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Observations of water masses and circulation in the Eurasian Basin of the Arctic Ocean from the 1990s to the late 2000s

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Abstract

The circulation and water mass properties in the Eurasian Basin are discussed based on a review of previous research and an examination of observations made in recent years within, or parallel to, DAMOCLES (Developing Arctic Modelling and Observational Capabilities for Long-term Environmental Studies). The discussion is strongly biased towards observations made from icebreakers and particularly from the cruise with R/V *Polarstern* 2007 during the International Polar Year (IPY). Focus is on the Barents Sea inflow branch and its mixing with the Fram Strait inflow branch. It is proposed that the Barents Sea branch contributes not just intermediate water but also most of the Atlantic layer that is found in the Amundsen Basin and also in the Makarov and Canada basins. Only occasionally would high temperature pulses originating from the Fram Strait branch penetrate along the Laptev Sea slope across the Gakkel Ridge into the Amundsen Basin. Interactions between the Barents Sea and the Fram Strait branches lead to formation of intrusive layers, in the Atlantic layer and in the intermediate waters. The intrusion characteristics found downstream north of the Laptev Sea are similar to those observed in the Northern Nansen Basin and over the Gakkel Ridge, implying a flow from the Laptev Sea towards Fram Strait. The formation mechanisms for the intrusions at the continental slope, or in the interior of the basins if they are reformed there, have not been identified. The temperature of the deep water of the Eurasian Basin has increased in the last 10 yr rather more than expected from geothermal heating. That geothermal heating does influence the deep water column was obvious from 2007 *Polarstern* observations made close to a hydrothermal vent in the Gakkel Ridge, where the temperature minimum usually found above the 600–800 m thick homogenous bottom layer was absent. However, heat entrained from the Atlantic water into descending boundary plumes may also contribute to the warming of the deeper layers.

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1 Introduction

Recent studies of the Arctic Ocean circulation have focussed on three discoveries, or rediscoveries, occurring in the 1980s and 1990s. (1) The realisation that the inflow of Atlantic water from the Norwegian Sea over the Barents Sea to the Arctic Ocean is of similar magnitude as the inflow of Atlantic water taking place in Fram Strait (Rudels, 1987; Blindheim, 1989). (2) The shelf contribution to the formation of the halocline does not just involve the creation of dense water by brine rejection that eventually penetrates into the deep basin water column but also the outflow of low salinity shelf water that overruns the winter mixed layer in the Nansen Basin, transforming it into a halocline water mass (Rudels et al., 1996). (3) The inflow of anomalously warm Atlantic water to the Arctic Ocean in the late 1980s and early 1990s (Quadfasel et al., 1991), and the relocation of the upper low salinity surface water from a large part of the Amundsen Basin into the Makarov Basin and even into the Canada Basin, and the shift of the surface front between the Atlantic and the Pacific derived waters from the Lomonosov Ridge to the Mendeleyev Ridge (Carmack et al., 1995; Morison et al., 1998; Steele and Boyd, 1998). The present study reviews how these findings have guided recent research and in particular the oceanography work from icebreakers conducted within DAMOCLES. It also describes how observations made the late part of the 2000s have improved our understanding of the Arctic Ocean processes and circulation.

2 The Barents Sea inflow branch

2.1 The inflow over the Barents Sea

The assumption that much, if not most, of the water that enters the Arctic Ocean from the Nordic Seas passes through the Barents Sea is not new. It was held by Nansen in the beginning of the 20th century (Nansen, 1906) and it was to a large extent accepted by Russian researchers, who commonly considered the inflows through Fram Strait and the Barents Sea to be of comparable magnitude (see e.g. Nikiferov and Sphaiker, 1980). The sudden departure from this view occurred, when the results from the first direct current measurements in the West Spitsbergen Current at 79° N were reported and indicated a northward transport of almost 8 Sv (1 Sv = 10^6 m^3 s^{-1}) (Aagaard et al., 1973). This put focus on the Fram Strait exchanges. In the mass, heat and salt budgets proposed by Aagaard and Greisman (1975) the exchanges through Fram Strait were assumed to balance with a northward flow of 7.1 Sv of Atlantic water and a southward flow of 1.8 Sv of low salinity Polar water and 5.3 Sv of returning, cooled and modified Atlantic water (Arctic Atlantic water – AAW). By contrast the inflow over the Barents Sea was estimated at 0.7 Sv, 10 % of the northward flow in the West Spitsbergen Current. The Fram Strait transport estimate given by Aagaard and Greisman did not include any deep-water exchanges.

The importance of the Barents Sea inflow to the Arctic Ocean was reintroduced in the late 1980s, when Rudels (1987) estimated the transport over the Barents Sea by formulating a heat budget for the Atlantic water and for the Norwegian Coastal Current water entering from the Norwegian Sea. These waters mainly leave the Barents Sea through the passage between Novaya Zemlya and Franz Josef Land, continuing down the St. Anna Trough, but also flow north in the Victoria Channel and the coastal water may pass through the Kara Gate south of Novaya Zemlya. However, Rudels assumed that a large fraction, two thirds, of the Atlantic water recirculated in the Hopen Deep and returned as cooler water in the Bear Island Channel to the Norwegian Sea. The inflow to the Arctic Ocean was estimated at 1.2 Sv of which the Norwegian Coastal Current supplied 0.8 Sv. Blindheim (1989) obtained from 3 weeks of direct current measurements a similar inflow from the Norwegian Sea but found that the recirculation was considerably smaller than Rudels (1987) had assumed and estimated a net inflow of 1.8–1.9 Sv. This included the contribution from the Norwegian Coastal Current, which was not given separately by Blindheim but was later estimated by Aagaard and Carmack (1989) based on Blindheim's work.

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Current measurements made continuously since 1997 between Nordkapp and Bjørnøya, the Barents Sea Opening, have largely confirmed Blindheim's estimate, but also revealed large seasonal and monthly variability of the transports (Ingvaldsen et al., 2004a,b). Recent direct velocity observations of the Norwegian Coastal Current have, however, shown that this transport is ~ 1.8 Sv, considerably larger than previously estimated (Skagseth et al., 2011). Only twice, 1991–1992 and 2008–2009 has a current meter array been deployed and recovered in the passage between Novaya Zemlya and Franz Josef Land, the expected main pathway for the Barents Sea inflow to the Arctic Ocean. The observed transports (~ 2 Sv) for the 1991–1992 array (Loeng et al., 1993; Schauer et al., 2002a) agree well with the inflow at the Barents Sea Opening found by Blindheim (1989). The data from the 2008–2009 array are not yet available.

2.2 The impact on the Arctic Ocean water column

The impact of the Barents Sea branch on the Arctic Ocean water column was recognised by Rudels et al. (1994) from the hydrographic observations in the Nansen and Amundsen basins obtained on the Arctic Ocean-91 expedition with IB *Oden*. The presence of inversions and intrusions in the water column, in the warm Atlantic layer as well as in the intermediate layers below, was interpreted as results of interactions between two inflow branches, a warm and saline Fram Strait inflow branch and a cooler and fresher Barents Sea inflow branch, taking place as they meet on the continental slope north of the Kara Sea. The *Oden* stations were located away from the slope and west of the area, where the branches were expected to meet. However, the observations indicated that the fraction of Barents Sea branch water in the water column increased towards the Lomonosov Ridge. This implied that part of the Barents Sea branch water, and most likely also much of the Fram Strait branch water encountered in the interior of the Eurasian Basin away from the continental slope, flows towards Fram Strait. Such a flow pattern requires at least a partial turning of the two branches north of the Laptev Sea before they reach the Lomonosov Ridge, causing much of the Atlantic inflow to remain in the Eurasian Basin.

Subsequent observations north of the Laptev Sea confirmed the presence of less saline Barents Sea branch water at the slope (Schauer et al., 1997), and strong interleaving between the two branches were observed north of Severnaya Zemlya on the shoreward side of the Fram Strait branch in 1995 (Rudels et al., 2000a). Farther offshore the temperature maximum of the Fram Strait branch was unaffected by the intrusions. North of the Laptev Sea the temperature of the Atlantic layer was considerably lower than north of the Barents Sea (Schauer et al., 1997) and north of the eastern Kara Sea (Rudels et al., 2000a). However, large differences were observed between profiles obtained at stations with multiple casts, indicating large spatial (drift of the ship), but perhaps also temporal, variability as the water column advects along the slope (Rudels et al., 2000a).

Cold, low salinity lenses were observed at intermediate depth in the Makarov Basin in 1994 (e.g. Swift et al., 1997), and water with similar characteristics was found at the sill (1600 m) of the Lomonosov Ridge close to the Siberian continental slope (Rudels et al., 2000a). An inflow of Barents Sea branch water across the Lomonosov Ridge would largely explain the presence of these cold, low salinity lenses in the Amerasian Basin (Swift et al., 1997). Barents Sea branch water has lately been almost exclusively associated with this low salinity, cold, intermediate water mass. This is a simplification. The Barents Sea branch comprises a large range of densities and contributes significantly to the warm Atlantic core in the Amerasian Basin, usually associated with the Fram Strait branch, as well as partly supplies water to the lower halocline (Rudels et al., 2004; Aksenov et al., 2011).

3 Changes in the Fram Strait inflow branch

The Atlantic water entering through Fram Strait and the variability of the Fram Strait inflow branch have received considerable attention. Fram Strait is the only deep passage connecting the Arctic Ocean with the Nordic Seas and across the Greenland-Scotland Ridge to the North Atlantic. It has long been considered the most important passage for oceanic heat to the Arctic Ocean, and it is the main exit for the Arctic sea ice and the only opening allowing exchanges of intermediate and deep waters. The estimates of the Atlantic water transport into the Arctic Ocean have varied from 2–3 Sv to the high value of 7 Sv in the 1970s (Aagaard and Greisman, 1975) and then again to a low value of 1 Sv in the late 1980s (e.g. Aagaard and Carmack, 1989). After the regular monitoring of the exchanges in Fram Strait began with VEINS (Variability of Exchanges in the Northern Seas) in 1997 the northward transport of Atlantic water in the West Spitsbergen Current has been found to range between 3–5 Sv, with large variations but no decisive trend (Schauer et al., 2004; Rudels et al., 2008; Schauer et al., 2008; Beszczynska-Möller et al., 2012). An increased northward flow in the West Spitsbergen Current is often compensated by larger recirculation in the strait and warmer Atlantic water in the East Greenland Current. Part of the inflow thus recirculates within the strait and about 2–3 Sv contribute to the eastward boundary current along the Eurasian continental slope (Schauer et al., 2008).

In 1990 anomalously warm Atlantic water was observed at the continental slope of the Eurasian Basin (Quadfasel et al., 1991). It was the first observation of a warmer, and likely also stronger, inflow taking place in the late 1980s and early 1990s. The warmer water could be followed across the Lomonosov Ridge into the Makarov and Canada basins (Carmack et al., 1995; McLaughlin et al., 1996). The advection of the warmer Atlantic water then offered an opportunity to determine the pathways of the Atlantic water in the Arctic Ocean. Anomalously high temperatures were observed in the Atlantic layer west of the Chukchi Borderland in 1998 (Shimada et al., 2004), but its subsequent movement around the Chukchi Borderland or between the Chukchi Borderland and the continental slope appeared to be much slower than the passage from Fram Strait to the Chukchi Sea slope. It is also not obvious which route, around the Chukchi Borderland or along the slope, is the most important one. However, low Atlantic water temperatures have been reported from the Alpha Ridge (Falkner et al., 2005), and the temperature section across the Amerasian Basin taken by IB *Oden* on the Beringia expedition 2005 showed the lowest Atlantic layer temperatures at the

Alpha Ridge. Similar low Atlantic layer temperatures were also observed over the Alpha Ridge on the *Polarstern* cruise in 2007 (Fig. 1). This suggests that older, and colder, Atlantic water has been displaced northward from the southern Canada Basin and is now entering the northern Canada Basin from the North American continental slope along the Alpha Ridge.

The temperature of the Atlantic water on the Makarov Basin side of the Lomonosov Ridge had increased by 2004 (Kikuchi et al., 2005), and the observations from *Oden* 2005 indicated that the temperatures at the Makarov Basin side were as high, or even higher, than on the Amundsen Basin side (see Figs. 1 and 13). This implies first, that a part of the warm Atlantic water pulse had entered the Makarov Basin at the Mendeleyev Ridge, circulated around the basin in about 10 yr and was now returning towards Siberia along the Lomonosov Ridge, and second, the Atlantic water on the Amundsen Basin side had been replaced by colder Atlantic water, presumably from a later, colder inflow event. A reduction in the Atlantic water temperatures in Fram Strait was observed between 1984 and 1997 (Rudels et al., 2000b), and the arrival of colder Atlantic water at the NABOS (Nansen-Amundsen Basin Observational Studies) stations and moorings north of the Laptev Sea was detected in 2003 (Dmitrenko et al., 2005).

Since 1997 the VEINS/ASOF/DAMOCLES mooring array (ASOF – Arctic-Subarctic Ocean Fluxes) in Fram Strait has revealed several pulses of warmer Atlantic water passing into the Arctic Ocean, and there has been a general trend towards higher temperature, having a maximum in 2006 and slightly decreasing thereafter (Hughes et al., 2011). The higher temperatures observed at the NABOS moorings north of the Laptev Sea after 2003 are likely related to the arrival of these pulses, and the time evolution of the Atlantic water at different positions has been documented and summarised by Dmitrenko et al. (2005) and Polyakov et al. (2005, 2011).

The upstream sources of the warm Atlantic inflow pulses have also been investigated. Much of the higher temperature relates to higher air temperature and less cooling in the Norwegian Sea connected with the positive NAO state prevailing in the late

1980s and early 1990s (Karcher et al., 2003). However, some changes can be traced farther south (Holliday et al., 2008). It has been proposed that the strength of the northward flow of Atlantic water across the Greenland-Scotland Ridge is related to the wind stress curl at 55° N (Orvig and Skagseth, 2003) and it has also been noticed that the subpolar gyre south of the Greenland-Scotland Ridge has a two-mode structure. Either a larger part of the water is brought northward into Nordic Seas, or more is kept in the gyre and carried westward into the Irminger Sea and ultimately into the Labrador Sea (Hátún et al., 2005). Both these processes, related to the atmospheric circulation and probably connected, will influence the upstream conditions of the Atlantic water eventually entering the Arctic Ocean.

4 Variability in the Eurasian Basin and interactions with shelf outflows

The largest and most intense transformations of the Atlantic inflow occur in the Nansen Basin and on the shelves of the adjacent Barents and Kara seas. In this area the heat of the Atlantic water has the largest possibility to become mixed into the surface layer and reach the sea ice and the atmosphere. The SPACE (Synoptic Pan-Arctic Climate and Environmental Study) cruise with R/V *Polarstern* in 2007 extended over much of this area, comprising 4 sections reaching from the shelf into the deep basin (Fig. 2a). The sections provide enough spatial resolution to determine the changes in the water mass characteristics along and across the Nansen Basin and to form the basis for a study of the transformations and changes occurring in the basin; in particular, changes in the shelf-basin interactions.

Furthermore, the SPACE sections are reoccupations of previous sections worked from the shelf into the basin and thus provide a temporal dimension of the changes over the last 10–15 yr (Fig. 2b). The two main SPACE sections extended over two thirds of the Arctic Ocean, reaching from the Eurasian shelf to the Alpha Ridge crossing not only the Nansen Basin but also the Amundsen and Makarov basins as well as the Gakkel and Lomonosov ridges (Fig. 2a). Of the four sections extending from the shelf into the

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Nansen Basin the first ran along 30° E, east of Svalbard, and on the second section the ship returned from the basin along 50° E towards Franz Josef Land. The third, main, section was taken from the eastern Kara Sea along 90° E across the Nansen and Amundsen basins into the Makarov Basin, and the final section (4) was along the Gakkel Ridge reaching the Laptev Sea at 126° E. An examination of the shelf-basin exchanges thus naturally starts with section 1.

The section along 30° E extended farthest onto the shelf, and it almost captured the Barents Sea inflow branch close to its entrance at the Barents Sea Opening. The most significant features are seen in the upper 400 m (Fig. 3). On the Northern Barents Sea shelf the cold remnant of the winter mixed layer dominates between 50 m and 100 m It is almost as cold but slightly more saline and denser than the winter mixed layer in the Nansen Basin farther north. The two distinct winter mixed layers derive from the interactions between sea ice and the two Atlantic inflows. Ice melts on top of the warmer Atlantic water and the resulting less saline surface waters are homogenised by cooling and later by ice formation in winter (Rudels et al., 1996, 2004). The higher salinity in the Barents Sea winter mixed layer has been suggested due to the fact that the Atlantic water initially encountering sea ice is colder in the Barents Sea than in the Nansen Basin north of Svalbard and then a smaller fraction of the heat loss goes to ice melt (Rudels et al., 2004).

No connection is seen between the two low salinity upper layers at the shelf break in the northern Barents Sea, where they are separated by warmer and more saline water associated with upwelling of the Atlantic core at the continental slope. However, at section 3 at 90° E in the Kara Sea east of the St. Anna Trough the two mixed layers are observed flowing side by side (see θS diagram in Fig. 4). This suggests that transformations of the upper part of the Barents Sea inflow branch occur also farther east in the Barents Sea and in the Kara Sea, and that this part of the branch primarily enters the Nansen Basin via the Kara Sea. Because the Barents Sea branch winter mixed layer is denser than that found in the Nansen Basin, it interacts with, and cools,

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the thermocline and the upper part of the Atlantic core of the Fram Strait branch as the two branches meet and mix north of the Kara Sea (Fig. 4).

The Atlantic water approaches the sea surface at, or close to, the shelf break north of the Barents Sea, which suggests that this region might be one, where a significant heat loss from the Atlantic water to the mixed layer, to sea ice and the atmosphere occurs. This is corroborated by the fact that no upper temperature minimum is found between the surface water and the Atlantic core. This upper minimum is normally a remnant of the local winter convection, and its existence almost everywhere in the Arctic Ocean indicates that the vertical heat transfer in summer is small (Fig. 5). In winter the homogenisation of the upper layer reaches the thermocline in the Nansen Basin but not elsewhere in the Arctic Ocean, and it is possible that upward vertical heat flux occurs in winter in the Nansen Basin. The absence of an upper temperature minimum at the shelf break shows that heat from the Atlantic water reaches the seasonally heated surface layer in summer, and this heat flux should be larger in winter when the stability is weaker. Farther to the east the Barents Sea inflow branch deflects the Fram Strait branch from the slope, and the vertical heat flux at the shelf break becomes smaller and likely less important.

At the southern end of the section the Atlantic water over the Central Bank in the Barents Sea displays high salinity and density and fairly high ($> 0\,^{\circ}$C) temperature. The bottom density over the bank is as high as or higher than that of the Arctic Ocean deep waters. Whether this high density would survive an advection towards either the Arctic Ocean via the St. Anna Trough or back to the Norwegian Sea in the Bear Island Channel, allowing for a ventilation of the deeper layers in the basins, is another matter. What is significant is that in the 1980s highest densities over the Central Bank were associated with ice formation and the temperature at the bank was close to the freezing point, while in years with no ice and warmer, more saline water over the bank the water column exhibited lower densities, also in comparison with 2007 (Quadfasel et al., 1992). This indicates a change, or at least a variation, of the conditions in the Barents Sea that could be significant for the deep-water ventilation.

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In the northern, deep part of the section there were differences between the intermediate waters close to the continental slope and those farther to the north (Fig. 6). A sharp decrease in salinity and in temperature of the Atlantic core was observed between the boundary current and the northern part of the Nansen Basin and large and regular intrusions were found on the northern stations. At the slope a weak salinity minimum is seen at 1500 m indicating the presence of Arctic Intermediate Water (AIW) or Nordic Seas Deep Water (NSD) from the Greenland Sea. This minimum is not found on the northern stations. Instead a salinity minimum is observed around 900 m. The most probable source for this minimum is the Barents Sea inflow branch (Fig. 6). This suggests that the water column in the northern part of the section either partly derives from the Barents Sea branch or has been in close, mixing, contact with the St. Anna inflow.

The second section, extending from the Nansen Basin towards Franz Josef Land, showed a similar warm Atlantic core as section 1, but the section evidently did not extend far into the Basin to observe the temperature and the salinity decrease seen in section 1. However, close to the slope the profiles became more rugged and irregular and cold intrusions appeared (Fig. 7). This resulted in a lowering of the salinity and temperature of the Atlantic core. The origin of the cold intrusions were outflows of cold, less saline water from the Barents Sea, which would mainly occur in the Victoria Channel between Victoria Island and Franz Josef Land. Cold, low salinity water was observed below the Atlantic layer at the slope north of Franz Josef Land in 1980 (Rudels, 1986) and in 1993 (Schauer et al., 1997) and these observations were taken as evidence that dense outflows from the Victoria Channel were taking place. The earlier observations showed one strong cold and fresh intrusion, and not the vertically more widespread interleaving observed in 2007 (Fig. 7). There was also colder, denser water present at the upper part of the slope that eventually might penetrate to larger depth (Fig. 8, blue station). The difference between the years could be due to ice formation and brine rejection in the polynya area around Franz Josef Land, which might have been more efficient at creating cold dense water in 2007 than on the other occasions. Some stations

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were also obtained in the western St. Anna Trough, unfortunately not capturing the main Barents Sea outflow but showing Fram Strait branch water entering the St. Anna Trough from the north and mixing with colder, less saline shelf water (Fig. 8).

The third section started in the eastern Kara Sea east of the St. Anna Trough, the major passage for the dense Barents Sea branch water entering the Arctic Ocean. The difference between the winter mixed layer in the Nansen Basin and in the Barents Sea branch has been discussed above and is shown in Fig. 3. The denser waters of the Barents Sea branch, which on the section are located above 400 m on the upper part of the slope, are dense enough to mix into the core of the Atlantic layer and to penetrate into the intermediate layers below. The densest part of the Barents Sea branch inflow appears to form a separate stream and sink directly from the St. Anna Trough into the intermediate layer creating a deep less saline intrusion in the boundary current (Fig. 8).

On the shelf three distinct dense contributions from the Barents Sea branch can be identified: a warm Atlantic core with temperatures slightly above 1 °C and salinity around 34.88, below the warm core a temperature minimum with $\theta \sim -1$ °C and salinity just above 34.8 is present, and at the bottom a dense part with temperature around −0.5 °C and salinity ∼ 34.88 is found. These characteristics are close to those observed in St. Anna Trough in 1996 (Schauer et al., 2002a), where the cold intermediate part was attached to the eastern flank of the trough, while the densest, slightly warmer water was found in the deepest part of the trough. A simple minded interpretation would then be that the densest part derives from the Barents Sea, probably from the shallow area west of Novaya Zemlya, while the colder contribution is formed on the northern banks of the Kara Sea east of the St. Anna Trough from where it drains to join the main Barents Sea outflow. The temperature maximum would be the remnant of the Atlantic water core entering at the Barents Sea opening (Fig. 9).

At section 3 the stream located at the shelf break has not yet begun to mix with the Fram Strait branch, and the front between the two branches is distinct and narrow. Only one station was showing large intrusions, especially in the warm core. The temperature and the salinity of the main part of the Fram Strait branch Atlantic core is similar to that found on sections 1 and 2, implying that no significant cooling has occurred between Svalbard and the eastern Kara Sea. Most of the intrusions and interleaving observed in the Fram Strait branch were here found above and at the temperature maximum, but above the salinity maximum (Fig. 9).

At section 4, along the Gakkel Ridge onto the Laptev Sea this had all changed. The temperatures and salinities of the temperature and salinity maxima of the two branches have become closer. The range is still fairly wide with a temperature difference of one degree and a salinity difference of 0.05 compared to two degrees and 0.1, respectively, at section 3 (Fig. 10). The Atlantic core displays distinct intrusions located primarily at and below the temperature maximum but above the salinity maximum, and the salinity minimum from the Barents Sea branch has spread from the slope into the Fram Strait branch.

The temperature reduction observed in the Fram Strait branch Atlantic core between the Kara Sea and the Laptev Sea is the largest one occurring along the Eurasian slope. This is more clearly seen by comparing the temperature and salinity sections taken across the slope at the different crossings (Fig. 11). The warm and saline Fram Strait branch is affected somewhat by mixing with Barents Sea branch water north of Franz Josef Land. North of the Kara Sea it is forced off the slope by the Barents Sea inflow branch east of the St. Anna Trough, but it still retains its high temperature. North of the Laptev Sea, really large changes can be seen in the characteristics of the Fram Strait branch. The Atlantic water core has become colder by 1 degree and its salinity been reduced by 0.05 and the core is displaced farther away from the slope.

One obvious and possible cause for the reduction in temperature is isopycnal mixing with the colder Barents Sea branch, and the intrusions found in the Atlantic layer indicate strong lateral mixing between the branches. Still a frontal zone remains at the slope between the Fram Strait branch and the Barents Sea branch. On the basin side of the Fram Strait branch, continuing section 4 along the Gakkel Ridge, the temperature and salinity of the Atlantic core again decrease and approach the values found for the Barents Sea branch at the slope (Fig. 12). Interleaving is present, and the maxima

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and minima of the different stations appear to line up in the θS curves and also in the profiles.

A similar structure is found at section 3 on the basin side of the warm Atlantic core (Fig. 13). Here the frontal structure is more distinct. Just beyond the warm Fram Strait branch the temperature and salinity decrease by more than 0.6 °C and 0.04, respectively, to values close to the maxima observed on the Laptev Sea section. Farther into the basin another front is present and the temperature and salinity drop to just above 1 °C and below 34.9, close to the values of the Barents Sea branch at its temperature maximum, and similar to what was observed at the Laptev Sea slope. These temperature and salinity characteristics then hold for the Atlantic layer in the entire Amundsen Basin up to the Lomonosov Ridge, where the temperature again increases slightly (see e.g. Fig. 15e below).

Interleaving structures are present at the basin side of the first front as well as an intermediate salinity minimum, less distinct over the Gakkel Ridge and in the Amundsen Basin in spite of lower salinity because of the lower temperature and salinity in the Atlantic layer. The lower salinity and the interleaving structures on the basin side of the Fram Strait branch indicate water with a significant, or even dominant, Barents Sea branch component. The temperature and salinity in the Atlantic layer are close to those of the Atlantic layer of the Barents Sea branch at the Kara Sea slope, but the salinity minimum is significantly reduced (Figs. 9 and 13). This is likely due to the higher density of the less saline water, which north of the Kara Sea was located high at the slope and had to descend, mixing with ambient Fram Strait branch water, before it reached its neutral density level (Fig. 11c and d).

This implies that the Barents Sea branch in general supplies most of the Atlantic water masses beyond the Laptev Sea and that the Fram Strait branch becomes markedly eroded by lateral mixing with Barents Sea branch water, initially in the boundary current along the Eurasian slope north of Severnaya Zemlya, but also by mixing on the basin side of boundary current. Here interleaving and large intrusions are seen in the Atlantic layer between the temperature maximum and the salinity maximum, whereas interleaving and intrusions with smaller vertical extent are observed below the salinity maximum and in the underlying salinity minimum. The transport of the Fram Strait branch water in the boundary current would then be all but extinguished north of the Laptev Sea. Beyond the Gakkel Ridge, in the Amundsen Basin and in the Amerasian Basin beyond the Lomonosov Ridge, the Atlantic inflow, except the Fram Strait contribution to the lower halocline, would mainly derive from the inflow over the Barents Sea (Fig. 14).

That the influence of the Atlantic inflow through Fram Strait would be limited to the Nansen Basin is surprising, considering that the estimate of Atlantic water entering the Arctic Ocean through Fram Strait, 2–4 Sv (Schauer et al., 2004, 2008; Beszczynska-Möller et al., 2012), is almost twice that entering the Barents Sea Opening (Skagseth et al., 2008). If most of the Fram Strait branch water hardly enters the Amundsen Basin, not to mention the Amerasian Basin, the situation observed in the Amundsen Basin and over the Gakkel Ridge in 2007 can be understood. Such close recirculation would explain the high temperatures of some of the Arctic Atlantic water returning towards Fram Strait. It would also explain the lower salinity and the strong interleaving found on section 3 and on the basin side of the boundary current on section 1. The lower salinity and the intrusions would be remnants from the initial mixing with the Barents Sea branch water that began at the slope north of Severnaya Zemlya.

This situation, observed in 2007, cannot always be present. The inflow of warm Atlantic water into the Makarov Basin observed in 1993 (Carmack et al., 1995), and the high temperatures found on the Arctic Ocean section in 1994 (Carmack et al., 1997; Swift et al., 1997) and the *Polarstern* section in 1996 (Schauer et al., 2002b) at the Lomonosov Ridge, indicate that Fram Strait branch water occasionally passes beyond the Gakkel Ridge, not just by mixing and heating the Barents Sea branch but also as a distinct contribution, retaining some of its high temperature and salinity characteristics. It would then recirculate along the Lomonosov Ridge and in the Amundsen Basin and also cross the Lomonosov Ridge into the Amerasian Basin. The reports from the NABOS moorings north of the Laptev Sea show that the temperature of the Atlantic

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layer increased suddenly in the early part of 2000s (Dmitrenko et al., 2005; Polyakov et al., 2005). Whether this was mainly due to warmer Fram Strait water entering the Arctic Ocean, or if different conditions north of the Laptev Sea can either allow the Fram Strait branch water to pass farther to the east or confine it to the Nansen Basin, is to date an unsolved question. Should the recent observed warming and changes in the Amerasian Basin be due to an increased inflow of warm Fram Strait branch water passing eastward at Severnaya Zemlya, the Laptev Sea area would be a critical choke point for the circulation of Atlantic water in the Arctic Ocean. Another possibility is that the changes downstream of the Laptev Sea mainly reflect variability in the Barents Sea branch, bringing either warmer or colder Atlantic core water over the Barents Sea into the Arctic Ocean.

The "Atlantic" part is here interpreted as a part of the Atlantic water from the Norwegian Sea that has survived the heat loss in the Barents Sea and retained a comparably high temperature. Another possibility or modification also exists; warm Fram Strait branch water that enters the St. Anna Trough from the north could become mixed into the Barents Sea branch in the trough and "jump stream" and join the Barents Sea branch on the shelf, contributing to the comparably high Atlantic water temperatures observed there (Fig. 4).

5 Time variability in the Eurasian Basin

A comparison of temperature and salinity sections taken in the Eurasian Basin from the Gakkel Ridge across the Amundsen Basin and over the Lomonosov Ridge into the Makarov Basin in 1991, 1996, 2001, 2005 and 2007 shows the time variability of the temperature of the Atlantic layer. We assume that the area covered, except the Makarov Basin, represents a flow towards Fram Strait (Fig. 15). The situation in 1991 shows that the Atlantic layer over the Gakkel Ridge is warmer than in the Amundsen Basin and at the Lomonosov Ridge. The temperature at the Lomonosov Ridge is higher than in the Amundsen Basin but both temperatures are close to the temperature maximum of the

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Barents Sea branch. The temperature in the Makarov Basin is considerably lower. In 1996 the Atlantic water temperature over the Gakkel Ridge is slightly lower, while the temperature over the Lomonosov Ridge has become significantly higher and is even higher than at the Gakkel Ridge. The Amundsen Basin has also become markedly warmer and the Makarov Basin slightly warmer. The changes in the Amundsen Basin and at the Lomonosov Ridge can be understood, if the inflow of warmer Atlantic water observed in 1990 (Quadfasel et al., 1991), which by then would have penetrated beyond the Laptev Sea choke point, was returning towards Fram Strait, more rapidly along the Lomonosov Ridge than in the Central Amundsen Basin.

In 2001 the Gakkel Ridge area was again the warmest and this also held for 2005 and 2007. This returning branch will contain a large fraction of Fram Strait branch Atlantic water and the variations between the years could reflect, with a time lag, the temperature variations of the Fram Strait inflow returning over the Gakkel Ridge. The temperature over the Lomonosov Ridge decreased between 1996 and 2001, while the temperature in the Amundsen Basin increased slightly. Both the Amundsen Basin and the Lomonosov Ridge temperatures dropped drastically in 2005. This only holds for half of the Amundsen Basin though. The stations closer to the Gakkel Ridge showed higher temperatures, gradually approaching those at the Gakkel Ridge. In the Makarov Basin the temperature had increased and was similar, or even higher, than at the Amundsen Basin side of the Lomonosov Ridge. This suggests that the warm Atlantic water entering the Makarov Basin in the 1990s (Carmack et al., 1995) now had circulated around the basin and was moving along the Lomonosov Ridge towards Siberia. A similar interpretation was made earlier by Kikuchi et al. (2005). In the Amundsen Basin and over the Lomonosov Ridge the lower temperatures imply that the warm Fram Strait branch water has been replaced by mostly Barents Sea branch water, supplying the Eurasian Basin beyond the Laptev Sea. This also correlates with the lower temperatures observed at the NABOS stations north of the Laptev Sea in the early 2000s (Dmitrenko et al., 2005). In 2007 the Makarov Basin temperature was somewhat reduced while the temperature at the Lomonosov Ridge was slightly higher than in 2005, indicating

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a possible presence of Fram Strait branch water or a warmer Barents Sea branch inflow at the Lomonosov Ridge. Over the period 1991 to 2007 there also appears to be a slight decrease in salinity of the Atlantic and intermediate layers between the Gakkel Ridge and the Makarov Basin. This all implies a rather rapid advection of the Atlantic water also in the interior of the basins and that the waters in the Amundsen Basin and at the Eurasian side of the Lomonosov Ridge are exchanged by advection from the Laptev Sea. On the Makarov Basin side of the Lomonosov Ridge the circulation is different, with a flow along the ridge towards Siberia.

6 Thermohaline intrusions

The almost universal presence of intrusions and interleaving structures in the Arctic Ocean deserves some comments. The intrusions are most frequent in frontal zones, indicating that they are created by the meeting and mixing of different water masses. This was the situation in 2007 north of Franz Josef Land and north of the Kara Sea. North of the Laptev Sea well-established intrusions were observed. This was also the case on the basin side of the Fram Strait branch on the third and the first sections (see Figs. 6, 7, 9, 10, 12, and 13 above). The presence of intrusions on the basin side of the Fram Strait branch has been taken as indication of a flow toward Fram Strait (Rudels et al., 1994). The advective velocities are with great certainty larger than cross frontal velocities of the intrusions caused by double-diffusive convection. If intrusions develop in a strong horizontal shear they would become stretched along the front by the shear, which likely would obstruct a significant cross frontal flow. The simpler view is that the two water masses move together, and the evolution of the intrusions can be described from a coordinate system following the motion. Since the less saline water on the basin side of the Fram Strait branch derives from the Barents Sea inflow, the most obvious route would be first eastward along the slope, after which both branches partly leave the slope and flow westward towards Fram Strait.

Whether the intrusions primarily are created during the eastward moving phase, when the gradients are strongest and subsequently advected westward, or if they form in the interior of the basin is not yet known. Much about the formation of Arctic Ocean intrusions is not well understood (for a short summary see Rudels et al., 2009). They are found in all possible background stratifications, also when both heat and salt are stably stratified, and linear stability theories do not work. Differential diffusion has been suggested as an important mechanism in these situations (Merryfield, 2002), but the time to establish the intrusions would then be on the order of years (Merryfield, 2002; Kuzmina et al., 2012). This appears long considering the interleaving structures in the intermediate depth range encountered along the continental slope and in the Eastern Nansen Basin (Figs. 7–9), and also by the occurrence of intrusions almost immediately, when the Makarov Basin deep water passes through the intra-basin in the Lomonosov Ridge into the Amundsen Basin (Björk et al., 2007).

7 Deep and bottom waters

The temperature and salinity profiles and the θS diagrams indicate that the deep and bottom waters of the Arctic Ocean are renewed, or at least transformed, within the Arctic Ocean. The strong stratification in the upper part of the deep basins excludes deep convection, and the most likely process is ice formation in lee polynyas over the shallow shelves (Nansen, 1906). Freezing and brine rejection create saline and dense waters that eventually cross the shelf break and descend into the deep Arctic Ocean as entraining boundary plumes. When the plumes reach their neutral density level they merge with the surrounding water. Depending upon their initial salinity (density) the plumes may supply the halocline, merge with the waters of the boundary current, or bypass the intermediate layers and enter the deep and bottom waters. Because of the entrainment of ambient water during their descent the temperature of the plumes increases and their salinity decreases. Below 2000–3000 m the plumes add both heat

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and salt to the water column creating a layer of increasing salinity and temperature towards the bottom (Rudels, 1986; Quadfasel et al., 1988).

The evolution of the deep water along the Eurasian Basin continental slope can be seen in the deeper layer on the different sections. At the first section a less saline layer at 1500–2000 m indicates a presence of Arctic Intermediate Water or Nordic Seas deep water (Fig. 16). This less saline layer is also observed on the second and third section but here the salinity has increased somewhat. The deep Barents Sea branch inflow occurs before section 3, and on section 3 it is seen as a reduction of the salinity between 1000 and 1500 m, and the vertical deep part of the θS curves of sections 1 and 2 starts to bend towards lower salinity (Fig. 17). First at the Laptev Sea section (4) there are indications of convecting water increasing the salinity in the deeper layer and thus change the slope of the θS curves, making it more perpendicular to the isopycnals (Fig. 18). The plumes observed in 2007 were, however, not dense enough to renew the deepest layers, which thus appear to be ventilated under harsher conditions than those prevailing during the last 10–15 yr.

Two features are remarkable about the deep waters in the Nansen, Amundsen and Canada basins. (1) A temperature minimum located about 800 m above the bottom (1200 m in the Canada Basin) and the slightly warmer, thick, 600–800 m, homogenous bottom layer (see Figs. 16–20). The increased temperature towards the bottom has been proposed to be caused by geothermal heating (Timmermanns et al., 2003; Björk and Winsor, 2006). That geothermal heating can influence the deeper layers is seen from stations taken close to subsurface volcano at the Gakkel Ridge (Fig. 19). The deep temperature minimum present elsewhere in the Eurasian Basin is here removed and also the layers above the injection are warmer than at the surrounding stations. A closer look at the deepest layers along the Gakkel Ridge also indicates a large variability of the deep waters with smaller scales intrusions of anomalous water that likely derives from the geothermal activity at the ridge (Fig. 19).

The Makarov Basin is different. Here no temperature minimum is present and as the salinity becomes constant about 800 m above the bottom the temperature continues to decrease another 200 m (Fig. 20). This led Jones et al. (1995) to propose a spillover of deep Amundsen Basin water across the central part of the sill. In 2005, when *Oden* studied the area where such an overflow could occur, no flow from the Amundsen Basin to the Makarov Basin was observed. Instead the densest water at the sill was from the Makarov Basin and a flow of Makarov Basin deep water into the Amundsen Basin was found (Björk et al., 2007). If a transport of deep water from the Amundsen Basin to the Makarov Basin takes place, this overflow must be intermittent (Rudels, 2012).

The bottom water temperature has increased in the Makarov Basin and some indications of a salinity increase are seen but the changes are very close to the observational accuracy (Fig. 20). If both a temperature rise and a salinity increase are seen, it could be due to advection and a gradual stronger presence of the warmer and more saline Canada Basin deep and bottom water in the absence of a spill-over of less saline Amundsen Basin deep water (Fig. 20).

A rough comparison between the *Polarstern* sections across the Eurasian Basin taken in 1996 and 2007 shows an increase in temperature in the bottom 1500 m by 0.015 °C, indicating a heat flux of $\sim 275\,\mathrm{m\,W\,m^{-2}}$, which is 2–4 times larger than the average geothermal heat flux 50–100 $\mathrm{m\,W\,m^{-2}}$ (Langseth et al., 1990) (Fig. 21). An additional heat source is then required. The fact that the salinity is increasing towards the bottom cannot be explained by geothermal heating and slope convection must, at least intermittently, reach the deepest layers. Depending upon the characteristics of the ambient waters, which the descending plumes pass through, their final temperature will vary with time, and slope convection can thus also contribute to the observed heating of the lower layers. However, no significant increase in salinity is observed between the sections, suggesting weak deep reaching boundary convection during this period.

That shelf-slope convection can provide water dense enough to enter the deep Arctic Ocean has been observed in Storfjorden (e.g. Quadfasel et al., 1988) and also in the Arctic Ocean north of Severnaya Zemlya have saline, dense and warmer bottom layers been observed (Rudels et al., 2000a). In 2007 warmer, more saline and dense bottom

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layers were observed at the Kara and Laptev Sea slopes between 2500 m and 3500 m, indicating that dense plumes were present also in 2007 (Figs. 17 and 18).

On the Laptev Sea slope at the 2000 m isobath an about 1000 m thick, warm and saline layer was observed between 750 m and 1750 m, almost reaching the bottom (Figs. 10 and 22). This layer, not present west of Severnaya Zemlya, could not derive from the colder and less saline Barents Sea branch inflow, and a likely source is again the shelf area around Severnaya Zemlya. Fram Strait branch water could, by transient eddy motions or by meandering of the boundary current, be brought close to the slope. Saline water formed on the shelf would then interact with the warmer Fram Strait branch instead of passing through the cold, less saline Barents Sea branch water. The mixing and heating would lower the density of the plumes and remove the thermobaric effect, which promotes the sinking of colder plumes. Such a process could explain the observed warm, less dense characteristics and the fact that the water mass had already become detached from the bottom, perhaps eventually to form a thick, warm subsurface eddy.

That eddies are created is evident from observations both from the near surface layer close to the Laptev Sea as well as for the intermediate water range in the intra-basin in the Lomonosov Ridge, where a 1000 m thick eddy with almost undiluted Barents Sea branch characteristics was found. How an eddy can survive such long translation is again an interesting question but could be taken as an indication that Barents Sea branch water does reach the Lomonosov Ridge and then perhaps also the Makarov Basin through the different rifts in the ridge (Fig. 22).

8 Summary

The impact of the Barents Sea inflow branch on the Eurasian Basin and the interactions between the Barents Sea and the Fram Strait inflow branches have been examined. The meeting between the two branches leads to strong isopycnal mixing and to the creation of intrusive layers, not only north of the Kara Sea between the main Barents

Sea branch inflow along the St. Anna Trough but also north of Franz Josef Land, where a smaller fraction of the Barents Sea inflow enters the Nansen Basin. The intrusions observed north of the Laptev Sea are similar to those found in the interior of the Nansen and Amundsen basins beyond the warm core of the Fram Strait branch. This indicates a turning of a large part of the boundary current within the Nansen Basin and a flow towards Fram Strait.

The temperature of the Atlantic layer in the Fram Strait branch stays fairly constant from Svalbard to Severnaya Zemlya but then drops by one degree between the Eastern Kara Sea and the Laptev Sea. This can (1) be explained by the cooling resulting from strong mixing between the inflow branches, but could (2) also be due to an extreme heat loss from the Atlantic layer north of Severnaya Zemlya. However, the lower temperatures could also be (3) an indication that a substantial fraction of the Fram Strait branch never penetrates across the Gakkel Ridge but returns towards Fram Strait, consistent with the existence of intrusions on the basin side of the Fram Strait branch. This is the interpretation favoured here.

The heat loss of the Atlantic water to the atmosphere and to sea ice seems to be confined to north of Svalbard and along the continental slope north of the Barents and Kara seas. Here warm Atlantic water comes close to the sea surface and wind mixing as well as other mechanical mixing processes generated at the slope might increase the entrainment of Atlantic water and heat into the upper layer (Fig. 5).

The implication is that the Atlantic and intermediate waters in most parts of the Arctic Ocean derive from the Barents Sea branch, not from the Fram Strait branch. The variations in temperature in the Atlantic layer in the Arctic Ocean should then be sought in the variability of the Barents Sea inflow, both at the Barents Sea Opening and by the atmospheric conditions over the Barents Sea, rather than in changes in the Atlantic water entering through Fram Strait. Only occasionally would the Fram Strait branch penetrate farther than the Nansen Basin to contribute to the changes in the Atlantic layer of the Arctic Ocean in the Amundsen Basin and beyond the Lomonosov Ridge.

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The deeper layers in the Eurasian Basin have shown an increase in temperature of 0.015 degrees over the last 10 yr, which is about 4 times that expected from geothermal heating. That geothermal heating can strongly influence the temperature profiles in the deeper layer was seen at stations taken close to a thermal vent in the Gakkel Ridge. Here the temperature minimum normally present above the homogenous 600–800 m thick bottom layer was absent. Slope convection can also contribute to the warming of the deeper layers, if slope plumes initially saline enough to enter the deeper layers are sinking through and entrain warmer water. Observations at the slope showed the presence of comparably warm and saline bottom layers at the Laptev Sea continental slope deeper than 2000 m.

Higher up on the Laptev Sea slope, at 2000 m, a 1000 m thick, warmer and more saline layer was observed between 750 and 1750 m. This layer was not present at the Eastern Kara Sea slope, which suggests that slope convection originating from around Severnaya Zemlya could have passed through warm Atlantic water, perhaps from the Fram Strait branch, and attained the higher temperature and salinity. This layer was less dense than the basin water column at the corresponding level and thus ready to detach from the slope, perhaps to form a warm intermediate water eddy.

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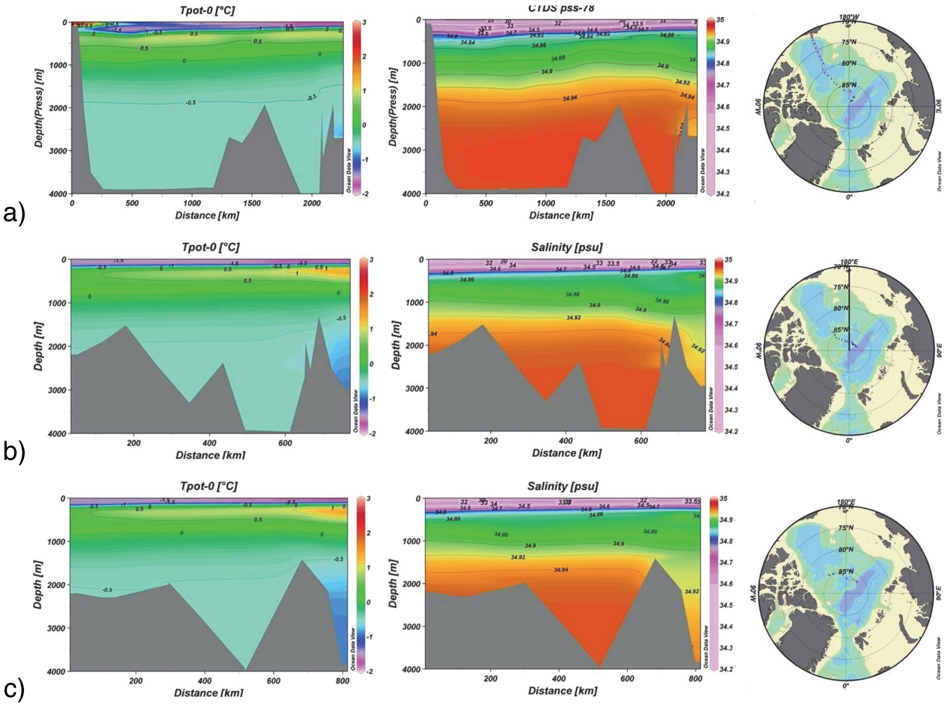

Fig. 1. (a) Sections of potential temperature and salinity from Alaska to the Lomonosov Ridge taken by IB *Oden* in 2005. **(b)** Sections of potential temperature and salinity between the Alpha Ridge and the Lomonosov Ridge taken by R/V *Polarstern* 2007 (western section). **(c)** Sections of potential temperature and salinity between the Alpha Ridge and the Lomonosov Ridge taken by R/V *Polarstern* 2007 (eastern section). Figures are created using Ocean Data View (Schlitzer, 2012).

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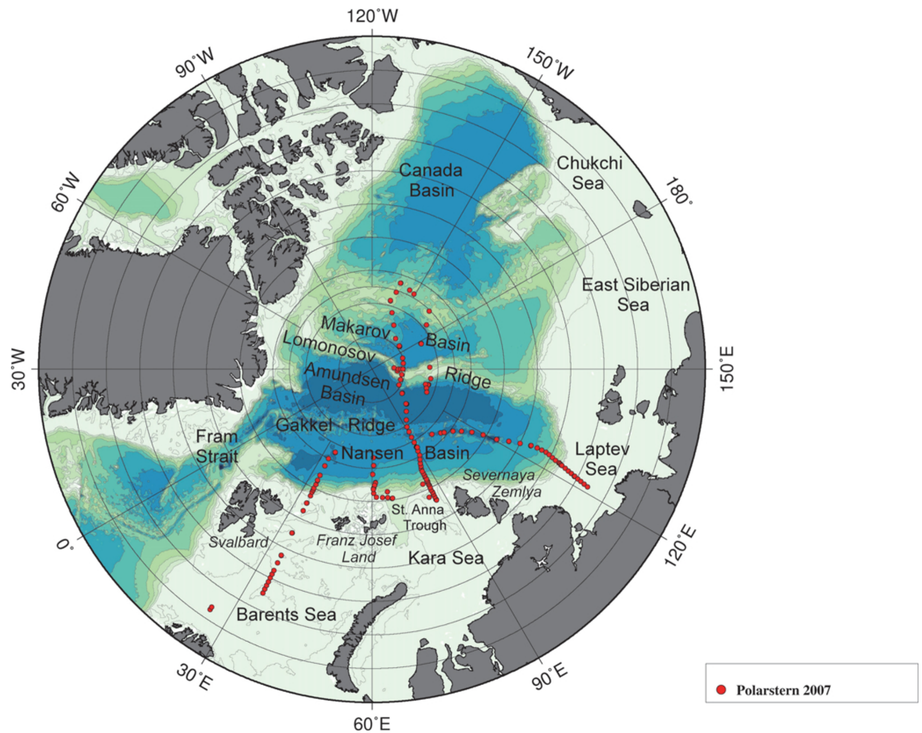

Fig. 2a. The station positions on the SPACE cruise 2007 with R/V *Polarstern*.

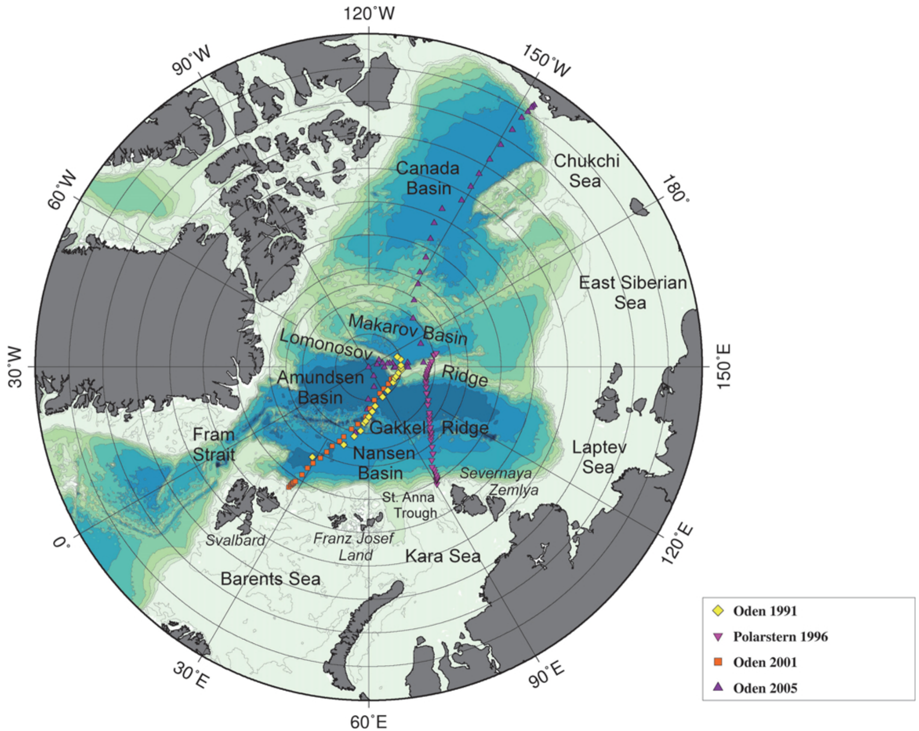

Fig. 2b. The positions of the stations from *Oden* 1991, *Polarstern* 1996, *Oden* 2001 and *Oden* 2005 used in this work.

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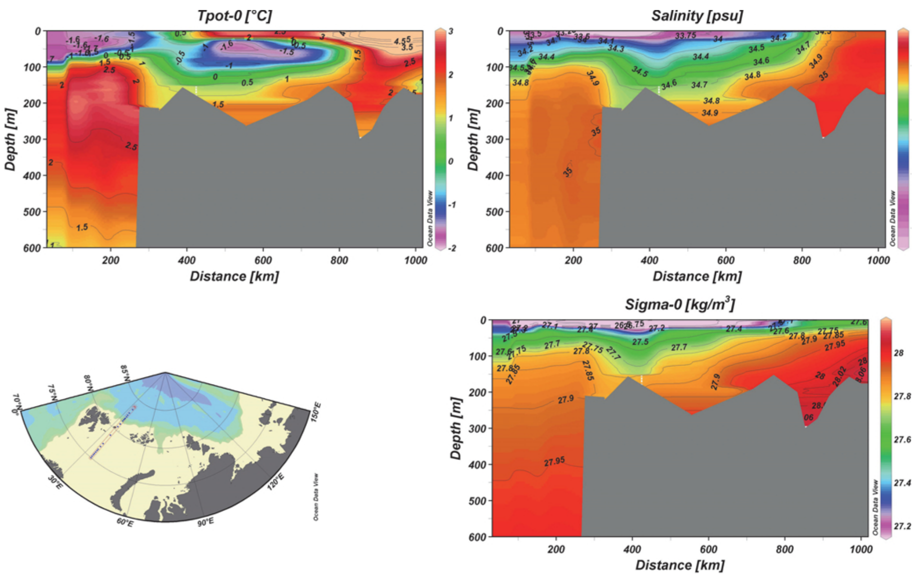

Fig. 3. Potential temperature, salinity and potential density sections from the SPACE transect along 30° E from the Barents Sea into the Nansen Basin.

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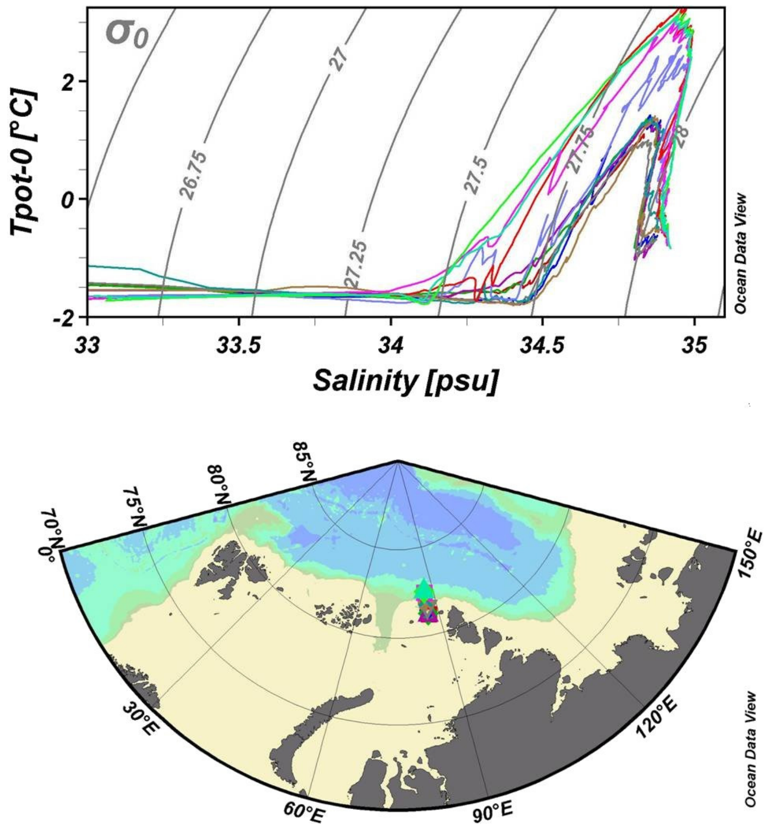

Fig. 4. θS curves from the shelf and slope in the eastern Kara Sea showing the colder, less saline Barents Sea branch on the shelf and the warmer more saline Fram Strait branch on the slope. The winter mixed layers are located above the thermocline and here the Barents Sea branch is more saline than the Fram Strait branch.

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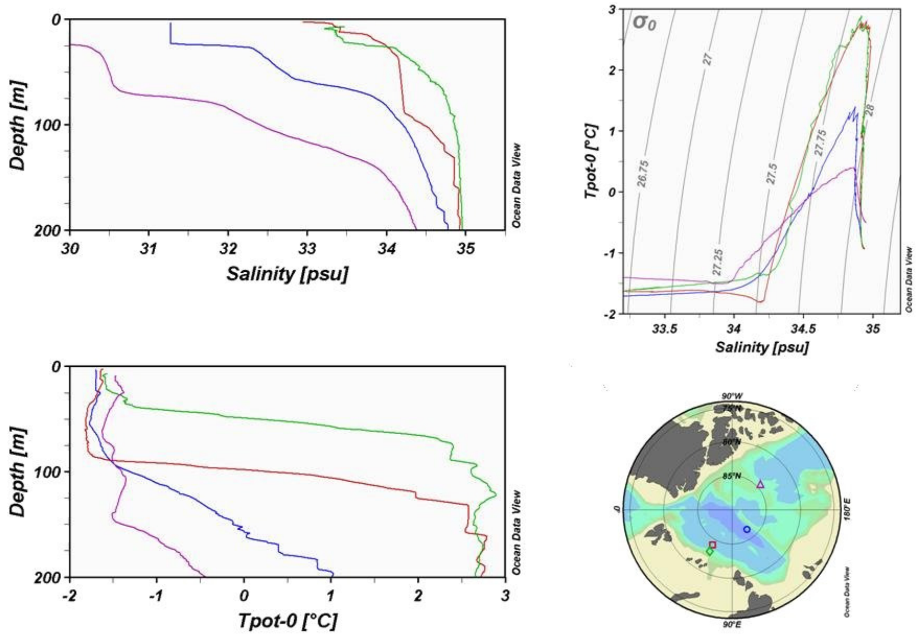

Fig. 5. Profiles of salinity and potential temperature and θS curves from the upper layers in different parts of the Arctic Ocean showing the upper temperature minima created by winter convection. Only at the Eurasian continental slope (green station) is the minimum absent, indicating that warmer water is entrained from below into the mixed layer in summer.

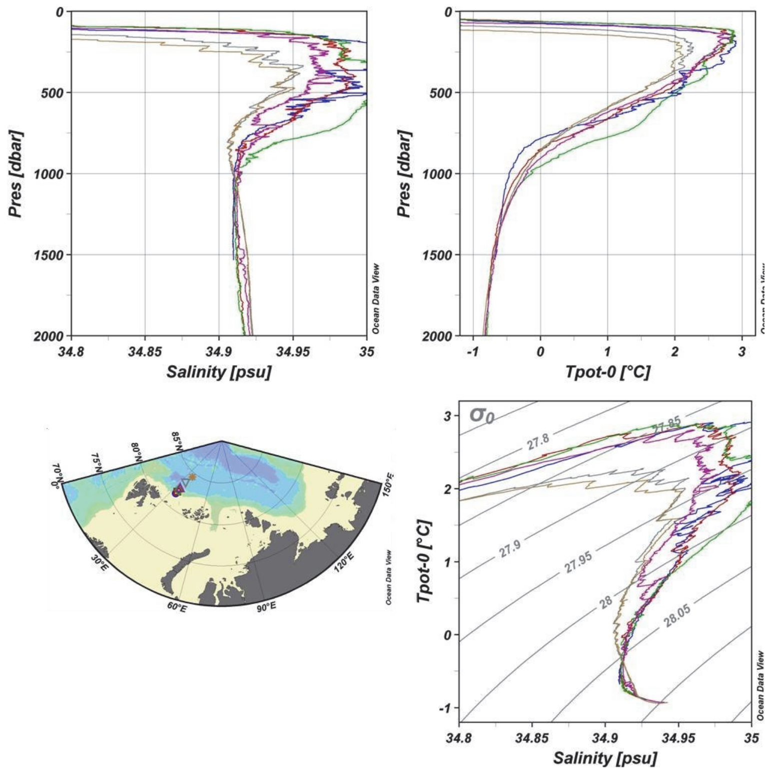

Fig. 6. Profiles of salinity and potential temperature, and θS curves from the deep stations on the first SPACE section. Notice the colder, less saline Atlantic water and the salinity minimum at 900 m on the northern stations and the deep (1500 m), weak salinity minimum closer to the slope (better seen in the θS diagram).

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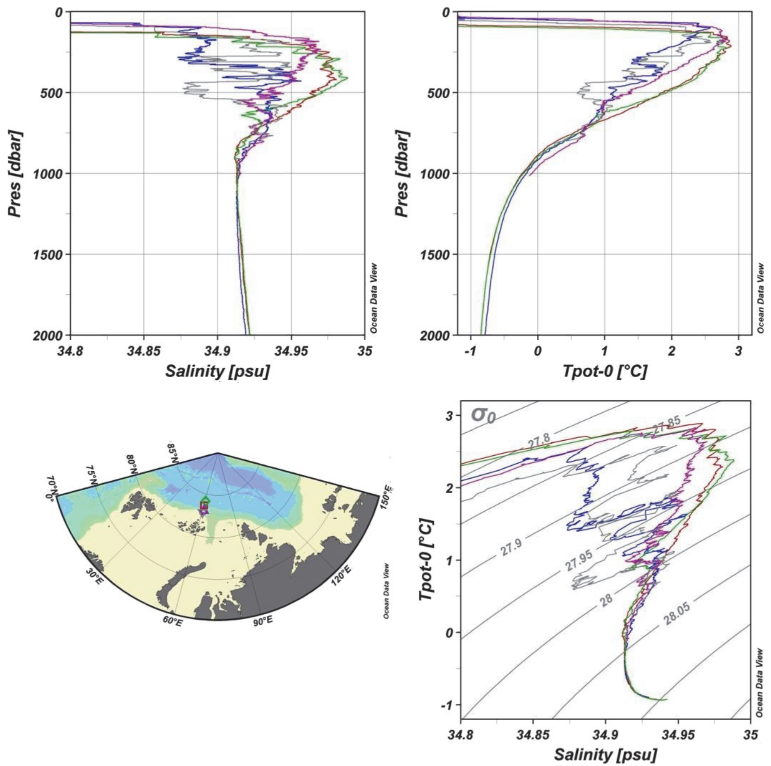

Fig. 7. Profiles of salinity and potential temperature and θS curves from the second SPACE section north of Franz Josef Land showing the intrusions of colder and less saline water from the Barents Sea in the water column at the slope.

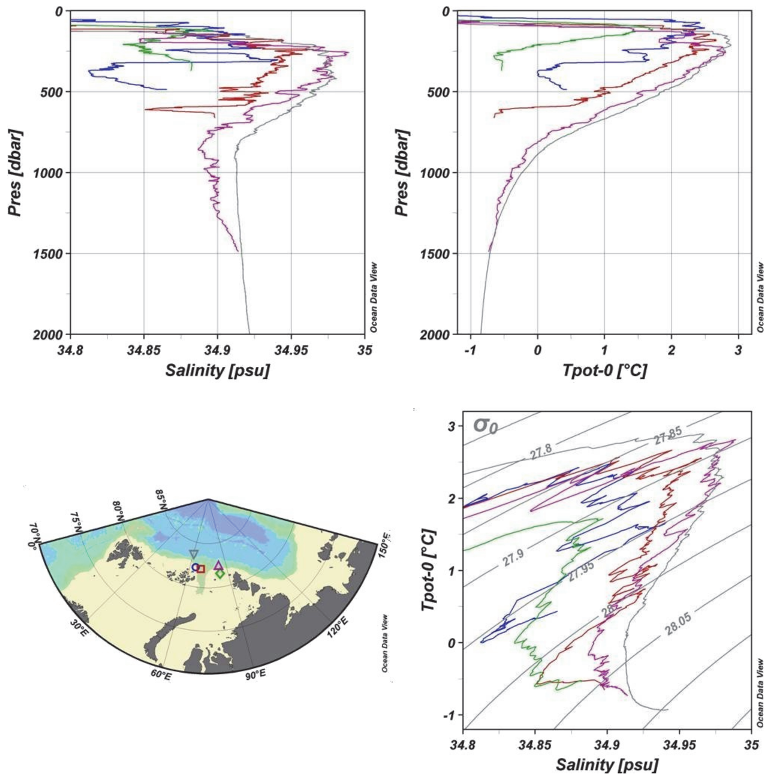

Fig. 8. Profiles of salinity and potential temperature and θS curves from the Nansen Basin (grey station) and from the shelf (blue station) north of Franz Josef Land and from St. Anna Trough (red station) showing the warmer Fram Strait branch Atlantic water and the denser water on the shelf and the Fram Strait branch water entering the trough from the north. The green and magenta stations on the Kara Sea shelf and slope show the Barents Sea branch remaining on the shelf, but the denser part of the Barents Sea inflow is seen as a low salinity layer at the slope between 750 m and 1500 m (magenta station).

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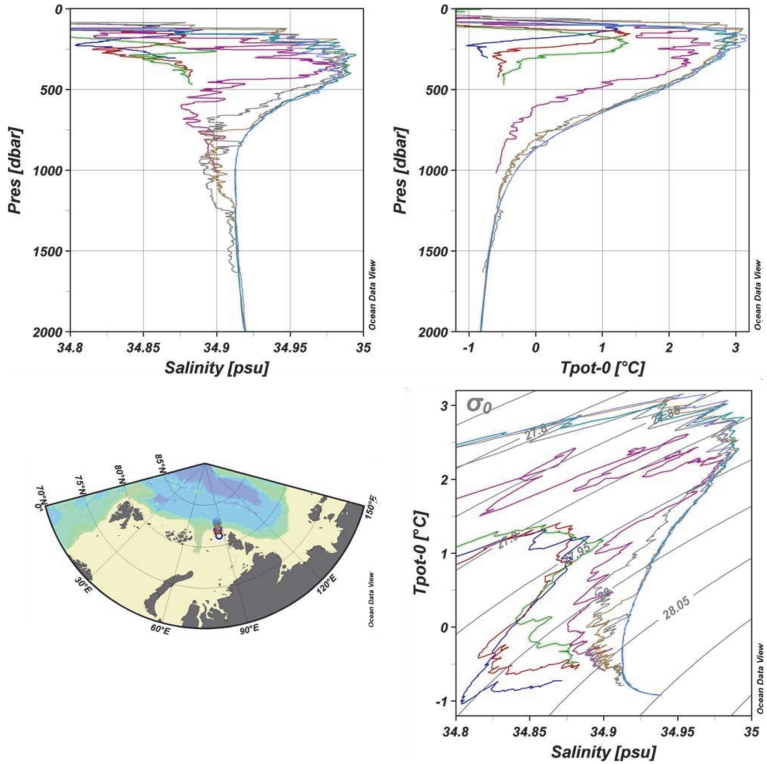

Fig. 9. Profiles of salinity and potential temperature and θS curves from the shelf and slope in the eastern Kara Sea. The three dense contributions from the Barents Sea are seen as a temperature maximum, a salinity and a temperature minimum, and a cold, more saline bottom water. The magenta station shows strong isopycnal mixing between the two branches. On the slope Barents Sea branch water that has entered the deep basin directly from the St. Anna Trough is seen as a low salinity layer (grey and yellow stations). The intrusions in the Fram Strait branch are found at the temperature maximum but above the salinity maximum.

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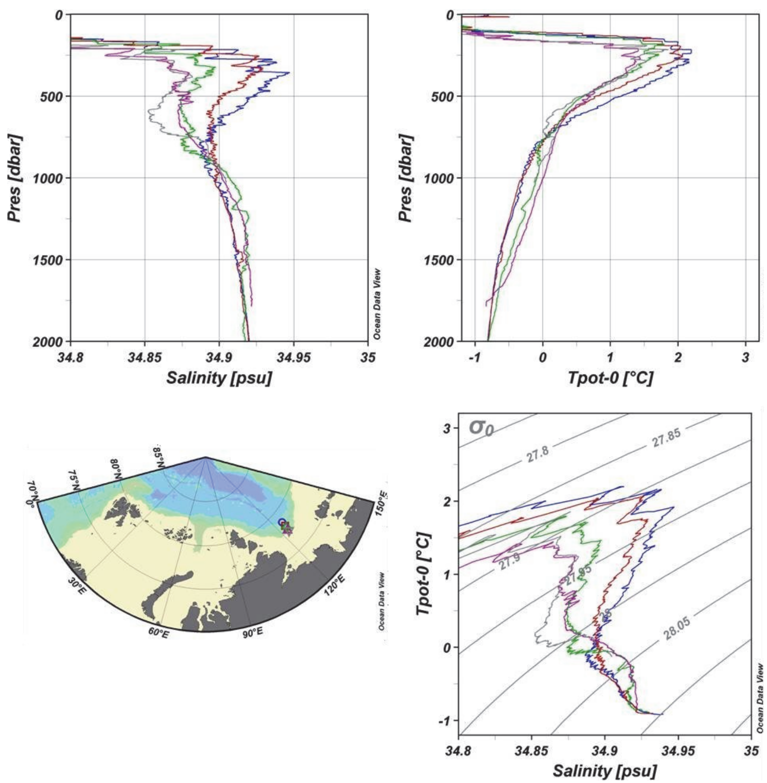

Fig. 10. Profiles of salinity and potential temperature, and θS curves from the Laptev Sea slope. No smooth temperature and salinity profiles are seen in the Fram Strait branch Atlantic core and the presence of intrusive layers indicates mixing between the branches. The Atlantic water is still warmer and more saline over the deeper part of the slope (red and blue stations) than closer to the shelf but has become considerably colder and less saline compared with section 3 (Fig. 9).

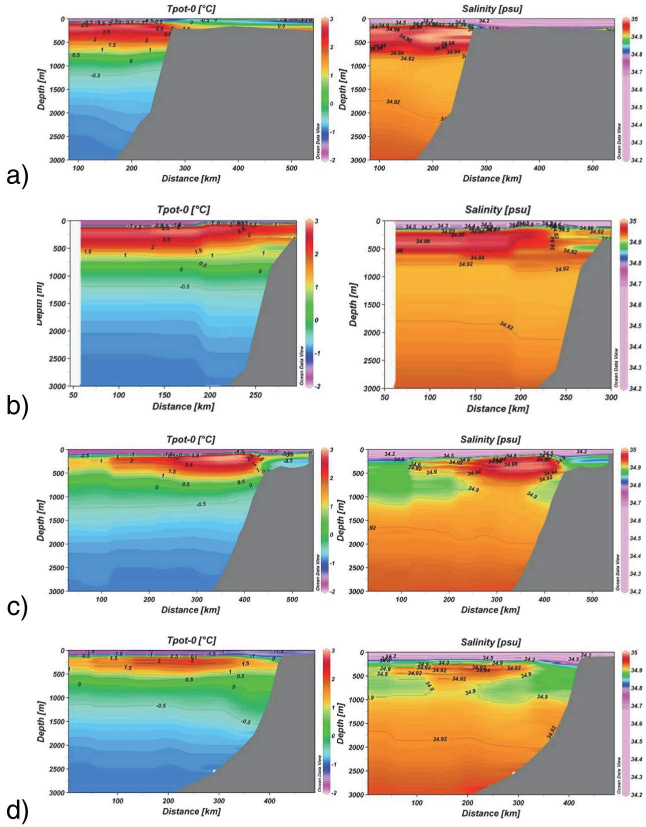

Fig. 11. Potential temperature and salinity sections from the shelf into the deep Nansen Basin in **(a)** western Barents Sea, **(b)** Franz Josef Land, **(c)** eastern Kara Sea, **(d)** Laptev Sea.

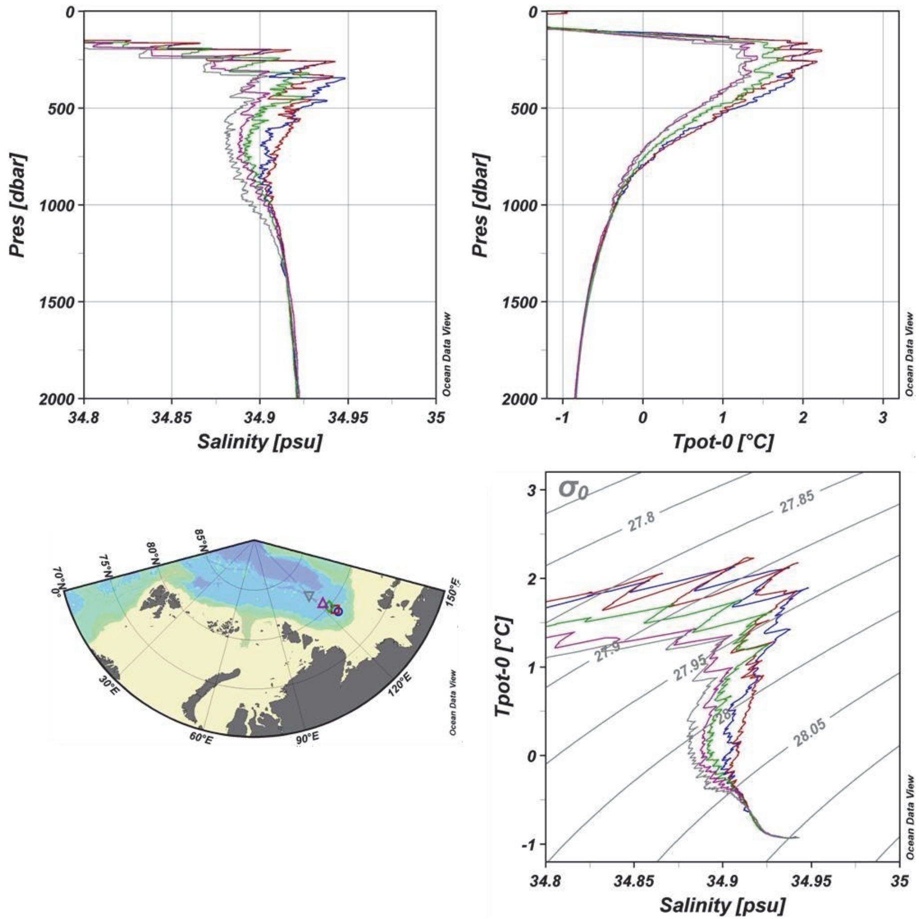

Fig. 12. Profiles of salinity and potential temperature and θS curves from SPACE section 4 along the Gakkel Ridge showing the regular interleaving layers and the gradual reduction in temperature and salinity from the slope to the basin.

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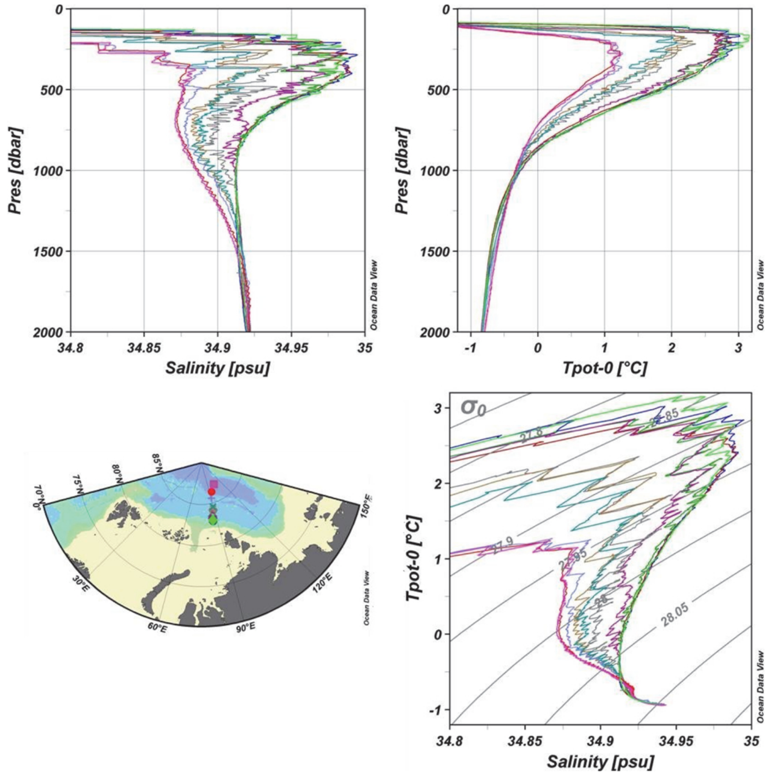

Fig. 13. Profiles of salinity and potential temperature, and θS curves from the SPACE section 3 from the slope into the Amundsen Basin. The temperature and salinity decrease towards the Amundsen Basin and two fronts are seen, both located in the Nansen Basin, where also the strongest interleaving is observed. The interleaving layers are here mostly located below the temperature maximum and at and below the salinity maximum. This is different from the situation at the slope (Fig. 9).

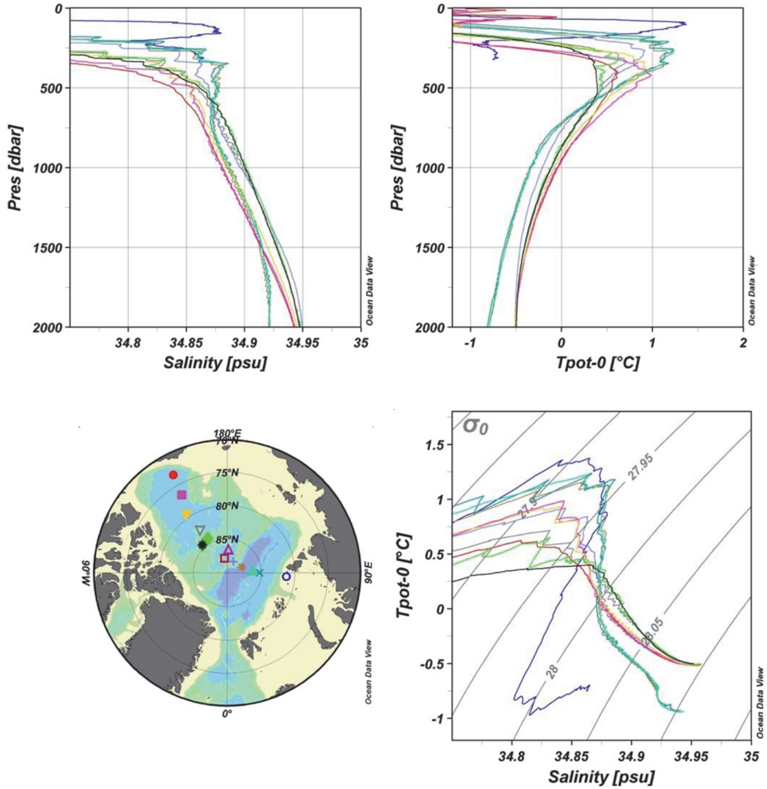

Fig. 14. Profiles of salinity and potential temperature and θS curves showing the properties of the Barents Sea branch and the characteristics of the water column in the Amundsen Basin, Makarov Basin and Canada Basin and over the Lomonosov Ridge and over the Mendeleyev and Alpha ridges. The Barents Sea branch is warm enough to supply the temperature maximum in most of the Arctic Ocean basins.

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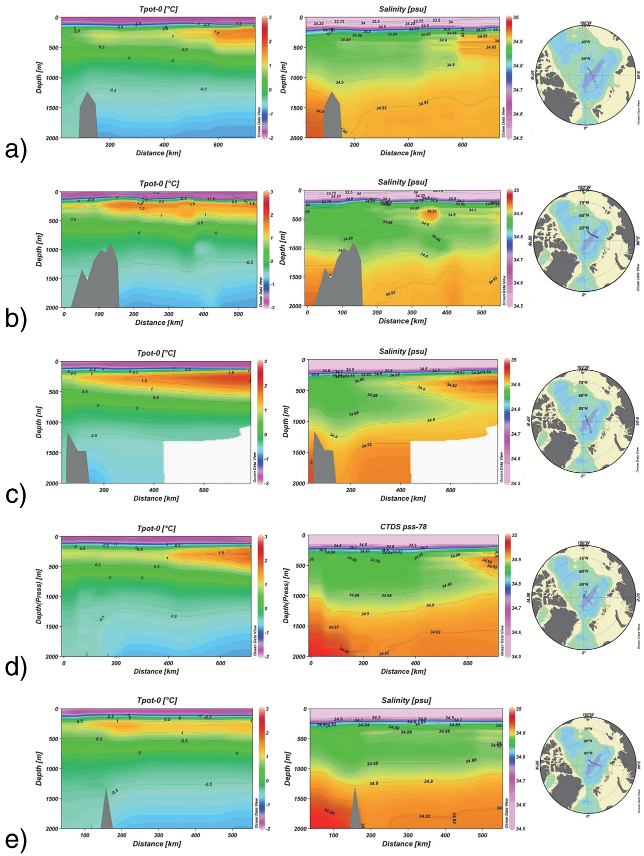

Fig. 15. Sections of potential temperature and salinity from the Gakkel Ridge (right) across the Amundsen Basin and the Lomonosov Ridge into the Makarov Basin from different years; **(a)** 1991 (*Oden*), **(b)** 1996 (*Polarstern*), 2001 (*Oden*), **(d)** 2005 (*Oden*), and **(e)** 2007 (*Polarstern*).

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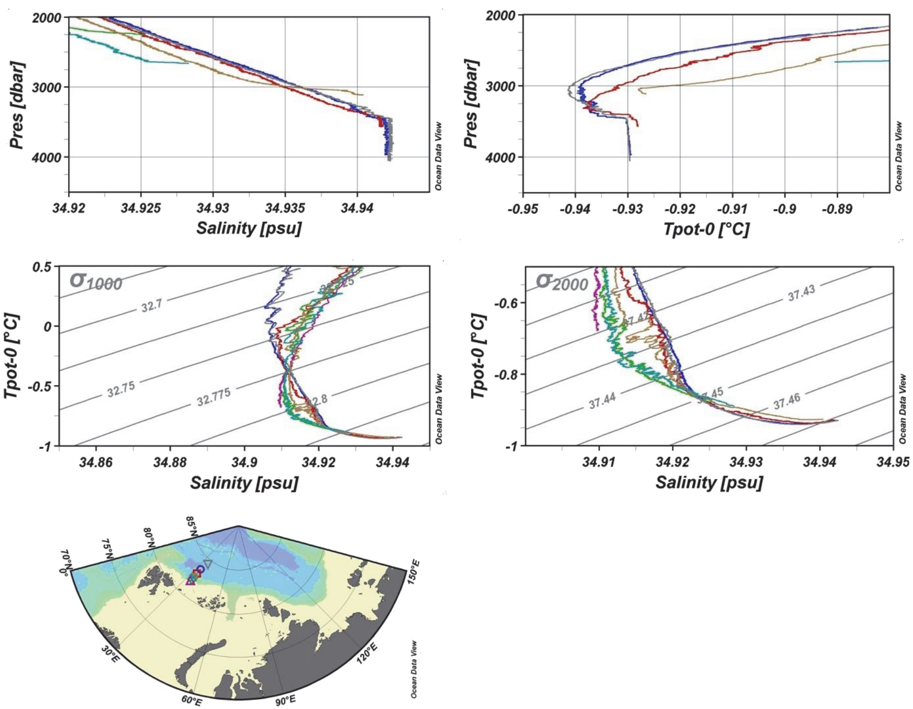

Fig. 16. Profiles of salinity and potential temperature and θS curves showing the characteristics of the deep and bottom water in the Nansen Basin on the SPACE section along 30° E and the presence of the deep temperature minimum and the homogenous bottom layer below (profiles) and the less saline deep inflow from the Nordic Seas through Fram Strait (θS curves).

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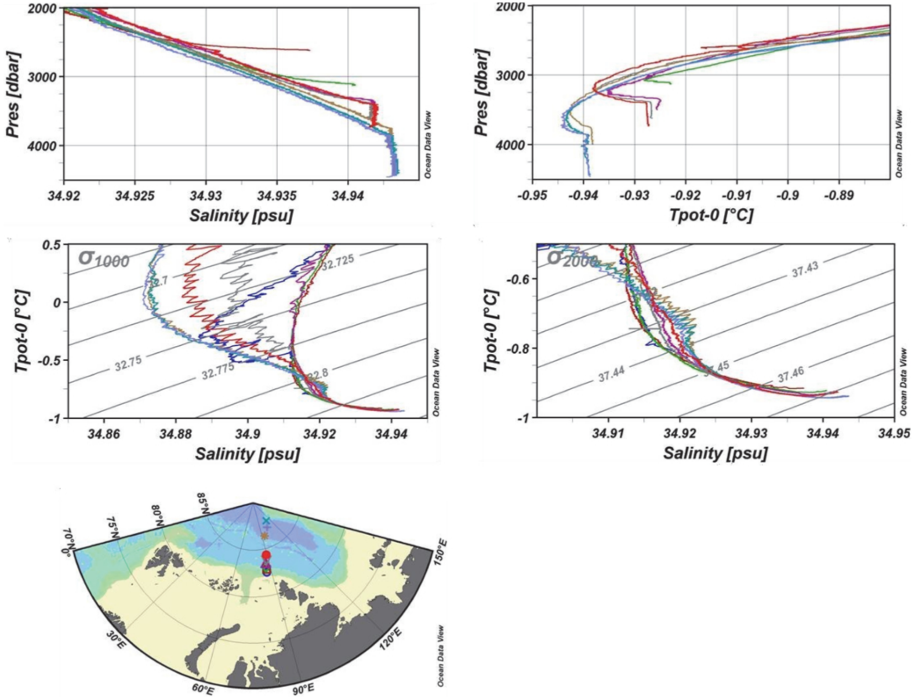

Fig. 17. Profiles of salinity and potential temperature and θS curves showing the deep and bottom water in the Nansen and Amundsen basins along SPACE section 3. A deep temperature minimum and a homogenous bottom layer are present in both basins. The Amundsen Basin is deeper and its bottom water colder and more saline. In the θS curves the less dense salinity minimum, found both at the slope (blue station) and in the interior, derives from the Barents Sea inflow. The deep salinity maximum is observed in the Amundsen Basin and indicates an inflow of Amerasian Basin deep water across the Lomonosov Ridge.

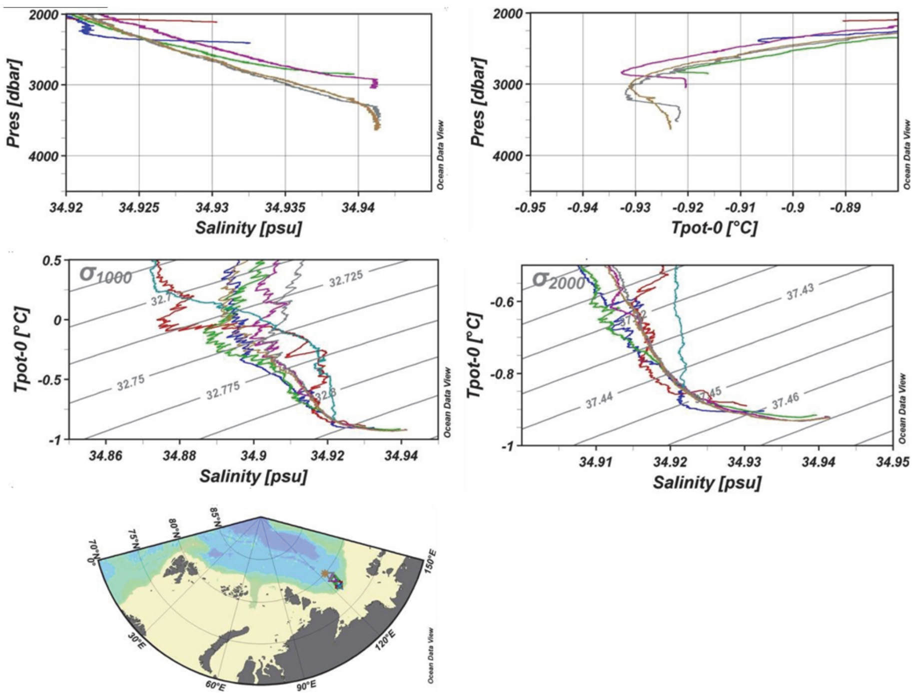

Fig. 18. Profiles of salinity and potential temperature, and θS curves from SPACE section 4 north of the Laptev Sea. Here the slope of the deeper part of the θS curves has become more perpendicular to the isopycnals, indicating both a salinity reduction due to the Barents Sea inflow and a deeper salinity increase due to slope convection and entraining, saline and dense plumes bringing heat and salt downwards. Bottom water with increasing temperature and salinity with depth are observed at the slope between 2000 m and 3500 m (green and magenta stations).

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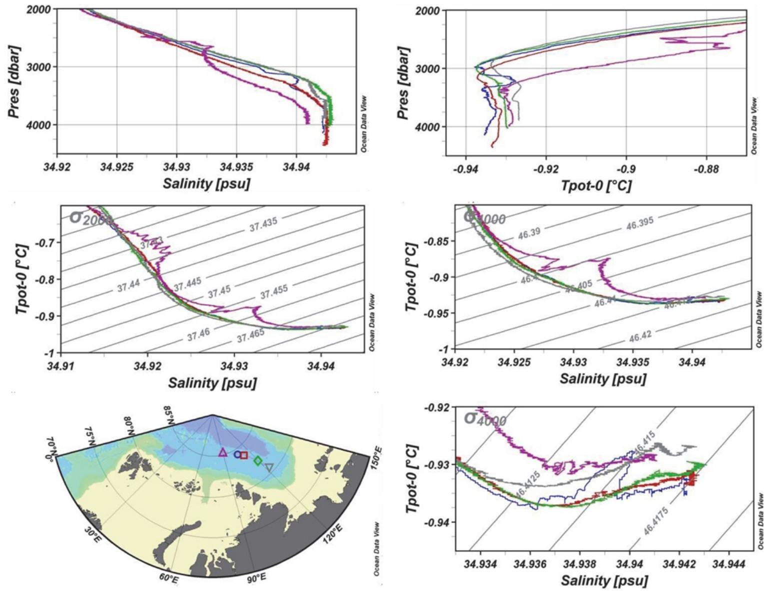

Fig. 19. Profiles of salinity and potential temperature, and θS curves shown with different blow-up levels from SPACE stations taken above the Gakkel Ridge. The simple deep water structure found in the basins is here disrupted by several different maxima and minima. A strong input of warmer and more saline water is found at 2700 m (magenta station) and a smaller deep intrusion is seen between 3100 m and 3300 m (blue station). The cause of these maxima could be geothermal activity at the ridge.

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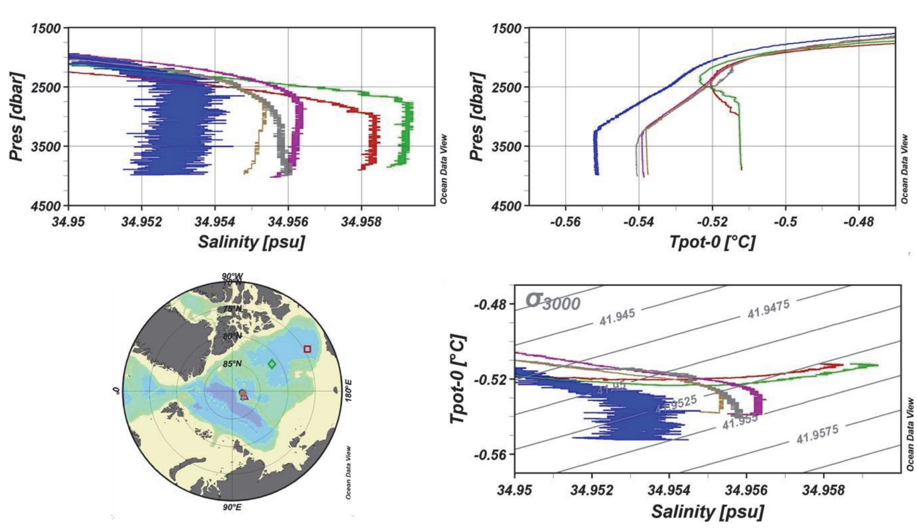

Fig. 20. Profiles of salinity and potential temperature, and θS curves from the Makarov Basin taken in 1991 (blue station), 2001 (grey station), 2005 (magenta station) and 2007 (brown station). The green and blue stations are from the southern (red) and northern (green) Canada Basin taken in 2005. The deep temperature minimum is absent in the Makarov Basin but the Makarov Basin deep water could supply the temperature minimum of the Canada Basin, which is located at the same level as the sill depth of he Alpha and Mendeleyev ridges. The temperature of the Makarov Basin bottom water has increased suggesting either heating or a temporary absence of colder deep water from the Amundsen Basin spilling over the ridge that otherwise would cool the deeper layers. The salinity in the bottom layer might be increasing with time, which would require some influx of the more saline Canada Basin deep water, not just geothermal heating. However, the observed salinity increase is not continuous in time and is very close to instrumental limitations.

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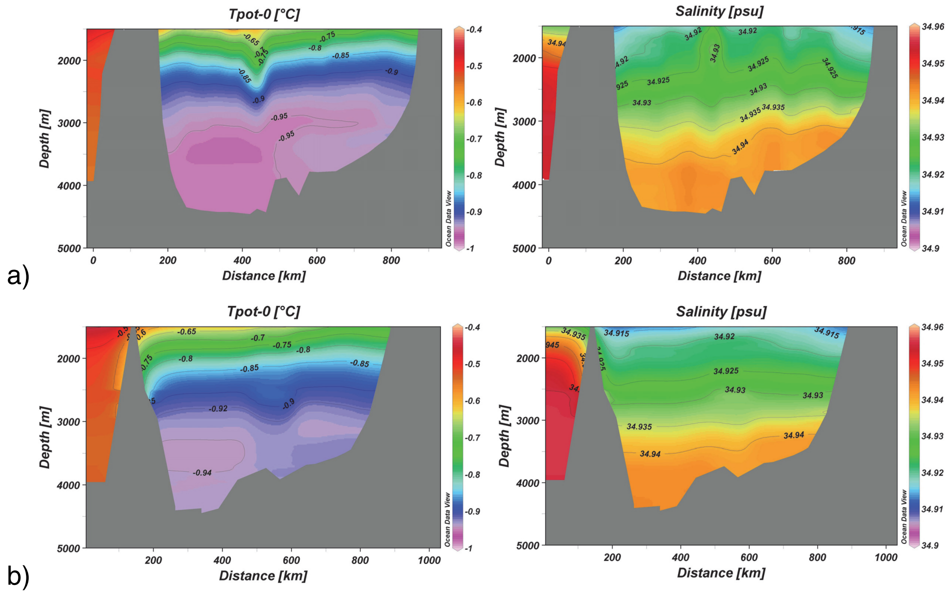

Fig. 21. Potential temperature and salinity sections across the Eurasian Basin taken in 1996 **(a)** and 2007 **(b)** showing the temperature increase and, a possible, salinity decrease below 1500 m. The mean temperature difference is estimated to 0.015 $^{\circ}$C.

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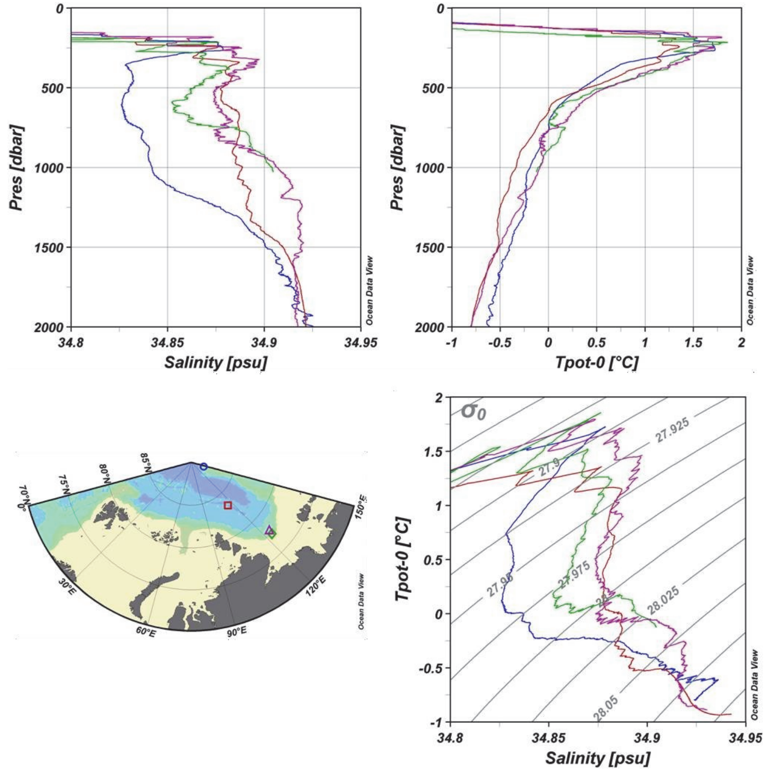

Fig. 22. Profiles of salinity and potential temperature and θS curves showing different eddies observed on the SPACE cruise 2007. Especially the 800 m thick eddy of Barents Sea branch water found in the intra-basin in the central Lomonosov Ridge (blue station) is remarkable.

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