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TWO SOURCES FOR THE LOWER HALOCLINE IN THE ARCTIC OCEAN

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Abstract

The lower halocline in the Amundsen and Makarov Basins of the Arctic Ocean is derived from the inflow from the Norwegian Sea through Fram Strait which interacts with sea ice to become fresher and colder, creating a deep winter mixed layer in the Nansen Basin that eventually supplies the lower halocline. The near surface inflow from the Norwegian Sea through the Barents Sea incorporates the run-off from the large Siberian rivers. It contributes to the Polar Mixed Layer, capping the denser mixed layer formed from the Fram Strait inflow. The main inflow from the Norwegian Sea over the Barents Sea enters the Arctic Ocean at the St. Anna Trough and its upper part provides a second source for the lower halocline. In the Nansen Basin this Barents Sea branch contribution to the halocline is as cold as the Fram Strait contribution but more saline. Further to the east it remains more saline, but becomes warmer than the Fram Strait derived lower halocline. This is a result of stronger vertical mixing and entrainment over the continental slope. The Barents Sea branch component, initially confined to the Siberian continental slope, eventually becomes the main source of the lower halocline in the Canada Basin beyond the Chukchi Cap.

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1 Background

The water masses of the Arctic Ocean derive from two sources – the Atlantic and the Pacific Oceans. The Atlantic contributes to the upper, intermediate and deep water masses. In particular, the inflow through Fram Strait maintains the warm subsurface "Atlantic" layer present throughout the Arctic Ocean. A second branch, entering over the Barents Sea, renews the intermediate layers and supplies the oceanic component to the shelf waters in the Arctic marginal seas, balancing the freshwater input from the rivers. The Pacific inflow, occurring through the Bering Strait, is smaller, less dense and mainly confined to the upper layers of the Canadian Basin. The broad, shallow shelves of the marginal seas play a dual role in the Arctic Ocean water mass transformations, supplying waters to the upper layers as well as ventilating the deep basins through shelf-slope convection.

The low salinity surface water creates a stratification strong enough to limit the depth of the winter convection in the deep basins, allowing the surface layer to attain freezing temperature. Ice forms and remains throughout the year. The presence of a permanent ice cover has a profound effect on the Arctic climate. The loss of sensible heat from the ocean is reduced and replaced by latent heat released by sea ice formation. The albedo increases and more of the incoming short wave radiation is reflected to space. In summer the solar radiation nevertheless leads to partial melting of the sea ice and creates a seasonal, low salinity, melt water layer, which is removed by ice formation in autumn. The less dense, upper waters isolate the Atlantic layer from the surface processes and act as a barrier that prevents the heat content of the Atlantic water from influencing the sea ice and the atmosphere.

The waters above the temperature maximum of the Atlantic layer can be separated into 3 and ½ layers. The ½ layer representing the seasonal low salinity surface water. The 3 more permanent layers are; 1) the Polar Mixed Layer (PML), homogenised in winter by haline convection driven by brine rejection during ice formation; 2) the halocline, a layer with approximately constant temperature, close to freezing, but with salinity increasing with depth; 3) the thermocline, where both salinity and temperature increase with depth. The last two layers together constitute the pycnocline between the PML and the temperature maximum of the Atlantic layer.

This basic structure is subject to large spatial, and as recent observations suggest, also to temporal variations. One recently observed, and intensely discussed, feature is a change in the Eurasian Basin from a 3 to a 2 layer structure consisting of a deep (~100m) upper layer with approximately constant salinity and with temperature close to freezing overlying a pycnocline with coinciding thermo- and haloclines. The pure halocline is missing. This change has been labelled "the retreat of the cold halocline (layer)" (Steele and Boyd, 1998; Martinson and Steele, 2000). Such change is believed to increase the heat flux from the Atlantic layer to the ice and the atmosphere and thus have a strong impact on the thickness and maintenance of the ice cover.

From TS properties it was early realised that the water mass present between the PML and the Atlantic layer could not have formed by vertical mixing between these two waters, but had to be introduced by advection. The TS structure is most simple in the Eurasian Basin where a halocline with temperature close to freezing lies above the thermocline. In the Canadian Basin the halocline is thicker and has wider salinity (32-34.5) and density ranges. It exhibits temperature maxima and minima that reveal different source waters, e.g. the Bering Strait summer and winter inflows, shelf drainage, and advection from the Eurasian Basin (Coachman and Barnes, 1961; Jones and Anderson, 1986; Carmack, 2000). In the Eurasian Basin the halocline is more saline (S~34.3) and denser, and since a part of it advects into the Canadian Basin, it is often referred to as the "lower halocline". Consequently the Canadian Basin "halocline complex" (Carmack, 2000), centred around a nutrient maximum at the salinity of 33.1, is sometimes called the "upper halocline".

In the Eurasian Basin, excluding the outflow area north of Greenland, the Atlantic inflow constitutes the principal source of the water masses and it must, through interactions with sea ice, river runoff and net precipitation, create the PML as well as the lower halocline. Coachman and Barnes (1962) suggested that water from the Atlantic layer is brought onto the shelves, mainly along the axes of submarine channel like the St Anna Trough. It becomes cooled and diluted by mixing with low salinity shelf water. Its density is reduced, and as it returns to the deep basin it enters between the PML and the Atlantic layer, providing water for the temperature minimum observed at about 100m in the Nansen basin.

Another process active on the shelves has also been proposed as the source for the waters of the halocline. The shallow depths of the marginal seas allow the winter haline convection to reach the bottom, and the salinity of the bottom water increases throughout the winter. A cold, fairly saline water mass is created which, as it eventually crosses the shelf break and enters the deep basins, intrudes between the PML and the Atlantic layer (Aagaard et al., 1981). Coachman and Barnes (1962) as well as Aagaard et al. (1981) assumed that the sources of the lower halocline were located in the Barents Sea and the Kara Sea, the only shelf seas having sufficiently high initial salinity.

A shelf source is, however, not necessary to explain the temperature minimum and the structure of the upper layers in the Eurasian Basin. The Atlantic water entering the Arctic Ocean through Fram Strait encounters and melts sea ice north of Svalbard. This changes the upper part of the Atlantic water into a less saline surface layer which during winter becomes homogenised down to the thermocline by haline convection. The subsequent summer ice melt creates a seasonal melt water layer which on a TS diagram takes the appearance of a halocline. Observations indicate that a similar upper ocean structure is present in the entire Nansen Basin (Figure 2), suggesting that the less saline upper layer is advected above the Atlantic water in the boundary current along the Siberian continental slope (Rudels et al., 1996).

The characteristics of the upper part of the St Anna Trough outflow suggest, however, that also the Barents Sea branch could, through the winter mixed layers formed in the Barents and Kara Seas, contribute to the lower halocline. The northern Barents Sea and also the northern Kara Sea (see below), even though the convection there does not reach the bottom, have a mixed layer that in winter commonly reaches salinities higher than those of the winter mixed layer in the Nansen Basin (Figure 3). Water with similar characteristics is present in the upper part of the outflow in the St Anna Trough (Figure 3) and can be followed in the inner part of the boundary current along the Siberian continental shelf. We will argue that this less dense part of the Barents Sea inflow branch is a further source of the Arctic Ocean lower halocline in addition to the interior basin source, deriving from the Fram Strait branch, and that it provides the main input to the lower halocline found north of the Alaskan shelf in the most remote part of the Canadian Basin. It can be identified north of Greenland, and it contributes to the outflowing Polar Surface Water in Fram Strait and it may also supply the deep and bottom waters in Baffin Bay which display similar properties.

2 Observations

2.1 Data used

The data examined are CTD observations from different cruises conducted with icebreakers and with ice strengthened vessels in the last 20 years (Table 1, Figure 1). Taken together they offer a fairly complete coverage of the Eurasian Basin. The observations are, however, spread over considerable time. The Arctic Ocean is a more

Figure 1: Map of the Arctic Ocean showing the positions of the discussed stations. The different cruises are identified. The colours of the stations are the same as in the other figures. Black indicates the presence of a lower halocline derived from the Fram Strait inflow, blue a lower halocline originating from the Barents Sea inflow branch. Stations in red show water entering the Arctic Ocean north of the Yermak Plateau. Yellow is used for especially mentioned stations and green indicates stations in the Barents Sea and in the St Anna Trough influenced by Arctic Ocean water masses. The 200m (black) and 2000m (blue) isobaths are shown.

variable ocean then previously assumed, and some more permanent changes may also take place. Difficulties in separating spatial and temporal variability may therefore rise. Stations where the lower halocline derives from the Fram Strait branch are shown in black. Stations having a Barents Sea branch lower halocline are shown in blue.

Table 1. List of expeditions:

YEAR	EXPEDITION	SYMBOL
1980	Ymer-80 (HMS Ymer) (80)	✉
1984	ARKII (RV Polarstern) (84)	⌨
1991	Arctic Ocean-91 (IB Oden) (91)	📪
1993	ARKIX (RV Polarstern) (93)	✂
1993	Larsen-93 (CCG Henry Larsen) (L93)	✇
1993	Arctic Radiation Study (ArcRad)	🕮
1994	Arctic Ocean Section (CCG Louis S. St Laurent) (94)	📭
1995	ARKXI (RV Polarstern) (95)	🗄
1996	ARKXII (RV Polarstern) (96)	✌
1997	ARKXIII (RV Polarstern) (97)	☜
1997	Joint Ocean Ice Study (JOIS)	🖱

2.2 The interior Nansen Basin

Observations show temperatures close to freezing over the larger part of the Nansen Basin (Figure 2), implying a cooling of the upper layer every winter, resetting the temperature by local convection. Seasonal heating from above and the weaker upward mixing of warmer water from the Atlantic layer below create a temperature minimum which roughly indicates the depth of last winter's convection. (Figure 2). The convection depth is commonly somewhat deeper than 100m, and the salinity of the homogenised layer is around 34.3. Spatial and temporal variations are, however, present. Apart from the 20m thick low salinity melt water layer present during the summer months the communication between the sea (ice) surface and the thermocline is not broken in the Nansen Basin.

The local winter convection occasionally reaches larger depths. On the 1996 Polarstern cruise to the central Eurasian Basin (Figure 1) stations taken in August in the northern Nansen Basin, over the Nansen-Gakkel Ridge and in the southern Amundsen Basin showed a very deep (~150m) mixed layer with hardly any melt water present at the surface. The salinity was constant at ~34.3, and the temperature was close to freezing also near the surface (Figure 2). Another discrepancy is the higher salinity in the upper layer observed close to the continental slope in 1995 (Figure 2, left column, yellow station). This more saline layer could originate from the Barents Sea branch, being mixed laterally into the interior of the basin.

2.3 The Eurasian continental slope and shelves

Closer to the continental slope the core of the Atlantic layer is elevated, and the surface melt water layer is more prominent. On the Barents Sea shelf further to the south the temperature minimum, indicating the vertical limit of the local winter convection, is located between 50 and 75m. It is usually more saline (~34.5) and denser than the

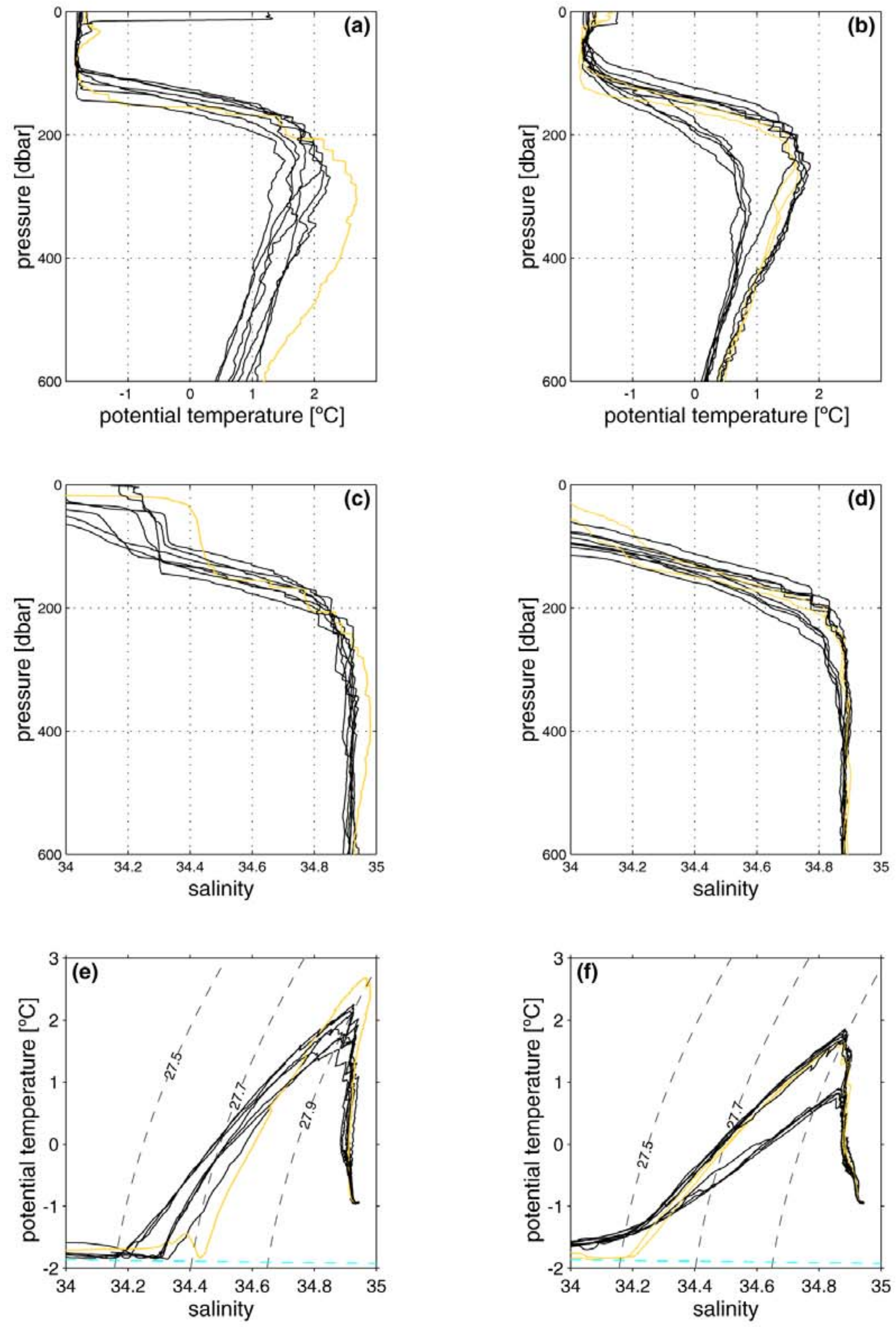

Figure 2: Potential temperature and salinity profiles down to 600m and Θ-S diagrams for stations 80/104, 91/5, 91/9, 96/43, 96/50, 95/27 (yellow), 95/75, 95/77, 93/53 in the Nansen Basin (left column), and for stations 91/47, 91/36, 91/31, 91/18, 96/55 (yellow), 96/57 (yellow), 96/58, 96/82, 95/47, 95/45, 95/44 in the Amundsen Basin (right column). The stations are listed from west to east. All stations display a lower halocline derived from the Fram Strait inflow and are shown in black, except the anomalous stations discussed in the text. The stations in the eastern part of the Amundsen Basin are warmer than in the western part because they are taken later and exhibit the warm inflow reported in the early 1990s (Quadfasel et al., 1991, Carmack et al., 1995)

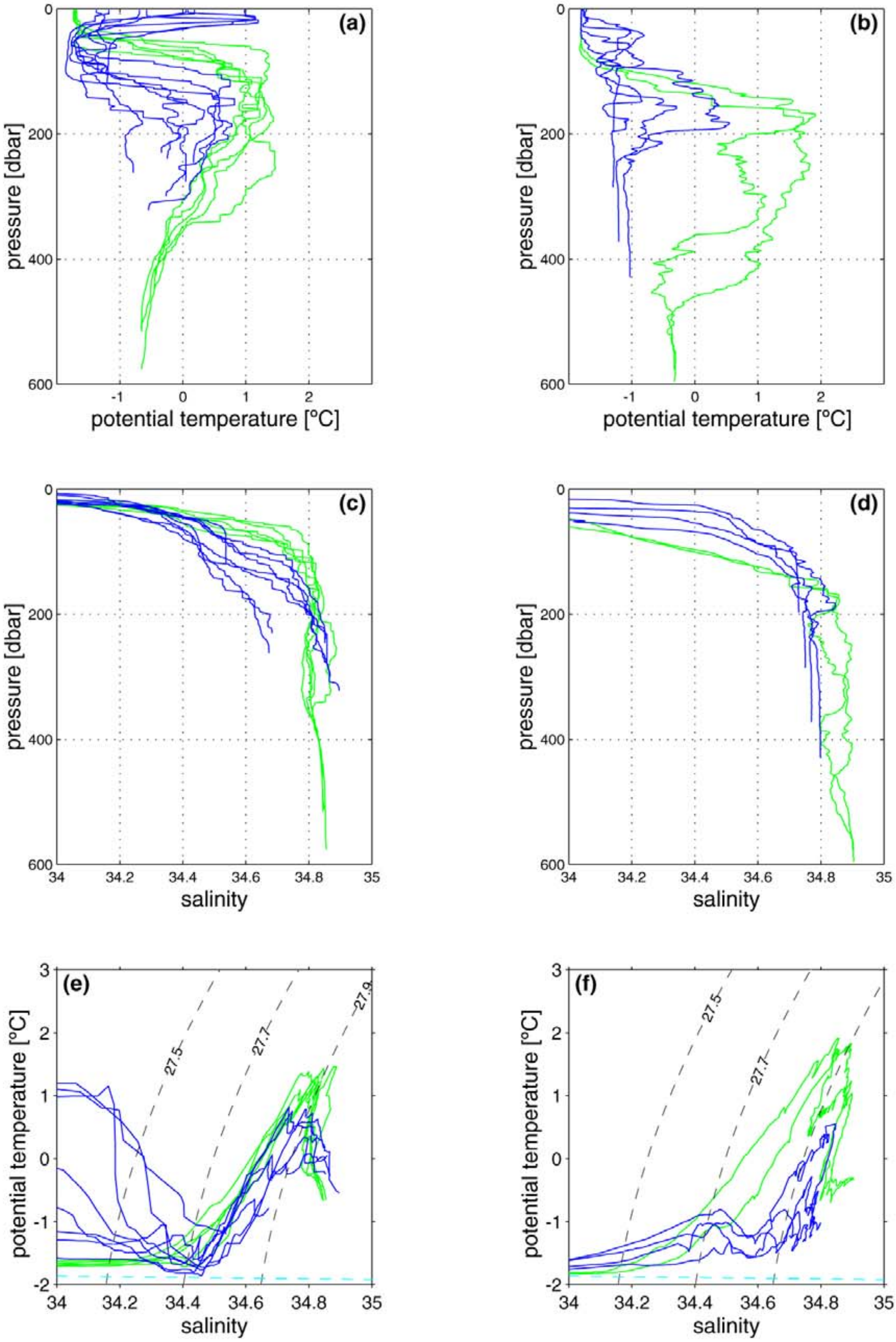

Figure 3: Potential temperature and salinity profiles down to 600m and Θ-S diagrams for stations 80/54, 80/55, 80/57, 80/59 (green) and 80/86, 80/88, 80/90, 80/91, 80/92. 80/94, 80/95 (blue) in the northern Barents Sea (left column) and stations 96/9, 96/10 (green) and 96/12 - 96/16 (blue) in the eastern St Anna Trough. Blue stations shows the embryo of the lower halocline derived from the Barents Sea inflow, created either in the Barents or the Kara Seas. Green indicates stations more influenced by Arctic Ocean water masses.

temperature minimum in the interior of the Nansen Basin (Figure 3). If this winter homogenised upper water is exported from the Barents Sea into the Arctic Ocean, it would be dense enough to penetrate into the thermocline, and into the core of the Atlantic Layer. Its salinity, however, varies from year to year, and salinities as low as 34.2 have been observed at the temperature minimum, lower than in the Nansen Basin. This salinity difference is, nevertheless, so small that an outflow from the Barents Sea entering the upper layer of the Nansen Basin could not prevent a homogenisation down to the thermocline the following winter, and such outflows would become an integral part of the Nansen Basin winter mixed layer.

The creation of the temperature minimum layer in the northern Barents Sea does not involve convection to the bottom and accumulation of saline water throughout the winter. It is a local winter homogenised upper layer, and the convection depth is restricted by the underlying stratification. It reaches a higher salinity and density than in the interior Nansen Basin because of either stronger ice formation or higher initial salinity. Convection to the bottom does take place in the Barents Sea bur further to the south, and the created, denser water mainly flows eastward into the Kara Sea. However, some dense outflows from the Barents Sea, sinking below the Atlantic Layer core, have been documented, especially in the Victoria Channel (Rudels, 1986; Schauer, 1997) (Figure 3).

The water column at the slope changes character between the Barents Sea and the Kara Sea east of the St Anna Trough, where the Barents Sea branch enters the Nansen Basin. The warm Atlantic core of the Fram Strait branch is displaced from the slope, but the cold layer above the core remains thinner and the thermocline lies shallower than further into the Nansen Basin and over the Nansen-Gakkel Ridge. The structure of the Barents Sea branch water column, now present at the slope, is more complicated (Figures 3 and 4).

The water masses observed at the St Anna Trough are similar to those found in the northern Barents Sea (Figure 3). This is to be expected since especially the densest contributions to the St Anna outflow are created in the Barents Sea (Midttun, 1985). Also the upper, less dense, temperature minimum, deriving from winter convection, has similar salinity (~34.5) as in the northern Barents Sea. However, the shelf water of the Kara Sea also experiences a seasonal cycle, and the upper temperature minimum is more likely to be a remnant of winter convection in the Kara Sea than in the Barents Sea.

The outflow at the St Anna trough is much stronger than that in the northern Barents Sea and the Barents Sea branch enters as a complete water column. This gives its upper part the possibility to initially remain separated from the winter mixed layer in the Nansen Basin. As the Barents Sea branch enters the boundary current its TS characteristics are located within the envelope of the Fram Strait branch TS curves (Figure 4). The temperature maximum is colder and less saline, as are the deeper layers, while the upper temperature minimum and the thermocline are more saline than in the Fram Strait branch. As the boundary current flows eastward, the waters of the two inflow branches gradually mix, especially in the Atlantic and deeper layers. In the upper part the temperature of the Barents Sea branch temperature minimum increases, while its salinity remains fairly constant, and the minimum gradually becomes transformed into a break (Figure 4 and 5).

Woodgate et al. (2001) noticed, from year-long moored seacat measurements as well as from hydrographic observations on the continental slope north of the Laptev and East Siberian Seas, the differences in characteristics of the halocline. They also contributed these differences to two different sources of halocline water, a colder, "convective" source and a warmer "advected" source located on the Barents Sea shelf.

As the boundary current crosses the Lomonosov Ridge, the break in the TS curves in the boundary current lies at the straight thermocline part of the interior Makarov Basin TS curves. Since waters from both branches enter

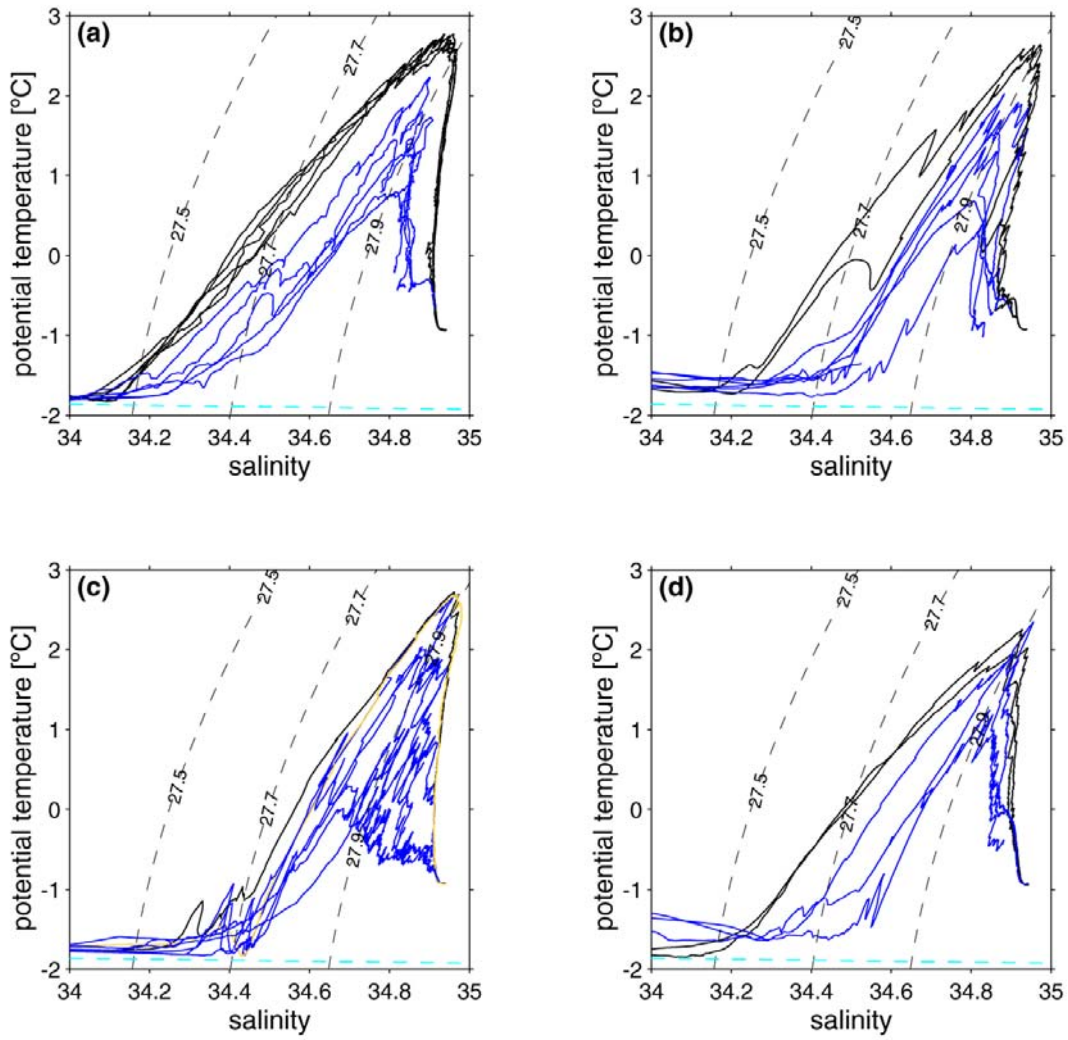

Figure 4: Θ-S diagrams for stations on the Eurasian continental slope north of the eastern Kara Sea and Severnaya Zemlya. a) Stations 96/30-96/34 (blue) and 96/35-96/39 (black), b) stations 95/95, 95/93, 95/92, 95/91, 95/88 (blue) and 95/89, 95/90 (black), c) stations 95/25, 95/29-95/32 (blue) and 95/26 (black), 95/27 (yellow), d) stations 95/78 - 95/80 (blue) and 95/77, 95/75 (black).

the Makarov Basin this implies that the temperature of the halocline in the Barents Sea branch increases more rapidly than in the halocline derived from the Fram Strait branch. The temperature increase cannot be due to lateral mixing with the waters of the Fram Strait branch but must be caused by entrainment of warmer water from below. Furthermore, the higher temperature also indicates that the winter convection occurring at the continental slope does not reach deep enough to cool the halocline. This could partly be caused by the larger ice thickness encountered in the ”ice massif” normally present north of Severnaya Zemlya. Further to the east the outflow of low salinity shelf water from the Laptev Sea cuts the communication with the sea surface for both inflow branches (see below).

The larger temperature increase in the halocline in the Barents Sea branch may partly reflect a thinner layer, but the observations indicate that after the intense mixing between the branches has abated the halocline at the slope constitutes a reasonably thick (100m) layer (Figure 6). However, it also suggests a stronger upward vertical heat flux over the slope in spite of the underlying Atlantic layer there being colder than further into the basins. The halocline is decoupled from the surface processes, at the slope as well as in the interior of the basins, and a larger heat flux is likely due to larger velocities and higher turbulent activity closer to the slope.

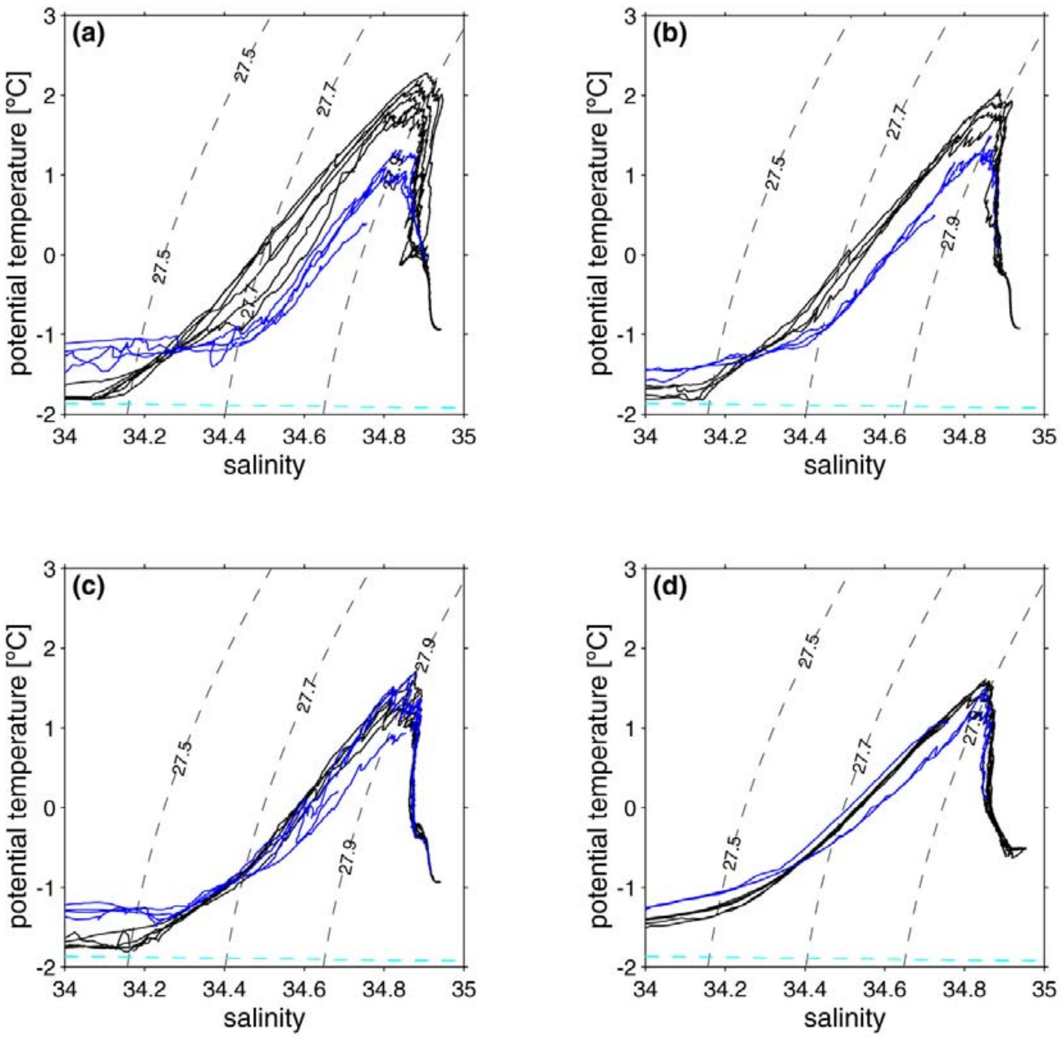

Figure 5: Θ-S diagrams for stations on the Eurasian continental slope north of the Laptev and East Siberian Seas. a) Stations 93/54-93/60 (black) and 93/61-93/65 (blue), b) stations 93/47-93/49 (blue) and 93/46, 93/50-93/52 (black), c) stations 96/99-96/104 (black) and 96/105 -96/107 (blue), d) stations 95/57-95/62 (black) and 95/63-95/65 (blue).

2.4 The Amundsen Basin and the Lomonosov Ridge

In the Amundsen Basin the halocline mainly derives from the Fram Strait branch. In the eastern part of the basin and, at least in 1991, also in the western part of the basin it was encountered as a proper halocline, covered by the Polar Mixed Layer (Figure 2). The higher temperatures in the Atlantic layer in the eastern part of the Amundsen Basin is due to the recent inflow of warmer Atlantic water to the Arctic Ocean (Quadfasel et al., 1991)

The lower salinity at the uppermost part indicates that a freshwater input occurs, or that less saline water is spreading at the surface. The Atlantic water entering the Barents Sea is not only transformed into dense water sinking down the St Anna Trough (Rudels et al., 1994; Schauer et al., 1997). Through mixing with waters of the Norwegian Coastal Current and the river runoff and because of the freezing and melting cycle of sea ice, it is also transformed into low salinity shelf water. As shelf water it spreads along the inner route south of Severnaya Zemlya and the East Siberian Islands into the Laptev and East Siberian Seas (Rudels et al., 1999). Some of this shelf water crosses the

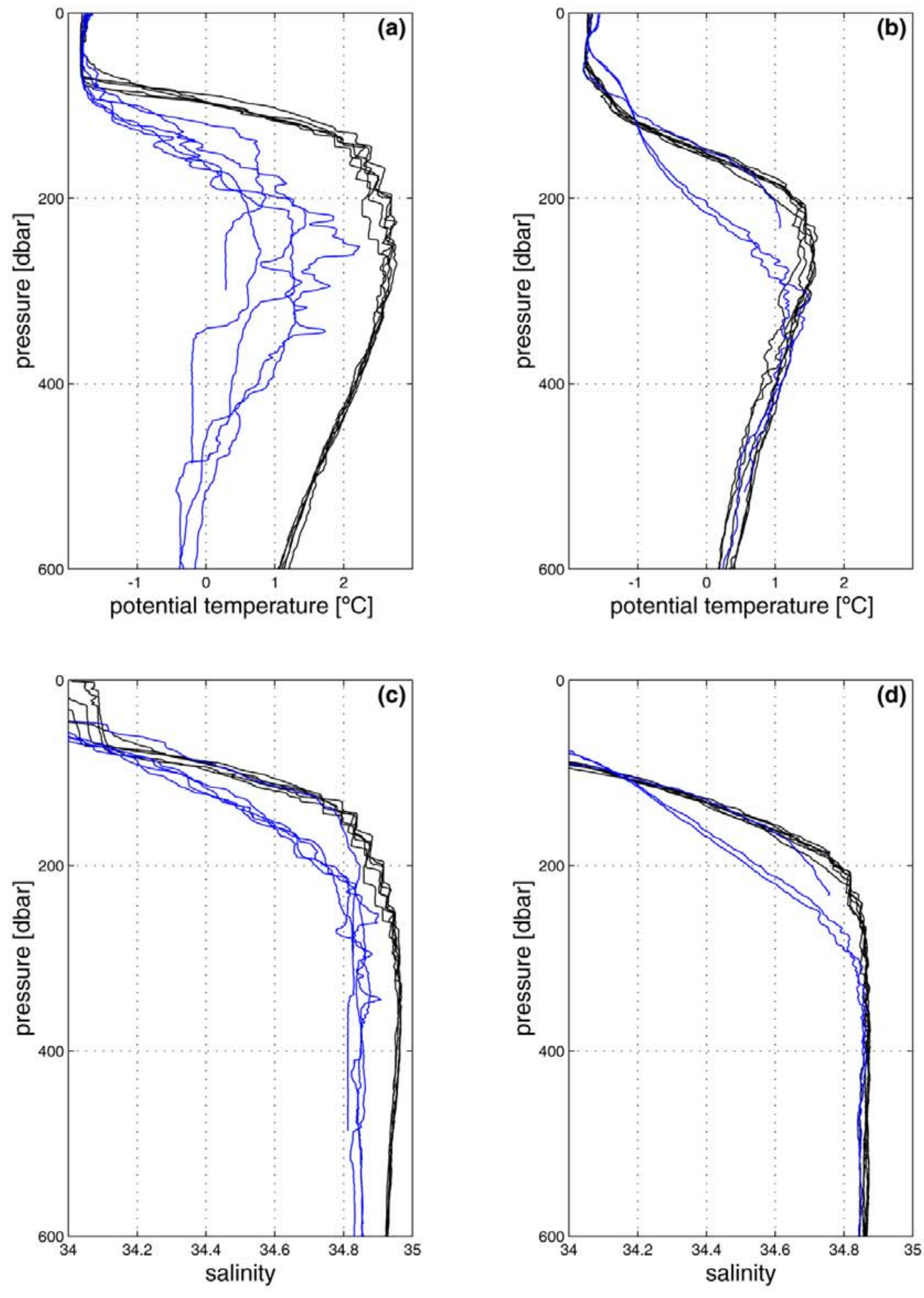

Figure 6: Potential temperature and salinity profiles down to 600m for stations 96/30-96/34 (blue) and 96/35-96/39 (black) north of the eastern Kara Sea (left column), and stations 95/57-95/62 (black) and 95/63-95/65 (blue) north of the western East Siberian Sea (right column).

shelf break of the Barents and Kara Seas, but the main part reaches as far east as the Laptev Sea before a substantial outflow across the shelf break occurs. Its salinity is here so low that the ice formation in the basin in winter cannot remove enough freshwater to maintain the convection to the thermocline. Winter convection becomes limited to the injected shelf water layer, which becomes the Polar Mixed Layer, while the denser layer below, constituting the earlier, deeper winter mixed layer is transformed into a permanent halocline (Rudels et al., 1996).

In the central part of the Amundsen Basin, close to the North Pole, the protective, low salinity Polar Mixed Layer present in 1991 was reported absent in 1995 (Steele and Boyd, 1998). In 1996 (Figure 2, right column, yellow stations) the central part of the Amundsen Basin also showed a cold, saline upper mixed layer. However, the depth of the layer varied from more than 100 to less than 60 suggesting that internal wave and eddy motions also contribute to the observed deepening of the upper layer. The area with a deep winter homogenised upper layer then extended from the Nansen Basin into the Amundsen Basin and up to the Lomonosov Ridge. Similar changes did not occur closer to the Eurasian continent, where a distinct halocline was observed in 1995 and 1996 (Figure 2).

2.5 The Makarov Basin

Water from both branches crosses the Lomonosov Ridge, providing two sources for the lower halocline in the Canadian Basin. The two contributions can be distinguished as far east as the Chukchi Cap. The observations made during the AOS94 expedition showed that the waters in the interior Makarov Basin were supplied by a branching of the boundary current at the Mendeleyev Ridge. This is the path for the warmer Atlantic water of the Fram Strait branch, but lenses of colder, less saline water were also observed in the interior of the Makarov Basin at about 1000m, showing that Barents Sea branch intermediate water was detached from the slope (Carmack et al., 1997; Swift et al., 1997). The halocline was cold and with a salinity of 34.3 (Figure 7). This indicates that the major contribution to the halocline in the Makarov Basin is from the Fram Strait branch, while the more saline and warmer Barents Sea branch halocline remains at the slope.

The 1994 observations showed that the halocline in the Makarov Basin had similar temperatures close to the Lomonosov Ridge as at the Mendeleyev Ridge. This suggests that the halocline water circulates around the Makarov Basin, or that a separation from the boundary current also occurs at the American side of the Lomonosov Ridge. The small temperature difference indicates that such input then also derives from the Fram Strait branch. The warmer Barents Sea branch halocline was not observed in the interior of the Makarov Basin.

2.6 The Canada Basin

The Fram Strait branch is not observed either at the slope or in the interior Canada Basin beyond the Chukchi Cap, and between the Mendeleyev Ridge and the Chukchi Cap the Barents Sea branch halocline becomes more prominent (Figure 7). A splitting of the boundary current at the Chukchi Cap has been suggested from the characteristics of the Atlantic and intermediate layers (Smethie et al., 2000; Smith et al., 1999). A weakening of the Fram Strait branch halocline at the slope is consistent with this view. The Fram Strait branch, also the lower halocline, would then be confined to the part of the Canada basin closest to the Mendeleyev and Alpha Ridges, while the Barents Sea branch waters continue in the boundary current along the slope and supply the most remote part of the Arctic Ocean, as seen from the Atlantic Ocean (Figure 8). The mixing in the boundary current has removed much of the difference between the two branches and some contribution from Fram Strait will also be present. Surprisingly the largest contrast is found in the halocline.

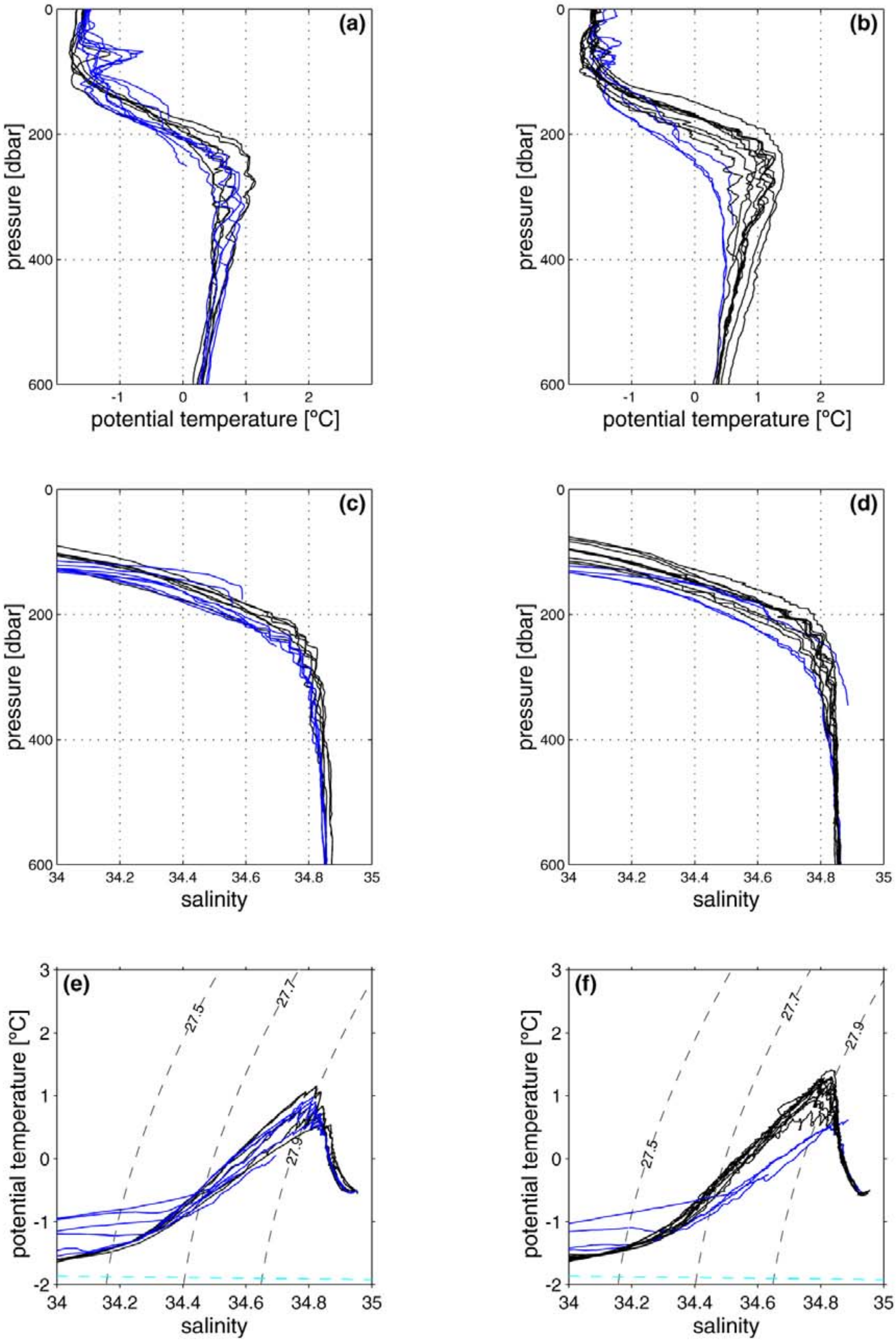

Figure 7: Potential temperature and salinity profiles down to 600m and Θ-S diagrams for stations from the Makarov Basin and the Eurasian continental slope between the Mendeleyev Ridge and the Chukchi Cap. Left column: stations 94/29, 94/26, 94/23, 94/20, 94/16, 94/12 (black) and 94/5 - 94/11 (blue). Right column: stations L93/50, L93/65, L93/67, L93/87 (blue) and L93/67, L93/69 - L93/71, L93/75, L93/81 - L93/85 (black).

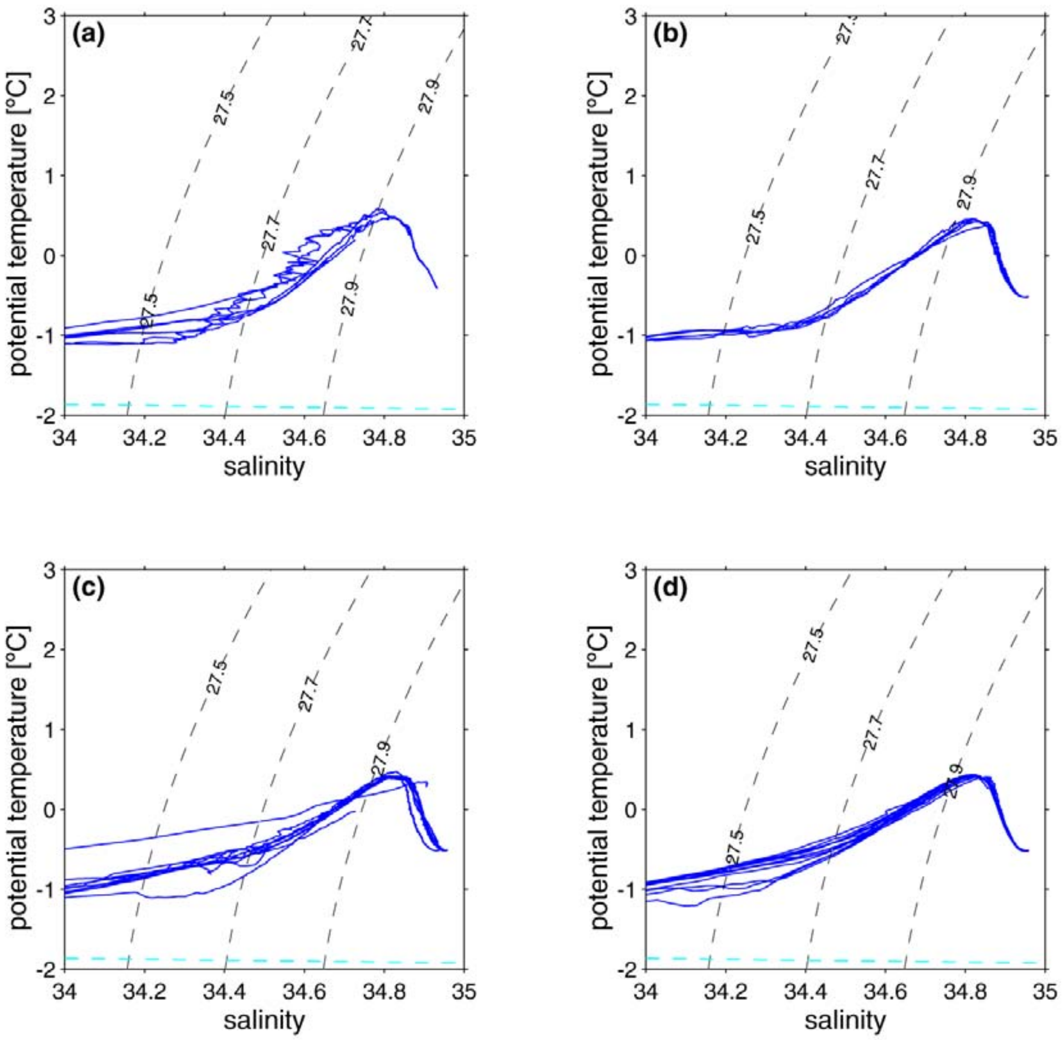

Figure 8: Θ-S diagrams for stations at the continental slope north of the Chukchi Sea and the Alaskan shelf. a) Stations ArcRad C/1 - ArcRad C/7 at the western Chukchi Cap (blue), b) stations ArcRad E/1 - ArcRad E/12 at the eastern Chukchi Cap (blue), c) stations L93/4 - L93/14, L93/98, L93/99, L93/106 north of the Alaskan Shelf (blue), d) stations JOIS97/1 - JOIS97/15 north of the eastern Alaskan shelf (blue).

The Pacific inflow enters the Canada Basin above the Atlantic derived lower halocline, increasing the thickness of the upper layers. This further distinguishes the Canada Basin from the rest of the Arctic Ocean, especially when compared to the Nansen Basin with its absence of a permanent, cold halocline.

2.7 The outflow area north of Fram Strait

The return flow toward Fram Strait occurs in the boundary current along the slope, crossing the Alpha and the Lomonosov Ridges. The waters from the different loops of the different basins come together at the Greenland continental slope between the Morris Jesup Plateau and Fram Strait. The return flow from the Canada Basin is located closest to Greenland and the contributions from the other basins line up parallel to the slope as one moves further into the basin, in the Atlantic and intermediate layers as well as in the surface waters (Rudels et al., 1994).

The observations closest to the Greenland coast and farthest to the north were made in 1980, 1984, 1991 and 1997. There are differences between the years. The Barents Sea branch halocline was more clearly identified in

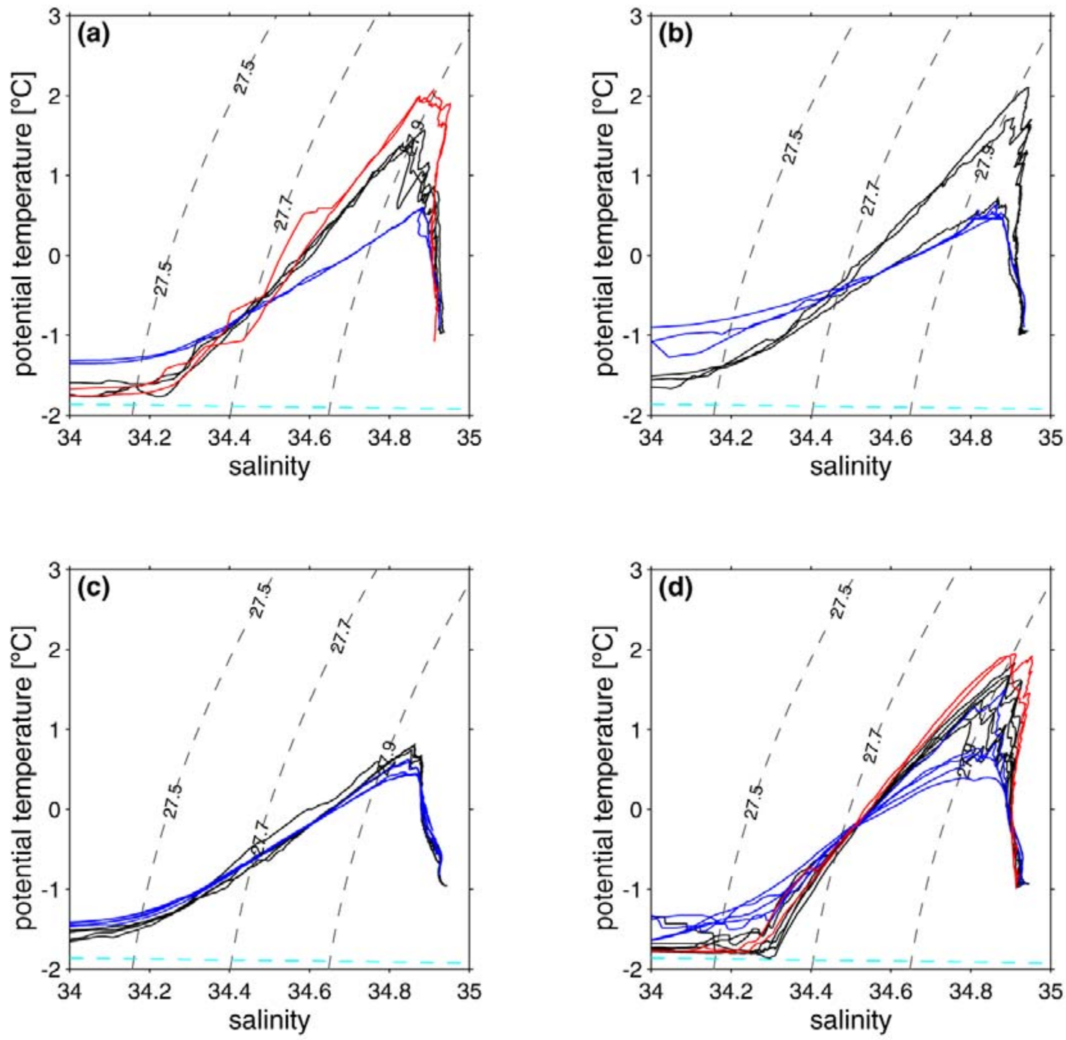

Figure 9: Θ-S diagrams for stations north of Fram Strait showing the different outflowing gyres and in red the inflowing water north of the Yermak Plateau. a) stations 80/164, 80/162 (blue) and 80/166, 80/168, 80/170 (black), and 80/171, 80/172 (red), b) stations 84/333 - 84/335 (black) and 84/29 - 84/331 (blue), c) 91/39 - 91/41, 91/43, 91/44 (blue) and 91/38, 91/45 - 91/47 (black), d) stations 97/68 - 97/70, 97/77, 97/78, 97/82 (black), and 97/72 - 97/76 (blue), and 97/86, 97/87 (red).

1984 than in 1997 (Figure 9). In some years the Canadian Basin upper waters, e.g. the upper halocline, were more prominent than in others. The Fram Strait branch was present in all years. Mostly it formed a proper halocline between the thermocline and the Polar Mixed Layer. However, in 1997 the Polar Mixed Layer was absent at some stations and a deep, saline upper layer with temperatures close to freezing was observed north-west of the Yermak Plateau (Figure 10). This suggests that a return flow of the saline deep mixed layer, observed in the interior Eurasian Basin in 1995 and 1996, was taking place. At one station the isothermal layer reached as deep as 240m, while the salinity showed a three layer structure, indicating that different streams converge north of Fram Strait and that the densest upper layer is depressed downward (Figure 10). Similar, but less extreme, convergences could occur in the interior of the Arctic Ocean where the deepest observed ventilated layers were less dense than the thermocline at the same depth on the neighbouring stations.

As the water of the lower halocline moves along the continental slope from the Canada basin towards the northern coast of Greenland, it rises in the water column from 300m north of Alaska to about 120m north of

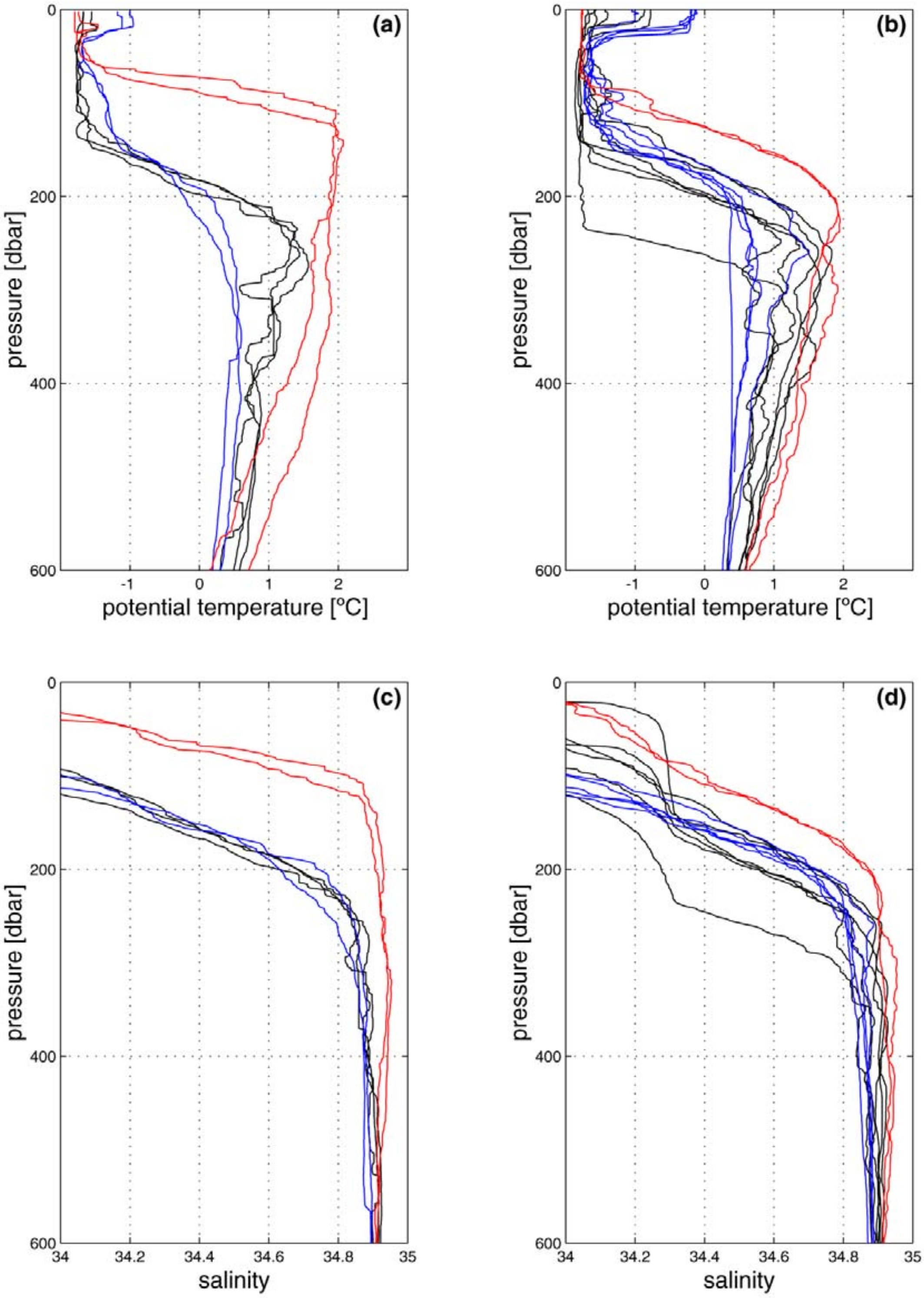

Figure 10: Potential temperature and salinity profiles down to 600m for stations north of Fram Strait showing the different outflowing gyres and in red the inflowing water north of the Yermak Plateau. Left column: stations 80/164, 80/162 (blue) and 80/166, 80/168, 80/170 (black), and 80/171, 80/172 (red), right column: stations 97/68 - 97/70, 97/77, 97/78, 97/82 (black), and 97/72 - 97/76 (blue), and 97/86, 97/87 (red).

Greenland. This rise is due to the gradual draining of the layers above, mainly of Pacific origin, through the Canadian Arctic Archipelago into Baffin Bay. As the upper waters are removed the lower halocline migrates upwards in the boundary current.

Since water with lower halocline characteristics is located shallower than 200m north of Greenland it lies above the sill depth of Smith Sound and it thus constitutes a possible source for the deep and bottom waters of Baffin Bay. That the characteristics of the deeper layers in Baffin Bay are similar to those found at 250m in the Beaufort Sea was noticed by Bailey (1956) and since this is at the level of the sill in Smith Sound he suggested the Arctic Ocean as the source for the deep waters of Baffin Bay. The fact that this water mass is located much shallower north of Greenland, closer to Smith Sound, further supports Bailey's suggestion.

Redfield and Friedman (1969) found from Deuterium observations that the deeper layers of Baffin Bay showed traces of brine rejection, which suggested that ice formation was one process that could increase the density of the Baffin Bay deep water. Local haline convection could then also contribute to the deep water in Baffin Bay. However, the brine rejection could have occurred earlier and elsewhere. The formation of the winter mixed layer of the Barents Sea branch inflow involves brine rejection and this brine rejection is not overcompensated by an initial ice melt as is the case with the winter mixed layer in the Nansen Basin north of Svalbard. The freshening of the upper part of the Barents Sea branch could partly be due to the excess freshwater of the Norwegian Coastal Current, partly be caused by the river input from Ob and Yenisei. This further supports the Barents/Kara Sea as a source not only for the lower halocline in the Canada Basin but also the deeper layers of Baffin Bay. The part that does not exit through Smith Sound would continue south and join the East Greenland Current on the eastern Greenland shelf.

It should be pointed out that Bourke et al. (1989) and Bourke and Paquette (1991) did not observed any continuous flow of Arctic halocline water through Smith Sound into Baffin Bay in 1986, and they proposed that local dense water formation and convection contribute to the Baffin Bay deep and bottom water. If the lower halocline water is the most important source for the deep water in Baffin Bay, the inflow must be intermittent and the entrainment into the inflowing water must also be small since the ambient Baffin Bay waters are warmer and a denser contribution from the Arctic Ocean also involves water of higher temperatures.

3 Summary

Different possible sources for the lower halocline have been examined. It is found that the water of the lower halocline in the Amundsen, Makarov and part of the Canada Basin derives from the winter mixed layer of the Nansen Basin and the Fram Strait branch, while the lower halocline at the continental slope and in the remoter part of the Arctic Ocean (seen from the Atlantic Ocean) originates from the upper part of the Barents Sea branch entering through St Anna Trough. By contrast, the upper temperature minimum in the Eurasian Basin, often attributed to outflow from the Barents and Kara Seas, derives from the Fram Strait branch.

The creation of the Barents Kara Sea lower halocline does not involve either of the two commonly proposed shelf processes for halocline formation: Cooling and freshening of Atlantic water brought from the interior of the Arctic Ocean onto the shelves, or the accumulation of saline water on the bottom of the shelves during winter due to freezing and brine rejection. Both the Fram Strait branch and the Barents Sea branch contributions rather involve the winter homogenisation of an upper, less saline layer above the Atlantic water entering from the Norwegian Sea. In the Arctic Ocean the Barents Sea branch source is recognised by a higher salinity and temperature than the Fram Strait branch source, due to a stronger entrainment of warmer water from below. It is the possible source for the deep and bottom waters in Baffin Bay, where it would enter through Smith Sound. The remaining part of the Barents Sea branch derived lower halocline water exits through Fram Strait as a part of the Polar Surface Water of the East Greenland Current.

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