alterations. It also allows tracking the Mg incorporation into different species during seasonal cycles. However, in these studies the depth habitat and calcification temperature of foraminifera is estimated from measured ($\delta^{18}O_{foraminifera}$) and calculated ($\delta^{18}O_w$) variables. Another way of water column sampling is the collection of planktic foraminifera by nets. Using this approach for calibration, we have the possibility to determine the actual depth habitat of the studied species and to obtain, parallel to the plankton samples, *in situ* water column properties (temperature, salinity, $\delta^{18}O$ and Mg/Ca value) from the same locations. In this way we minimize the uncertainties concerning the temperature that controlled the Mg incorporation into the tests.

Here we report on the results of Mg/Ca thermometry in planktic foraminifera collected in the water column and on the sediment surface, combined with *in situ* measurements of water column properties from the Fram Strait. This strait connects the North Atlantic Ocean with the Arctic Ocean and comprises strong oceanographic contrasts (Fig. 4.1). The Mg/Ca signal of foraminiferal tests from such high-latitude areas has been shown not to follow the logarithmic temperature calibration curve (Nürnberg, 1995; Elderfield & Ganssen, 2000), which increases the uncertainty of palaeotemperature reconstructions based on Mg/Ca thermometry in cold water-regions. Meland et al. (2006) and Kozdon et al. (2009) analysed the dominating planktic foraminifera species, *Neogloboquadrina pachyderma* (sinistral) from sediment surface samples from the Nordic Seas and found that in Polar waters the temperature information in the Mg/Ca signal is lost. Jonkers et al. (2013) applied sediment-trap time series and assumed that temperature might not be the driving factor behind *N. pachyderma* (sin.) Mg/Ca changes in the Irminger Sea. Here we test the reported limitations of the Mg/Ca approach in high-latitude areas on the polar *N. pachyderma* (sin.) and the subpolar *Turborotalita quinqueloba* sampled from the upper water column. We compare the Mg/Ca compositions of net-sampled tests to the Mg/Ca values of ambient seawater and to that of shells from sediment surface samples, and evaluate the temperature dependence of Mg/Ca in the tests. Eventually we discuss factors other than temperature that might control the Mg/Ca proxy signal in the two dominating planktic foraminifera species in this high latitude region.

4.2 Material and methods

4.2.1 Study area

Plankton tow, sediment surface and water samples used in this study were obtained during expedition ARK XXVI/1 with research vessel *Polarstern* in June/July 2011 in the Fram Strait. Samples were collected at 10 stations along a transect at 78°50'N across the Fram Strait (Fig. 4.1, Table 4.1). The samples from station 127 did not contain enough foraminifera for Mg/Ca measurements; this station is therefore excluded from the study.

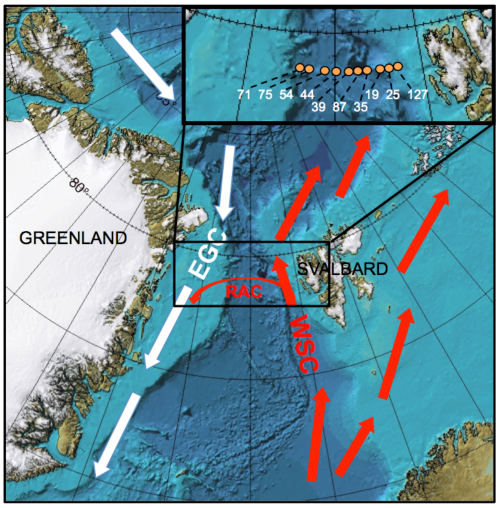

Figure 4.1. Map and schematic surface circulation of the Fram Strait. Red arrows indicate the West Spitzbergen Current (WSC) and the Return Atlantic Current (RAC), white arrows indicate the East Greenland Current (EGC). The enlarged part shows the sampled stations (orange dots with station numbers) along the transect at 78°50'N across the Fram Strait. Map source: the International Bathymetric Chart of the Southern Ocean (Jakobsson et al., 2012).

The oceanographic properties of the study area during sampling have been extensively discussed previously (Pados & Spielhagen, 2014). Briefly, the area is dominated by two major surface current systems: in the West Spitsbergen Current (WSC) warm and saline water masses flow northwards while the East Greenland Current (EGC) carries cold and fresh Polar waters from the Arctic Ocean into the Atlantic Ocean (Johannessen, 1986) (Fig. 4.1). Oceanographic measurements during cruise ARK XXVI/1 (Beszczynska-Möller & Wisotzki, 2012) provide a detailed profile across the Fram Strait at the time of sampling (Fig. 4.2). In the WSC at the surface an approx. 100 m thick, relatively warm layer was observed, with temperatures up to 7-8°C. The remaining water column in the upper 500 m was characterized by temperatures between 2-4°C and salinities near 35. During the time of sampling, between 2°E and 4°E, cold and fresh water masses intruded into the upper 50-100 m. After one week these water masses were apparently not present anymore at the same location (see water column properties at the station 87 that was sampled 8 days later; Fig. 4.2 and Beszczynska-Möller & Wisotzki, 2012), in agreement with the high variability of oceanic fluxes in this region observed earlier (Beszczynska-Möller et al., 2011). The Polar waters in the EGC extended down to ~200 m water depth and had an average temperature of −1.5°C and salinities around 33 at the time of the sampling. Below ~200 m water depth submerged warmer and salty Atlantic waters were found (Fig. 4.2). The Polar waters were mainly ice-covered (sea-ice margin located at ~2°W) and on the very surface low salinities (~30) were found that were probably caused by meltwater.

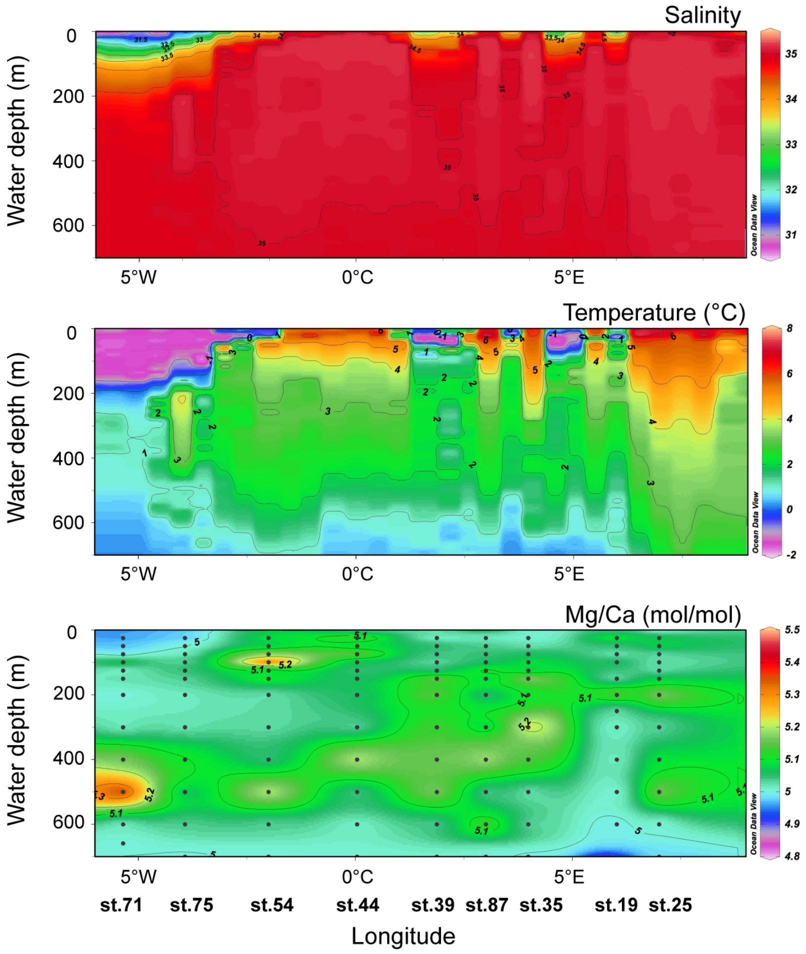

Figure 4.2. Salinity, temperature and Mg/Ca ratios of the water column in the upper 700 m along a transect at 78°50'N across the Fram Strait. The salinity and temperature figures comprise data from 30 CTD stations, equally distributed along the transect (Beszczynska-Möller & Wisotzki, 2012). The Mg/Ca figure displays Mg/Ca ratios measured in selected water depths at nine sampled stations (black dots).

4.2.2 Foraminifera/water sampling and analyses

Plankton samples were collected by a MultiNet sampler (net opening 0.5 m^2); the nets of 63 µm mesh size were towed vertically on regular depth intervals (500-300 m, 300-200 m, 200-100 m, 100-50 m, 50-0 m). Sediment surface samples were obtained from multicorer deployments at the same stations. Sample collection and handling procedures are described in detail in Pados & Spielhagen, 2014. Temperature and salinity of the water column were measured by a Conductivity-Temperature-Depth (CTD) profiler (Beszczynska-Möller & Wisotzki, 2012), directly before the deployment of plankton tows. Data of ice coverage were obtained by shipboard observations. Water samples for stable isotope and Mg/Ca analyses were taken from the rosette sampler, filled into glass bottles (50 ml) and sealed by plastic screw-on caps. Stable isotope measurement procedures and database for $\delta^{18}O$ values are described in Pados et al. (2014).

A new analytical method was developed for the determination of molar Mg/Ca and Sr/Ca ratios in seawater using ICP-optical emission spectrometry (ICP-OES). Seawater samples were 50-fold diluted and introduced into the simultaneous ICP spectrometer (Spectro Ciros SOP), equipped with micronebulizer (200 µl/min sample uptake, SeaSpray™, Glass Expansion, Australia) and cyclonic spray chamber (Glass Expansion, Australia) at the Institute of Geosciences, CAU Kiel University. A series of reagent blanks, calibration solutions and reference standards were analysed along with the samples. Emission lines at 315.887 nm and 317.933 nm (Ca), 279.079 nm and 279.806 nm (Mg), 407.771 and 421.552 nm (Sr), and 597.159 nm (Ar) were monitored during data acquisition, and Sr-407, Mg-279.0, and Ca-317 giving lowest analytical error were selected for subsequent data evaluation. Data read out was truly simultaneous after 20 s measuring time for all spectral lines. Raw data output was then transferred to spreadsheet software and processed manually. We applied a combination of intensity ratio calibration (De Villiers et al., 2002) and external normalization (Schrag, 1999) for the calculation of molar Mg/Ca and Sr/Ca ratios. IAPSO seawater was used for external normalization of Mg/Ca = 5.185 mmol/mmol and Sr/Ca = 8.286 µmol/mmol. Further details on instrumentation and analytical procedures can be found in Garbe-Schönberg et al. (in prep). The average internal analytical error of 5 runs was 0.07 and 0.08 %RSD (1sigma) for Mg/Ca and Sr/Ca, respectively. The external error as estimated from 5 replicate analyses over 18 hours was 0.11 and 0.12 %RSD for Mg/Ca and Sr/Ca.

For foraminiferal Mg/Ca measurements 10-55 specimens of *N. pachyderma* (sin.) and 15-40 specimens of *T. quinqueloba* were picked from the 100-250 µm fraction (plankton net and sediment samples). The number of picked tests was restricted by the number of available, same-sized and clearly identifiable shells. The picked shells were crushed and homogenized. The monospecific samples were cleaned with methanol, hydrogen peroxide and repeated treatments with ultrapure water, followed by a weak acid leach and final dissolution (Martin & Lea, 2002; Barker et al., 2003). Element analyses were performed at the Kiel University, on the Spectro CircoCCD ICP OES (analytical error for Mg/Ca ratios ±0.1‰). The certified reference material ECRM 752-1 with an expected Mg/Ca ratio of 3.9 mmol/mol (Greaves et al., 2008) was run as standard. The measurements revealed low mean Mn/Ca and Fe/Ca ratios of 0.04 (±0.03) mmol/mol and 0.68 (±0.53) mmol/mol (respectively) in 22 samples, indicating proper cleaning.

Correlation analysis between Mg/Ca ratios, temperature, and salinity measured in the water column was performed by Spearman rank correlation.

Table 4.1. List of the stations sampled during cruise ARKXXVI/1 in June/July 2011.

Station	Latitude	Longitude	Water depth	Date	Ice cover
PS78-19	78°49.84´N	6°0.69´E	2464 m	25.06.11	no
PS78-25	78°49.962´N	7°0.077´E	1465 m	26.06.11	no
PS78-35	78°49.772´N	3°58.380´E	2335 m	28.06.11	no
PS78-39	78°50.09´N	1°54.56´E	2554 m	28.06.11	no
PS78-44	78°49.972´N	0°4.630´E	2636 m	29.06.11	no
PS78-54	78°50.02´N	2°0.21´W	2714 m	01.07.11	ice margin
PS78-71	78°49.66´N	5°20.99´W	684 m	04.07.11	ice covered
PS78-75	78°49.74´N	3°55.44´W	1978 m	04.07.11	ice covered
PS78-87	78°50.44´N	3°0.19´E	2454 m	06.07.11	no
PS78-127	78°49.84´N	8°1.33´E	1019 m	10.07.11	no

4.3 Results

4.3.1 Mg/Ca in the water column

The Mg/Ca ratio of the water samples varies between 5.4 and 4.9 mol/mol (Table 4.2). The two highest values were measured at station 54 in 100 m water depth and at station 71 in 500 m water depth (2°W and 5°W, respectively). Moreover, high values mark the halocline that separates the Atlantic Layer from the Polar Mixed Layer in ~500 m water depth. Low values characterise water masses below this halocline and on the surface of Polar waters in the EGC (stations 71 and 75) and in the upper 100 m at stations 35 and 39 where cold and fresh water masses intruded in the upper layers of the water column (Fig. 4.2). Correlation analysis between Mg/Ca ratios, salinity and temperature of the water column revealed stronger correlation between Mg/Ca and temperature ($r=0.539$, $p<0.01$; Spearman rank correlation) than between Mg/Ca and salinity ($r=0.470$, $p<0.01$; Spearman rank correlation).

4.3.2 Mg/Ca in planktic foraminifera

The Mg/Ca ratio in polar *Neogloboquadrina pachyderma* (sin.) from the water column is varying between 1.26 and 2.89 mmol/mol, in subpolar *Turborotalita quinqueloba* between 2.15 and 2.88 mmol/mol. The values of *N. pachyderma* (sin.) are on average 0.7 mmol/mol lower than the values measured in *T. quinqueloba*, taken at the same stations from the same depth intervals. The Mg/Ca ratio of tests from the sediment surface is ranging from 1.11 to 2.65 mmol/mol in *N. pachyderma* (sin.) and from 1.62 to 2.86 mmol/mol in *T. quinqueloba* (Table 4.3). The offsets between the two species from the same sediment surface samples do not show a well-pronounced trend. If we compare the Mg/Ca values of the shells of given species collected in the water column to those from the sediment surface at the same locations, the offset is ranging from −0.84 to +1.55.

Table 4.2. Mg/Ca ratios measured in the water column at certain water depths.

Station	Depth (m)	Mg/Ca (mol/mol)
PS78-19	0	5.037
	25	5.107
	50	5.042
	75	5.008
	100	5.073
	150	5.007
	200	5.168
	250	5.007
	300	4.998
	400	5.006
	500	4.995
	600	4.989
PS78-25	0	4.993
	25	5.068
	50	5.089
	75	5.021
	100	5.021
	125	5.004
	150	5.087
	200	5.178
	300	5.025
	400	5.091
	500	5.182
	600	4.996
	700	4.987
PS78-35	0	4.984
	25	4.996
	50	5.013
	75	5.060
	100	5.099
	125	5.037
	150	5.237
	200	5.110
	300	5.262
	400	5.143
	500	5.000
	600	5.011
	700	4.994
PS78-39	0	5.005
	25	4.991
	50	5.051
	75	5.027
	100	4.997
	125	5.003
	150	5.218
	200	5.175
	300	5.144
	400	5.160
	500	5.206
	600	5.014
	700	5.008
PS78-44	0	5.019
	25	5.158
	50	5.075
	75	5.183
	100	5.036
	125	5.024
	150	5.047
PS78-44	200	5.047
	300	5.008
	400	5.236
	500	5.008
	600	5.001
	700	5.002
PS78-54	0	5.086
	25	5.094
	50	5.063
	75	5.194
	100	5.452
	125	5.024
	150	5.126
	200	5.011
	300	5.019
	400	5.022
	500	5.244
	600	5.023
	700	5.000
PS78-71	0	4.954
	25	4.954
	50	4.965
	75	4.973
	100	5.113
	125	4.988
	150	5.020
	200	4.996
	300	5.002
	400	5.165
	500	5.440
	600	4.992
	660	4.998
PS78-75	0	4.963
	25	4.966
	50	4.991
	75	4.993
	100	5.002
	125	5.048
	150	5.023
	200	5.012
	300	5.010
	400	5.105
	500	5.018
	600	5.028
	700	4.996
PS78-87	0	5.091
	25	5.013
	50	5.036
	75	5.044
	100	5.043
	125	5.151
	150	5.039
	200	5.003
	300	5.031
	400	5.227
	500	4.988
	600	5.179
	700	4.997

Shell Mg/Ca in *N. pachyderma* (sin.) from plankton tows does not show a clear correlation with temperature (Fig. 4.3a) or Mg/Ca ratios (Fig. 4.3g) of the water mass from where the tests were collected. Comparing the $\delta^{18}O_{calcite}$ values to the Mg/Ca ratios in the same shells, we can see a broad range of Mg/Ca values (1.3-2.9 mmol/mol) relating to relatively similar $\delta^{18}O_{calcite}$ values (Fig. 4.3c). In *T. quinqueloba* the correlation seems to be stronger between Mg/Ca ratio and temperature than between Mg/Ca and $\delta^{18}O_{calcite}$ (Fig. 4.3b and 4.3d), but the low number of samples containing enough specimens of *T. quinqueloba* for Mg/Ca analysis allowed only 6 measurements, which may be too few for a robust evaluation.

To compare the calcification temperature of *N. pachyderma* (sin.) from core top samples, the average depth of calcification was calculated from the modern standing stock, assuming modern-like habitats for the time interval represented by a surface sample (see Pados et al. 2014). When doing so, above 0°C the Mg/Ca values of *N. pachyderma* (sin.) from core tops seem to decrease with increasing calcification temperature (Fig. 4.3e). At the same time, we cannot observe any correlation with the $\delta^{18}O_{calcite}$ values measured in the same samples (Fig. 4.3f). Because of the low number of *T. quinqueloba* samples from sediment surface (four) we refrain from discussing the temperature dependence of Mg/Ca content in these shells.

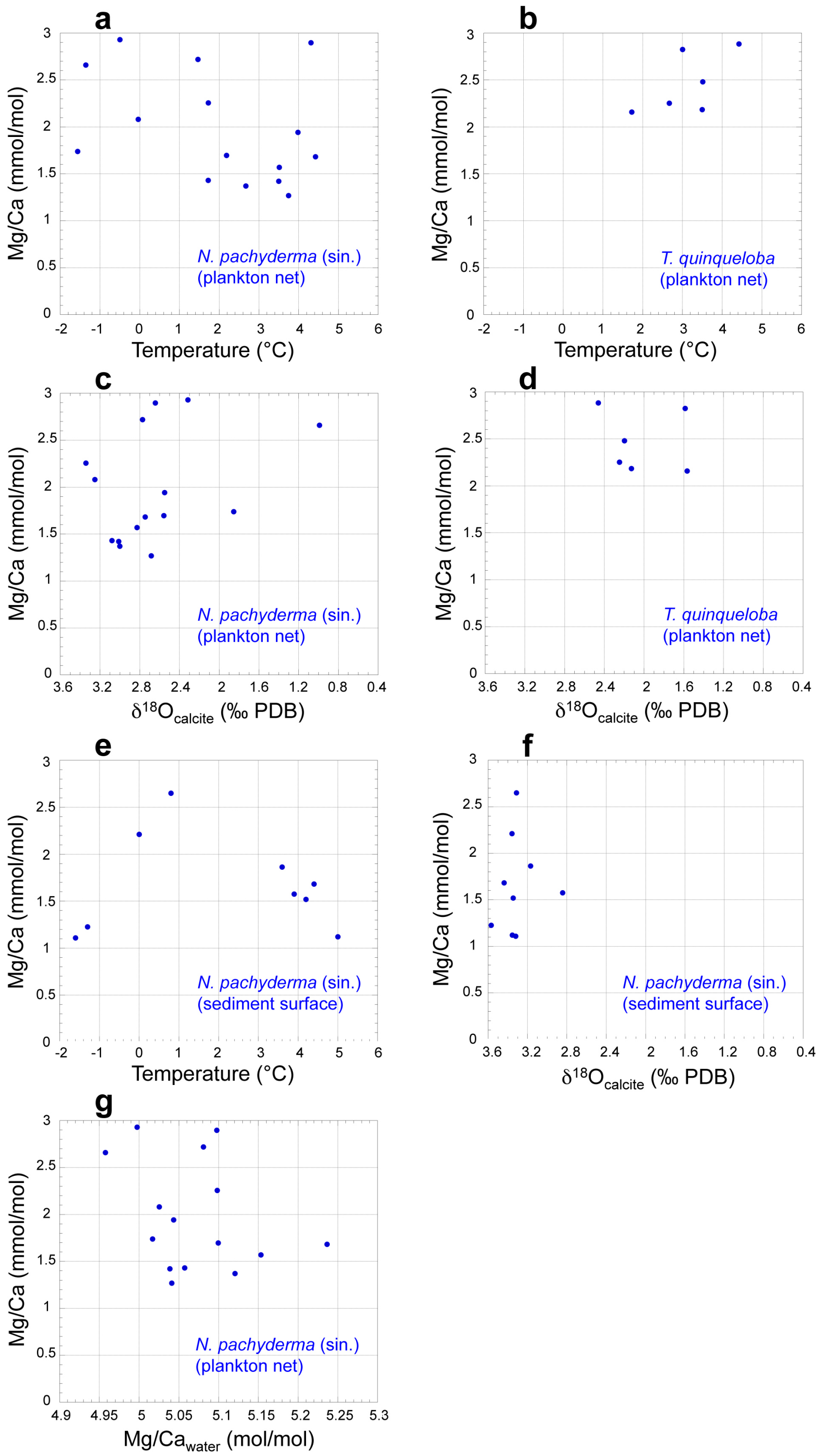

Figure 4.3. (a) Mg/Ca ratios measured in net-sampled *N. pachyderma* (sin.) versus *in situ* water temperatures in the collection depth. **(b)** Mg/Ca ratios measured in net-sampled *T. quinqueloba* versus *in situ* water temperatures in the collection depth. **(c)** Mg/Ca ratios versus $\delta^{18}O$ measured in net-sampled *N. pachyderma* (sin.). **(d)** Mg/Ca ratios versus $\delta^{18}O$ measured in net-sampled *T. quinqueloba*. **(e)** Mg/Ca ratios measured in *N. pachyderma* (sin.) sampled from the sediment surface versus water temperatures in the assumed calcification depth. **(f)** Mg/Ca ratios versus $\delta^{18}O$ measured in *N. pachyderma* (sin.) sampled from the sediment surface. **(g)** Mg/Ca ratios of net-sampled *N. pachyderma* (sin.) versus Mg/Ca ratios of the water in the collection depth.

4.4. Discussion

4.4.1 Critical evaluation of the data points

N. pachyderma (sin.) builds its secondary calcite shell while descending to a certain depth level for reproduction (Kohfeld et al., 1996; Simstich et al., 2003; Schiebel & Hemleben, 2005). Thus, the tests collected close to the sea surface are most probably incompletely calcified. Secondary crust and generally final chambers of foraminifera have been shown to yield significantly lower Mg/Ca ratios compared to the earlier formed parts of the shells (Anand & Elderfield, 2005; Sadekov et al., 2005; Marr et al., 2011). Tests without secondary crust have therefore – independent of the water temperature – increased Mg/Ca ratios compared to individuals that completed their life cycle before collection. For this reason we decided to omit the data points belonging to samples collected in the 0-50 m depth interval when discussing the results. Additionally, we rejected from further consideration the sample from the depth interval 50-100 m at station 71 that yields – similar to the above-mentioned samples from the 0-50 m depth interval at stations 35 and 71 – anomalous $\delta^{18}O_{calcite}$ value (Fig 4.3c; Pados et al., 2014).

4.2 Difference between species

In general, we found in the symbiont-bearing subpolar *T. quinqueloba* significantly higher Mg/Ca ratios than in the symbiont-barren polar *N. pachyderma* (sin.) taken from the same samples (Table 4.3). Jorgensen et al. (1985) measured increasing pH at the shell surface of symbiotic foraminifera as a result of photosynthesis. However, shell Mg/Ca and $[CO_3^{2-}]$ has been shown to have an inverse relationship (Russell et al., 2004). Thus, if the variation in pH in the calcifying microenvironment would alter the Mg/Ca ratio of the symbiont-bearing tests, we would expect to find lower Mg/Ca ratios in *T. quinqueloba* than in *N. pachyderma* (sin.), which is not the case. The offset is more likely related to two other factors. First, *N. pachyderma* (sin.) goes through a secondary calcification, unlike *T. quinqueloba* (Hemleben, 1989; Kohfeld et al., 1996; Simstich et al., 2003; Schiebel & Hemleben, 2005), and as mentioned above, the presence of secondary crust decreases the Mg/Ca ratio measured in the whole shell (Anand & Elderfield, 2005; Sadekov et al., 2005). Second, *T. quinqueloba* has been assumed to have a shallower calcification depth than *N. pachyderma* (sin.) (Be, 1977). Simstich et al. (2003) reported on significantly different depth habitats in these two species in the Fram Strait, while Carstens et al. (1997) and Pados & Spielhagen (2014) found in the same area only slight differences between *T. quinqueloba* and *N. pachyderma* (sin.). However, along our sampled transect in the upper 150 m of the water column alone, a temperature decrease of up to 5°C was measured with increasing depth (Fig. 4.2), which could result in recording colder temperatures by the slightly deeper dwelling *N. pachyderma* (sin.). The observed difference in shell Mg/Ca between the two species may thus reflect a combined effect of secondary calcification and different calcification depths.

Table 4.3. Mg/Ca ratios measured in *N. pachyderma* (sin.) (*N. p.* (sin.)) and in *T. quinqueloba* (*T. q.*) from both sediment surface and water column in the sampled depth intervals that contained enough specimens for the analysis. Red numbers indicate anomalous values that were obtained probably due to unsuccesful measurements and were therefore excluded from the study.

		Mg/Ca (mmol/mol)	
Station	Depth (m)	*N. p.* (sin.)	*T. q.*
PS78-19	0-50		
	50-100	1.268	
	100-200		
	200-300		
	300-500		
	sed. surface	1.864	4.377
PS78-25	0-50		
	50-100	1.942	
	100-200		
	200-300		
	300-500	1.697	
	sed. surface	1.575	
PS78-35	0-50	2.930	
	50-100	1.431	
	100-200	1.371	2.253
	200-300		
	300-500		
	sed. surface	2.212	
PS78-39	0-50	4.730	
	50-100	2.081	
	100-200	2.256	2.159
	200-300		
	300-500		
	sed. surface	2.650	2.236
PS78-44	0-50	3.617	
	50-100	2.897	
	100-200	1.422	2.185
	200-300		
	300-500		
	sed. surface	1.519	2.860
PS78-54	0-50	2.720	
	50-100	1.682	2.884
	100-200	1.570	2.481
	200-300		2.825
	300-500		
	sed. surface	1.68	1.622
PS78-71	0-50	2.659	
	50-100	1.739	
	100-200		
	200-300		
	300-500		
	sed. surface	1.109	
PS78-75	0-50		
	50-100		
	100-200		
	200-300		
	300-500		
	sed. surface	1.226	
PS78-87	0-50		
	50-100		
	100-200		
	200-300		
	300-500		
	sed. surface	1.120	1.841

4.4.3 Controls on Mg/Ca ratios in planktic foraminiferal calcite in the Fram Strait

It has been shown that magnesium uptake during inorganic precipitation of calcite is substantially different from biologically mediated partitioning in foraminiferal shells (Nürnberg et al., 1996). It is therefore assumed that physiological processes are responsible for regulating the Mg concentrations in foraminifera. These processes seem to be predominantly controlled by temperature (Nürnberg et al., 1996; Mashiotta et al., 1999; Elderfield & Ganssen, 2000; Lea et al., 2000). However, the relation between temperature and Mg/Ca content of the shells appears to vary among species, requiring species (or genus)-specific Mg/Ca-temperature calibration equations (Lea et al., 1999; Anand et al., 2003). Elderfield & Ganssen (2000) published a Mg/Ca-temperature calibration equation for *N. pachyderma* (sin.) and (dex.) using core-top samples from the North Atlantic Ocean:

$$\mathrm{Mg/Ca}=0.50*e^{(0.10*\mathrm{Temperature})}. \qquad (1)$$

Analysing cultured *N. pachyderma* (dex.) von Langen et al. (2005) developed a similar equation:

$$\mathrm{Mg/Ca}=0.51*e^{(0.10*\mathrm{Temperature})}. \qquad (2)$$

However, these equations do not describe the relationship between Mg/Ca and water temperatures in our samples. Even though *N. pachyderma* (sin.) and *T. quinqueloba* from both plankton tow and sediment surface samples in general show increasing Mg/Ca ratios with increasing temperature, they have much higher Mg/Ca ratio than what the equations yield with respective water temperatures (Fig. 4.3b and 4.4). Our results coincide with those of Meland et al. (2006), Kozdon et al. (2009) and Jonkers et al. (2013) who also observed in sediment surface and sediment trap samples from the Nordic Seas Mg/Ca ratios in *N. pachyderma* (sin.) higher than expected from the above-mentioned equations.

For the Norwegian Sea Kozdon et al. (2009) recommended a modified, linear Mg/Ca-temperature relationship to refine the "cold end" branch of the existing calibrations for *N. pachyderma* (sin.):

$$\text{Mg/Ca}=0.13*\text{Temperature}+0.35. \qquad (3)$$

Jonkers et al. (2013) adjusted the calibration equation of Elderfield & Ganssen (2000) according to their results from the Irminger Sea:

$$\text{Mg/Ca}=0.89*e^{(0.08*\text{Temperature})}. \qquad (4)$$

Nevertheless, none of the above-mentioned equations fits our data set either (Fig. 4.4), they all lead to an overestimation of *in situ* water temperatures. Hypothetically assuming that the equations by Elderfield & Ganssen (2000) and von Langen et al. (2005) reflect the calcification temperatures of *N. pachyderma* (sin.), our measured Mg/Ca values would yield calcification temperatures higher than 8°C, which is very unlikely in this region. Applying the linear correlation introduced by Kozdon et al. (2009) we would get slightly lower calcification temperatures. Calculating hypothetical Mg/Ca ratios of foraminiferal shells for the measured seawater temperatures according to the equations, we get Mg/Ca ratios lower than 0.8 mmol/mol.

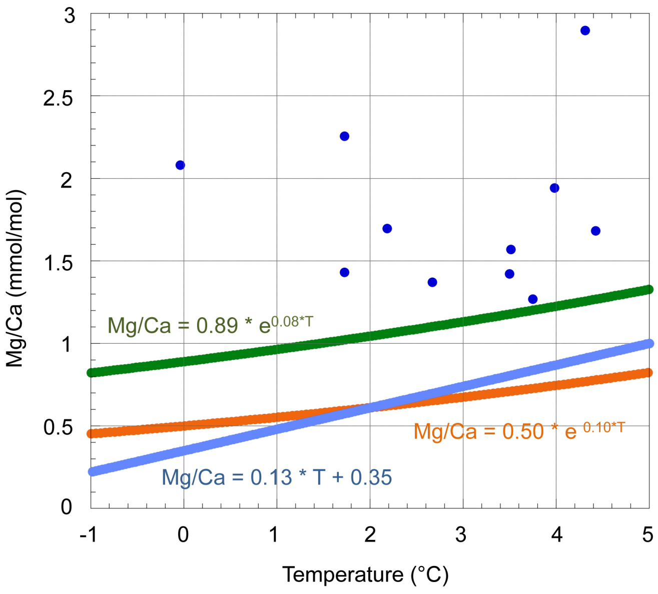

Figure 4.4. Mg/Ca ratios of *N. pachyderma* (sin.) from the water column compared to existing Mg/Ca calibrations of Elderfield & Ganssen (2000) (orange), Kozdon et al. (2009) (blue) and Jonkers et al. (2013) (green). Anomalous data points are omitted (see text).

Martínez-Botí et al. (2011) conducted an extensive study about Mg/Ca in plankton tow samples from various regions of the world ocean and also found elevated Mg/Ca ratios with respect to calibration predictions in distinct foraminifera species, just like van Raden et al. (2011) in the Mediterranean Sea and Friedrich et al. (2012) in the North Atlantic. Martínez-Botí et al. (2011) speculated that the specimens leading to overestimation of *in situ* temperatures were incompletely calcified, which resulted in increased Mg/Ca ratios compared to individuals that completed their life cycle before collection. Although the possible effect of incomplete secondary calcification cannot be discounted in our plankton tow samples, this hypothesis cannot explain the elevated Mg/Ca ratios in the tests from the sediment surface that were thick-walled and looked completely calcified. Lateral transport has been also suggested to resolve the enigmatic, anomalously high Mg/Ca ratios (Friedrich et al., 2012). However, as described by von Gyldenfeldt et al. (2000), mean transport distances of planktic foraminifera in the Fram Strait are short (ca. 25-50 km for *N. pachyderma* (sin.)) and the flow direction of warm and cold water masses is largely normal to the sampling transect, with only minor temperature changes on such distances. It thus seems unlikely that the majority of foraminifera calcified in water masses with temperatures significantly different from where they were caught. Moreover, if lateral transport would account for the anomalous Mg/Ca ratios, we would expect to find different Mg/Ca signatures in the core top and plankton tow samples, which is apparently not the case.

One possible explanation for the elevated Mg/Ca ratios found both in plankton and sediment surface samples and for the absence of a clear exponential or linear correlation between Mg/Ca values and water temperatures might be the influence of seawater chemistry on the Mg/Ca composition of the tests. This might mask the temperature-dependence, as already hypothesized by Meland et al. (2006). A relation of Mg/Ca ratios in foraminiferal shells to salinity was suggested from analyses of sediment samples (Nürnberg et al., 1996; Arbuszewski et al., 2010), sediment and plankton net samples (Mathien-Blard & Bassinot, 2009), and from culturing experiments (Lea et al., 1999). However, in our study the tests collected in the fresh waters of the EGC yield similarly high Mg/Ca values as the shells from saline Atlantic waters. Therefore salinity is unlikely to be of major influence here, just like in the laboratory experiments of Hönisch et al. (2013). Another factor that might influence the Mg/Ca composition of foraminiferal shells – and thus mask the temperature-dependence – is the variability of the Mg/Ca ratio in the water column. The Mg/Ca ratio of seawater is considered to be constant and termed "conservative" (Broecker et al., 1982). Nonetheless, our measurements show that the Mg/Ca distribution of the water column in the Fram Strait reflects quite well the distribution of the main water masses. Low ratios (<5 mol/mol) mark the cold and fresh waters of the Polar Layer in the west and that of the Mixed Polar Layer below the Atlantic Layer. In general, the Mg/Ca composition of the water apparently follows the temperature variability. In contrast, shell Mg/Ca of *N. pachyderma* (sin.) seems to be less related to the Mg/Ca ratios measured in the water column than to *in situ* temperatures (Fig. 4.3g). Thus we find it unlikely that Mg/Ca ratios of seawater may explain the discrepancies found.

To explain the elevated Mg/Ca ratios in our samples, we speculate that they might result from the effect of carbonate ion concentration ($[CO_3^{2-}]$) of the seawater on the tests′ biochemical composition. Lea et al. (1999) and Russell et al. (2004) reported on a negative correlation between Mg/Ca of foraminiferal tests and $[CO_3^{2-}]$ under constant temperatures in laboratory experiments. Moreover, in *N. pachyderm*a (sin.) from high latitude areas a possible increased sensitivity to $[CO_3^{2-}]$ changes has been assumed (Jonkers et al., 2013). The majority of our samples was taken from cold Polar water masses, characterized by low $[CO_3^{2-}]$ (Chierchi & Franson, 2009). This may explain the elevated Mg/Ca ratios found in the tests compared to the studies of Kozdon et al. (2009) and Jonkers et al. (2013), conducted in areas south of our study site. However, more investigations need to be done to evaluate the potential effect of *in situ* $[CO_3^{2-}]$ on Mg/Ca ratios of planktic foraminifera in regions with cold surface and subsurface waters.

4.5. Conclusions

Mg/Ca ratios in planktic foraminifera collected from the water column in the Fram Strait reveal a very weak correlation with *in situ* water temperatures. The two dominating species show a clear offset in Mg/Ca ratios: *N. pachyderma* (sin.) have significantly lower Mg/Ca ratios than *T. quinqueloba*, which might be explained by the combined effect of secondary calcite crust and different calcification depths.

N. pachyderma (sin.) both from the water column and from the sediment surface have elevated Mg/Ca ratios with respect to predictions from available equations involving water temperatures. Incompletely calcified specimens leading to an overestimation of water temperatures or lateral transport are unlikely the reasons explaining the discrepancies found. One possible explanation is that foraminiferal Mg/Ca ratios in the area may be significantly influenced by seawater chemistry that might mask the temperature dependence of this proxy. However, salinity and Mg/Ca ratios of the water column are probably not a major factor inducing the observed anomalous values. The elevated Mg/Ca ratios of *N. pachyderma* (sin.) in the Fram Strait are more likely related to the influence of sewater carbonate chemistry. Further *in situ*-studies are needed to confirm this hypothesis.

Acknowledgements

We wish to express our appreciation to the captain and crew of RV Polarstern and the Alfred Wegener Institute, Helmholtz Centre for Polar and Marine Research in Bremerhaven for support during the cruise ARK XXVI/1. Thanks are also due to Maciej Telesinski for collecting the sediment surface samples for this study and to Dr. Angelina Kraft for assisting in the deployments of the MultiNet sampler. We thank Dr. Agnieszka Beszczynska-Möller and Andreas Wisotzki for making available the CTD data for the study, Dr. Hartmut Schulz for providing the low temperature vacuum asher at the University of Tübingen and Karen Bremer for the Mg/Ca measurements. This research was conducted with funding of the European Community's 7th Framework Programme FP7 2007/2013, Marie-Curie Actions, under Grant Agreement No. 238111 – CASE ITN. R.F.Spielhagen acknowledges the financial support through the "Akademienprogramm" by the Academy of Sciences, Humanities, and Literature Mainz.

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Acknowledgements

First, I would like to thank my supervisor, Robert F. Spielhagen for giving me the opportunity to work on this interesting (and from my previous research so different) topic and to participate in an unforgettable research cruise; and for his supervision during the whole time of my PhD studies; Dorothea Bauch for her valuable advices and help during manuscript preparation-periods, and the whole "arctic-building 4-group" for the rewarding time I spent there.

Furthermore, I would like to thank my official supervisor Prof. Martin Frank and the whole Paleoceanography Department for the support.

Special thanks go to my friends who have been always there for me and without whose I would have never had such an amazing time in Kiel. You have listened to all my crazy/happy/sad stories, you comforted me when I was down, you have trusted me to share your secrets with me, you helped me whenever I needed help, you invited me to your wedding/birthday/whatever parties, you celebrated my 30th birthday with me, you have watched movies with me, you have cooked delicious dishes for me. You have simply made me feel good, even in this peculiar city. Thank you Zeynep, Anna, Mario, Idi, Gizem, Arne, Salman, Robert, Georgi, Daniel, Paula, Lukas I., Lukas II., Fernando, Nicolas.... and everybody else who I have forgotten to mention now. WE ARE THE INFINITY.

To my "not-Kiel-friends": Patri for the fun that we had in San Francisco, Agnieszka for being the best roommate ever on Polarstern, Daniele for the continuous Skype-support, and Krisztian for listening every time when I needed to be listened to.

Alba, my dear, thank you so much for the crazy times, for cruising with me through Europe. You are the best.

And Felice... Thank you for proving that sometimes even silly daydreams can come true and make the life sweeter.

Last but not least I thank very-very much my parents and my grandmother for supporting me without hesitation all the time and for making it possible for me to achieve my dreams. I could not wish a better family.