Atlantic sources of the Arctic Ocean surface and halocline waters

Bert Rudels, E. Peter Jones, Ursula Schauer & Patrick Eriksson

Data obtained during the last 20 years from selected hydrographic stations throughout the Arctic Ocean basins and at the continental slopes and northern parts of the surrounding shelf seas are examined to estimate the contribution of Atlantic water to the upper part of the Arctic Ocean water column, and to follow the circulation of the Atlantic derived halocline waters around the Arctic Ocean basins. A substantial fraction of the Atlantic water entering the Arctic Ocean in the two inflow branches, the Fram Strait branch and the Barents Sea branch, is transformed into less dense waters. The inflow through Fram Strait encounters and melts sea ice. Its upper part becomes less saline, and in winter, ice formation homogenizes this low salinity upper part into a winter mixed layer, which follows the boundary flow eastward. The inflow over the Barents Sea mainly becomes colder, less saline and denser due to cooling and net precipitation, but melting of sea ice also here creates a less dense upper layer, which is subsequently homogenized by haline convection. Both these components enter the Nansen Basin via the St. Anna Trough and flow eastward along the Siberian continental slope. A third component of the Barents Sea inflow, mainly comprising Norwegian Coastal Current water, remains on the shelf, where it absorbs the runoff from the large Siberian rivers and evolves into low salinity shelf water. This part continues as far east as the Laptev Sea before a major inflow to the deep Arctic Ocean basins occurs. Winter convection, which in most of the Nansen Basin extends to the Atlantic layer, then becomes limited to the injected low salinity shelf water, and the more saline upper layers of the Fram Strait and the Barents Sea branches are transformed into halocline waters. The Fram Strait branch supplies the halocline water of the Amundsen, Makarov and part of the Canada basins, while the upper part of the Barents Sea branch, initially confined to the Siberian continental slope, becomes the main source of the lower halocline below the Pacific water in the Canada Basin beyond the Chukchi Cap. Less dense Pacific water close to the North American continent prevents the Atlantic derived upper layers from flowing through the Canadian Arctic Archipelago and their main exit is Fram Strait. Only some of the Barents Sea branch halocline water passes through Nares Strait and may occasionally renew the deep and bottom waters of Baffin Bay.

B. Rudels & P. Eriksson, Finnish Institute of Marine Research, PO PL33, FI-00931 Helsinki, Finland, Bert.Rudels@fimr.fi; E. P. Jones, Ocean Science Division, Bedford Institute of Oceanography, 1 Challenger Drive, PO Box 1006, Dartmouth NS, Canada B2Y 4A2; U. Schauer, Alfred Wegener Institute for Polar and Marine Research, PO Box 120161, D-27515 Bremerhaven, Germany.

The upper 500 m of the Arctic Ocean basins are strongly stratified. A low salinity layer, the polar mixed layer (PML), overlies a salinity-stratified pycnocline whose upper part has temperatures close to freezing. In the lower part of the pycnocline there is a destabilizing thermocline as the temperature rises to the subsurface temperature maximum of the Atlantic layer. In summer the PML becomes stratified as the melting of sea ice creates a surface layer of still lower salinity. In winter this layer is removed by ice formation, brine release and subsequent haline convection. In summer the depth of the local winter convection is approximately indicated by a temperature minimum (Fig. 1). The pycnocline between the PML and the Atlantic layer is commonly called the halocline (Coachman & Aagaard 1974). It is typically 150 m thick, about three times the thickness of the PML.

Profiles of salinity and potential temperature, Θ, as well as ΘS curves, representative of each of the four major Arctic Ocean basins, are shown in Fig. 1. There are large differences between the basins. In the Canada Basin the PML is less dense and the halocline much thicker, spanning a larger density range than in the other basins. This is caused by the inflow of low salinity ($S \sim 32.5$) Pacific water through the Bering Strait (Coachman & Barnes 1961), and to some degree by runoff from the Mackenzie River. The Pacific inflow comprises the Bering Strait summer water and Alaskan coastal water, recognized by temperature maxima, and the Bering Strait winter water, forming a temperature minimum at 150 - 200 m (Coachman & Barnes 1961; Coachman et al. 1975). In the same density range as the winter water, the Pacific water also supplies the upper halocline, characterized by salinity around 33.1 and a nutrient maximum (Jones & Anderson 1986). The nutrient maximum arises as a result of contact with sediments and the uptake of remineralized nutrients on the Chukchi Shelf, implying that the water has experienced a winter of ice formation and convection to the bottom before it enters the deep Canada Basin (Jones & Anderson 1986). Recent work on the Bering Strait inflow and the distribution of Pacific water in the Canada Basin has revealed further details of the paths of the Pacific water across the Chukchi Shelf and in the Arctic Ocean (Jones, Anderson et al. 1998; Shimada et al. 2001; Steele et al. 2004). The Pacific water mainly leaves the Arctic Ocean through the Canadian Arctic Archipelago, but its movements across the Arctic Ocean appear to be determined by the meteorological forcing, such as the state of the Arctic Oscillation and the North Atlantic Oscillation and the two circulation modes (e.g. Proshutinsky & Johnson 1997) of the Arctic Ocean (Steele et al. 2004).

The Pacific inflow rarely supplies water to the interior Arctic Ocean more saline than 33.5, even if water entering through western Bering Strait occasionally has salinities above 34 (Roach et al. 1995) and salinities above 36 have been observed in the Barrow Canyon (Aagaard et al. 1985). Instead, the part of the halocline underlying the upper halocline in the Canada Basin as well as the more saline upper layers in the other basins derive from the Norwegian Sea and ultimately from the North Atlantic. In the Amundsen and Makarov basins a smooth, but distinct, halocline is encountered below the temperature minimum, while in the Nansen Basin the temperature and salinity stay almost constant until they both rapidly increase towards the temperature maximum of the Atlantic layer (Fig. 1). This suggests that no permanent halocline is present in the Nansen Basin, and, when the seasonal halocline is removed in winter, the upper layer is homogenized down to a pycnocline with coinciding thermocline and halocline lying above the Atlantic water, forming a "winter mixed layer" (Rudels, Anderson et al. 1996).

The heat content in the Atlantic layer, defined as water with temperatures above 0 ºC, is large enough to melt several metres of ice, should it be entrained into the PML. The halocline prevents this from happening. In view of recent reports of the diminishing of the Arctic Ocean ice cover (McPhee et al. 1998; Rothrock et al. 1999; Winsor 2001; Holloway 2001), of increased temperature of the Atlantic layer (Quadfasel et al. 1991; Carmack, MacDonald et al. 1995; Morison et al. 1998), and of a weakening of the two-layer structure of the PML and the halocline, the so called "retreat of the cold halocline layer" (Steele & Boyd 1998; Martinson & Steele 2001), an understanding of the formation and spreading of the Atlantic derived "lower" halocline waters is fundamental in predicting how the Arctic Ocean may respond to climate change.

Because of its initially high salinity, Atlantic water must be diluted by ice melt and/or by net precipitation and river runoff before it can renew the upper waters of the Arctic Ocean. The present work describes the areas and the proc-

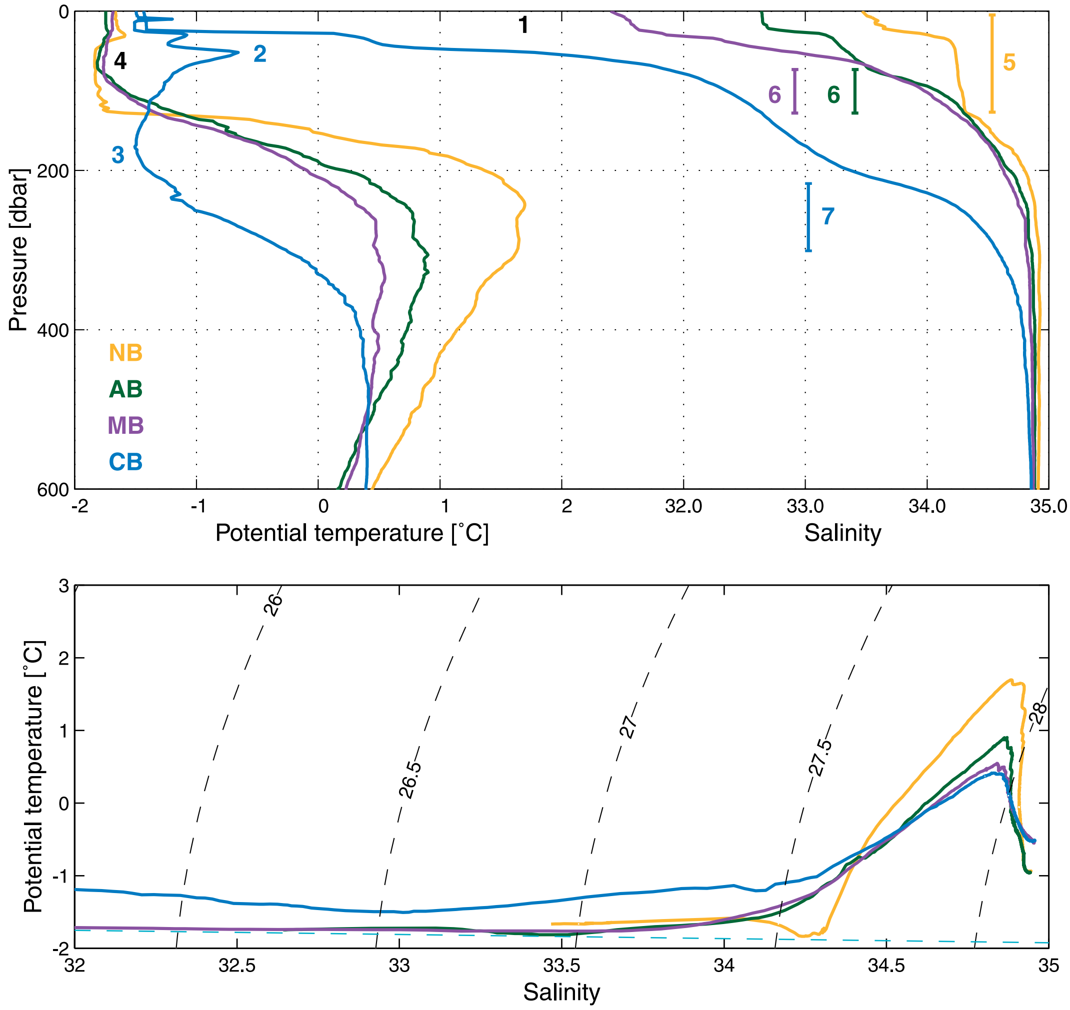


Fig. 1: The characteristics of the upper layers in the different basins in the Arctic Ocean. The upper panel shows the potential temperature and salinity profiles and the lower panel the ΘS curves for the Nansen Basin (NB, Oden/5 yellow), Amundsen Basin (AB, Oden/18 green), Makarov Basin (MB, AOS/29 violet) and Canada Basin (CB, Jois/14 blue) (see insert map). The exact positions of the stations are given in Table 2. The broken line in the ΘS diagram indicates the salinity dependence of the freezing temperature, 1) the low salinity surface water deriving from seasonal ice melt, 2) the temperature maximum of the Bering Strait summer water, 3) the temperature minimum of the Bering Strait winter water and the range of the upper halocline, 4) the temperature minimum of the winter convection, 5) the depth of the winter mixed layer in the Nansen Basin, 6) the range of the halocline in the Amundsen and Makarov basins, and 7) the range of the lower halocline in the Canada Basin.

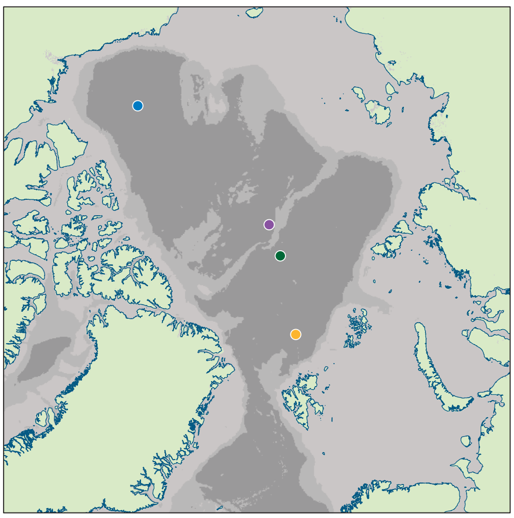


Table 1. List of expeditions:

Year	Expedition	Symbol
1980	Ymer (HMS *Ymer*)	■
1984	ArkII (RV *Polarstern*)	◀
1987	Baffin Bay (CCGS *Hudson*) (BB)	▣
1991	Arctic Ocean (IB *Oden*)	●
1991	Barents Sea (RV *Dalnye Zelitsi*) (DZ)	×
1993	ArkIX (RV *Polarstern*)	+
1993	Arctic Radiation Study (ArcRad)	◆
1994	Arctic Ocean Section (CCG *Louis S. St. Laurent*) (AOS)	▲
1995	ArkXI (RV *Polarstern*)	▼
1996	ArkXII (RV *Polarstern*)	★
1997	ArkXIII (RV *Polarstern*)	✷
1997	Joint Ocean Ice Study (JOIS)	▶

esses in the Arctic Ocean and in the Barents and Kara seas, where the inflow from the Norwegian Sea becomes transformed into less dense water masses, and where waters in the density range of the halocline are created. We trace the paths of the upper layers of the two Atlantic inflow branches—the Fram Strait branch and the Barents Sea branch—by following the temperature and salinity characteristics as these waters evolve and make their way throughout the Arctic Ocean basins before eventually exiting through the Canadian Arctic Archipelago and Fram Strait. We use data from several cruises (Fig. 2 and Tables 1 & 2) conducted during the last 20 years in the interior of the basins as well as on the continental slopes and northern parts of the shelves. The accuracy of the measurements has increased over the years and the early cruises show much larger error bars, perhaps as large as 0.02 in salinity and 0.02 ºC in temperature. This is sufficient for a discussion of the evolution of the upper and halocline waters.

Table 2: The positions of the discussed stations.

Cruise	Station	Latitude	Longitude
AOS	8	75.45	189.42
AOS	11	76.62	186.68
AOS	23	81.58	176.88
AOS	26	84.06	175.06
AOS	29	87.15	160.71
ArcRad	C01	73.90	-168.50
ArcRad	C05	74.79	-165.17
ArcRad	E06	73.09	-158.75
ArkII	329	81.77	-10.64
ArkII	330	81.84	-10.49
ArkIX	19	82.76	40.21
ArkIX	53	79.24	122.88
ArkXI	27	81.22	106.58
ArkXI	57	81.20	150.08
ArkXI	63	79.92	149.79
ArkXI	89	82.35	93.00
ArkXI	91	82.07	90.98
ArkXII	9	81.29	72.02
ArkXII	13	81.44	74.22
ArkXII	16	81.42	74.76
ArkXII	37	82.52	92.30
ArkXII	43	84.19	100.56
ArkXII	50	85.17	109.29
ArkXII	55	85.88	121.24
ArkXII	57	86.30	130.63
ArkXII	58	86.39	134.19
ArkXII	82	82.53	132.90
ArkXIII	68	81.42	-0.97
ArkXIII	69	81.49	-5.41
ArkXIII	72	81.58	-8.19
ArkXIII	73	81.58	-9.82
ArkXIII	82	82.38	3.70
ArkXIII	86	82.12	5.04
ArkXIII	87	82.04	5.31
BB	133	71.52	-62.56
DZ	19	74.67	53.13
Jois	4	70.56	-141.72
Jois	5	70.70	-141.78
Jois	7	71.03	-141.98
Jois	12	73.74	-143.81
Jois	14	75.20	-142.35
Oden	5	83.56	27.63
Oden	18	88.18	99.13
Oden	31	88.28	9.34
Ymer	59	80.81	41.34
Ymer	90	79.35	39.86
Ymer	91	79.32	41.99
Ymer	94	79.55	46.18
Ymer	104	82.35	24.20
Ymer	162	81.72	-8.85
Ymer	164	81.68	-9.10
Ymer	166	82.02	-7.12
Ymer	168	81.72	-3.52
Ymer	170	81.67	-3.73
Ymer	171	81.40	0.87
Ymer	172	81.12	3.20

Fig. 2: The Arctic Ocean and the positions of the stations from the different cruises. Symbols for expeditions are identified in Table 1. The positions of the stations discussed explicitly in the text and presented in the different figures are shown on maps in the individual figures and listed in Table 2.

Formation and source areas of Atlantic derived halocline waters

The label "halocline", as used in the Arctic Ocean, is misleading. "Cline" implies a transition zone between two water masses, as between the PML and the Atlantic layer, while in reality several water masses from different sources, all colder than 0 °C, could be layered above each other with the least saline on top and the most saline at the bottom (see e.g. Carmack 1990). The halocline also occupies a much larger depth range, commonly between 50 m and 250 m, than the upper 50 m of the PML. In this paper, the term halocline is retained for the depth interval between the

PML and the Atlantic layer, but when different water masses within the halocline are discussed, these are referred to as “waters” and identified by their origin, i.e., Barents Sea branch halocline water and Fram Strait branch halocline water. In the situation where the thermocline and halocline coincide, forming a pycnocline with destabilizing temperature and stabilizing salinity stratification, the two-layer PML–halocline structure is not present and the term “winter mixed layer”, introduced above, is applied for the water mass above the pycnocline.

The bend in the ΘS curves at $S \sim 34.2$ and $\Theta \sim -1.2$ to -1.8 (Fig. 1) indicates that the halocline does not form through direct mixing between the PML and the underlying Atlantic water. The halocline waters must be created elsewhere and be advected to different parts of the Arctic Ocean.

For the Atlantic derived halocline Coachman & Barnes (1963) suggested that Atlantic water is brought onto the shelves along deep troughs like the Victoria Channel and the St. Anna Trough, where it is cooled by heat exchange with the atmosphere and freshened by mixing with less saline shelf water. This colder, less saline and less dense water eventually returns to the deep basins, where it intrudes between the PML and the Atlantic layer.

Aagaard, Coachman et al. (1981) proposed another shelf process for creating halocline waters. If in some areas on the shelves sea ice is constantly removed by the winds, e.g., in polynyas leeward of coasts and islands, the ice production will remain high. The released brine convects and accumulates in the water column, eventually forming a bottom layer that becomes increasingly saline as the winter progresses. This bottom layer crosses the shelf break and flows into the basins, where it enters below the PML and creates the halocline. Such a process most likely creates the Pacific upper halocline water on the Chukchi shelf (Jones & Anderson 1986).

Rudels, Anderson et al. (1996) noted that accumulation of saline water on a shallow shelf is not really needed to create waters with halocline characteristics. What is needed is a fairly dense upper layer, where winter convection is limited by the stratification of the water column and the water above the pycnocline is homogenized into a winter mixed layer, which is advected with the mean circulation and eventually becomes isolated from the sea surface by a less dense water mass and evolves into an halocline.

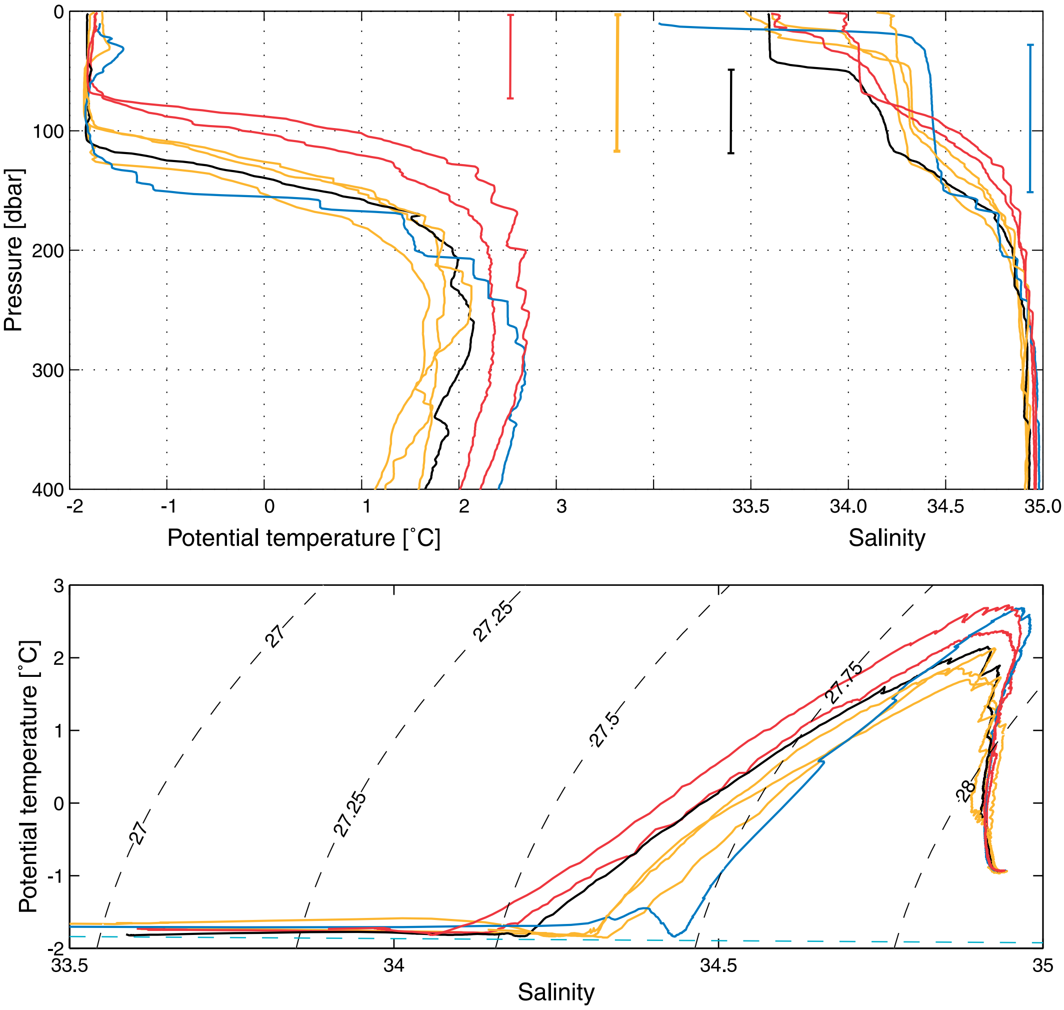
The Fram Strait branch and the Nansen Basin

The characteristics of halocline waters are already present in the winter mixed layer in the Nansen Basin. Rudels, Anderson et al. (1996) proposed that this winter mixed layer is initially formed as the Atlantic water entering the Arctic Ocean through Fram Strait encounters and melts sea ice north of Svalbard. This cooled and freshened surface water largely moves eastward with the underlying warm Atlantic core, becomes homogenized in winter and covered by less saline meltwater the following summer. This scenario repeats itself as the waters flow along the Nansen Basin as far east as the Laptev Sea, where commonly a large volume of less saline shelf water crosses the shelf break and enters on top of the winter mixed layer. Subsequently, winter convection is limited to the low salinity surface water, which becomes the PML, while the winter mixed layer is submerged and evolves into a halocline water mass. Rudels, Anderson et al. (1996) considered this the main mechanism for creating the Atlantic derived, lower halocline of the Arctic Ocean.

Winter convection in the Nansen Basin reaches about 100 m and the salinity of the homogenized winter mixed layer is 34.2 - 34.4 (Fig. 3). The convection depth is less, 50 - 100 m, closer to the continental slope (Fig. 3, red stations). Apart from the summer months, when the thin, low salinity meltwater layer is present at the surface, the communication between the sea (ice) surface and the Atlantic layer remains unbroken in most of the Nansen Basin. Only in the eastern part, close to the Laptev Sea, is a thicker, less saline surface layer present and only there has the almost constant salinity in the layer below evolved into a salinity gradient (Fig. 3, black station).

The Barents Sea branch and the Barents and the Kara Seas

The shallow Barents and Kara seas have long been considered plausible source areas for the (lower) halocline water with salinity about 34.5. Suggested mechanisms include: cooling and freshening of Atlantic water brought onto the shelves along submarine canyons (Coachman & Barnes 1963); freezing and accumulation of brine enriched water at the shelf bottom throughout the winter (Aagaard, Coachman et al. 1981; Jones &



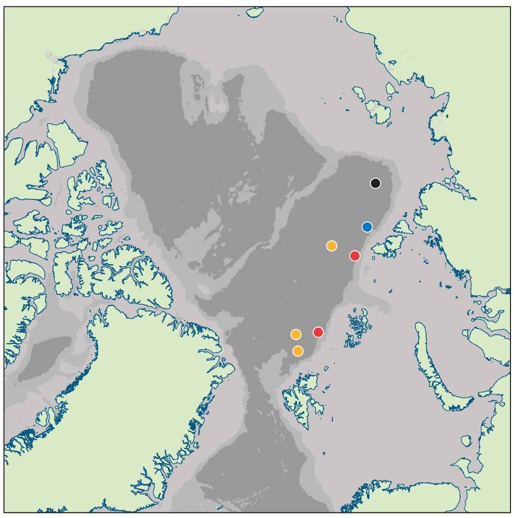


Fig. 3: The properties of the upper part of the Nansen Basin water column. The upper panel shows potential temperature and salinity profiles and the lower panel the ΘS curves from Ymer/104, Oden/5 and ArkXII/43 (yellow) from the interior of the Nansen Basin and ArkIX/19 and ArkXII/37 (red) from the southern boundary (see insert map). The yellow and red vertical bars indicate the depth of the winter mixed layer on the different stations. ArkIX/53 (black) from north of the Laptev Sea shows a thicker low salinity upper layer and an isolated winter mixed layer gradually evolving into a halocline (black vertical bar). ArkXI/27 (blue) indicates Barents Sea branch halocline water injected into a Fram Strait branch water column (blue vertical bar).

Anderson 1986; Woodgate et al. 2001); exchanges of Atlantic water across the Polar fronts, where ice melt preconditions the upper layers for ice formation and homogenization by haline convection the following winter (Rudels, Larson et al. 1991); and ice melting in the marginal ice zone (Steele et al. 1995).

The salinity of the Atlantic water entering the Barents Sea from the Norwegian Sea is initially high, and a substantial input of freshwater is needed for the surface water to attain the salinity of the halocline. This makes convection to and accumulation of brine enriched water at the shelf bottom a less likely formation process for halocline water in the Barents Sea. The density of the created bottom water would become too high and, as it continues into the Arctic Ocean, it would enter the Arctic Ocean water column at a much deeper level than the halocline. No large rivers enter the Barents Sea, and the main freshwater sources are net precipitation, the low salinity water of the Norwegian Coastal Current, runoff from Ob and Yenisey entering the Barents Sea from the Kara Sea (either as ice or as low salinity surface water), and ice drifting into the Barents Sea from the Arctic Ocean. In addition, the separator effect, i.e., ice formation over shallow areas and the accumulation of saline water at the bottom, together with export and melting of the sea ice elsewhere, also supplies freshwater to the upper part of the water column. The combined effects of all these sources and processes lead to the formation of a less saline upper layer in the northern and eastern Barents Sea.

In the northern Barents Sea a temperature minimum ($S \sim 34.45$-34.5, $\Theta \sim -1.8°C$) is located between 50 m and 75 m; this marks the limit of the local winter convection and defines the winter mixed layer in the Barents Sea (Fig. 4). The convection depth, as in the Nansen Basin, is determined by the underlying stratification, not by the depth of the shelf. The higher salinity makes the water exported from the Barents Sea dense enough to enter the pycnocline below the winter mixed layer of the Nansen Basin.

The higher salinities in the winter mixed layer could be caused by the lower temperature of the Atlantic water in the northern and eastern Barents Sea compared to north of Svalbard. During winter the air temperatures are below the freezing point of seawater, and when sea ice is present at the surface, the ocean loses sensible heat to the atmosphere as well as to the melting of sea ice. Rudels, Friedrich, Hainbucher et al. (1999) suggested that when a low salinity upper layer is formed through ice melt under such conditions, the partitioning of the oceanic sensible heat between the part going to the atmosphere and the part used for melting sea ice is such that the sea ice melt rate reaches a minimum. This results in higher salinities in the winter mixed layer than if all oceanic sensible heat is used for ice melt as suggested by Moore & Wallace (1988) and applied by Steele et al. (1995).

The fraction, f, of heat going to ice melt is then given by:

$$f = \frac{2\alpha L}{c(\beta S_A - \alpha \Delta T)} \approx \frac{2\alpha L}{c\beta S_A} \tag{1}$$

where α is the coefficient of heat expansion and $\beta = 8 \times 10^{-4}$ is the coefficient of salt contraction. $L = 336\,000$ Jkg^{-1} is the latent heat of melting, $c = 4000$ Jkg^{-1}degree^{-1} is the heat capacity of sea water, S_A is the salinity of the lower (Atlantic) layer, and ΔT is the temperature difference between the Atlantic and the upper layer. The salinity of sea ice is taken to be zero (for details see Rudels, Friedrich, Hainbucher et al. 1999).

The upper layer is a mixture of Atlantic water and ice melt and the amount of freshwater, F, added to the mass, M, of Atlantic water is:

$$F = \frac{Mfc\Delta T}{L} \tag{2}$$

Conservation of salt, $M \times S_A = (M + F) \times S_l$, gives the salinity S_l of the upper layer as:

$$S_l = \frac{S_A}{\left(1 + \frac{fc\Delta T}{L}\right)} \tag{3}$$

which, using (1), gives:

$$S_l = \frac{S_A}{\left(1 + \frac{2\alpha\Delta T}{\beta S_A}\right)} \tag{4}$$

α depends strongly upon temperature. The temperature, T_A, and salinity, S_A, of the Atlantic water north of Svalbard are around 2.5 ºC and 34.95, while in the northern Barents Sea T_A is about 1 ºC and S_A about 34.85. If the upper layer

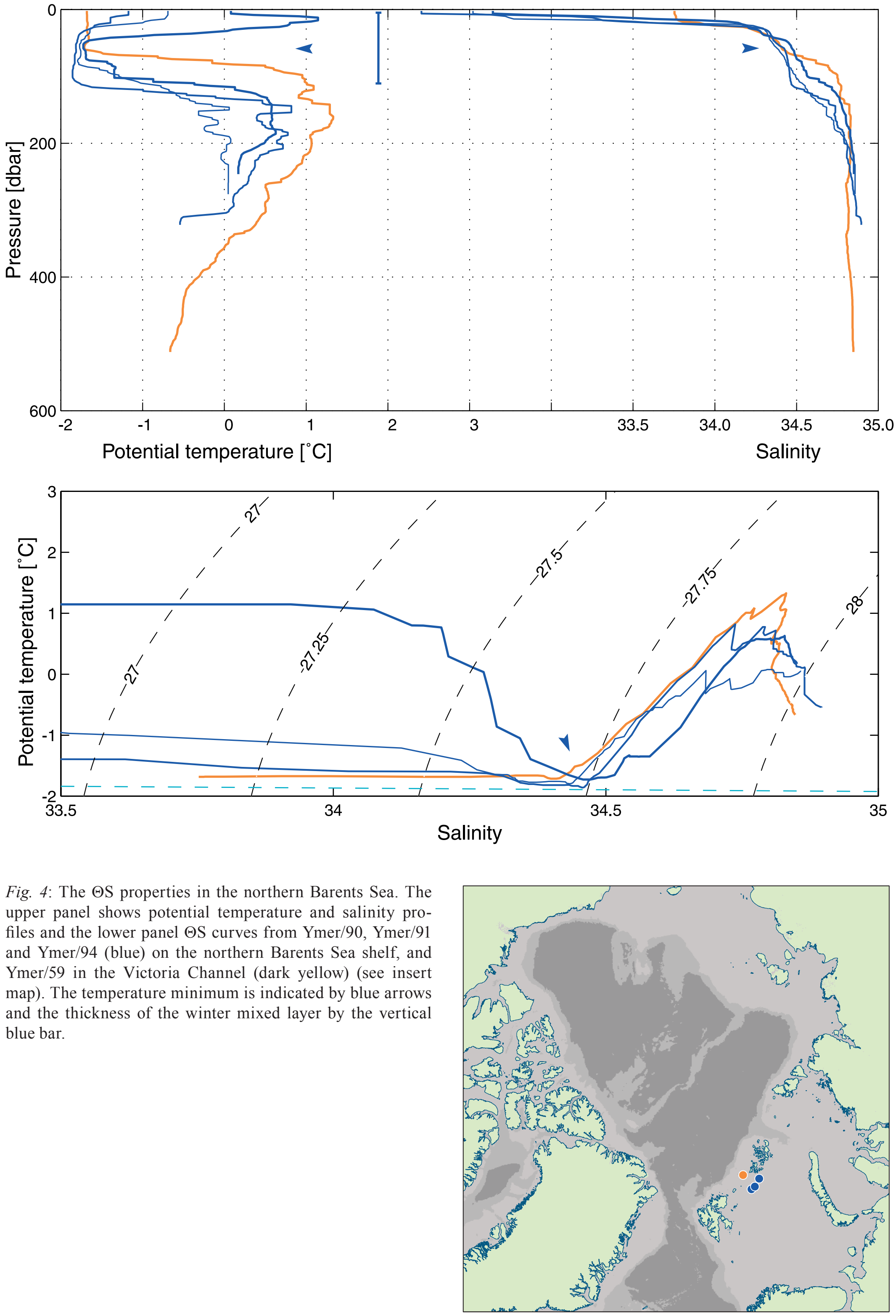


Fig. 4: The ΘS properties in the northern Barents Sea. The upper panel shows potential temperature and salinity profiles and the lower panel ΘS curves from Ymer/90, Ymer/91 and Ymer/94 (blue) on the northern Barents Sea shelf, and Ymer/59 in the Victoria Channel (dark yellow) (see insert map). The temperature minimum is indicated by blue arrows and the thickness of the winter mixed layer by the vertical blue bar.

is assumed to be at freezing temperature, α, taken for the mean temperature of the layers, becomes 0.56×10^{-4} north of Svalbard and 0.46×10^{-4} in the northern Barents Sea (Gill 1982, Table A3.1). This gives $f \approx 0.34$ north of Svalbard and $f \approx 0.28$ in the Barents Sea and leads to a salinity ~34.34 for the winter mixed layer north of Svalbard, and ~34.52 in the northern Barents Sea. This shows that lower temperature rather than higher salinity in the underlying (Atlantic) water leads to higher salinity in the upper layer. The computed values are close to those observed (Fig. 3 & Fig. 4). The salinity of the upper layer will be lower if an additional freshwater source is available, e.g., precipitation. It will be higher if the heat loss from the ocean is not enough to supply the heat demanded by the atmosphere, and new ice begins to form at the same time as existing ice is melting.

The water masses encountered in the eastern part of the St. Anna Trough are similar to those found in the northern and eastern Barents Sea (Fig. 5). The temperature minimum, deriving from winter homogenization of the upper layer, has somewhat higher salinity (34.5 - 34.6) than in the northern Barents Sea, but close to that observed west of Novaya Zemlya (Fig. 5, red station). On the eastern flank of the St. Anna Trough the temperature minimum is located between 50 m and 100 m. A slight temperature maximum is present above the minimum, while the surface water is colder and less saline. This structure could be explained, if the winter mixed layer, after homogenization in the eastern Barents Sea, is advected towards St. Anna Trough. The upper part is warmed by summer heating during the first part of the transit, but when it encounters the Arctic Ocean ice cover, its surface temperature and salinity decrease because of ice melt. We consider this winter mixed layer with salinity around 34.5 as the embryo of the Barents Sea branch halocline water.

Two other cold, but denser, water masses are present in the St. Anna Trough. The bottom layer in the deepest part is comparatively warm (–0.5 ºC) (Fig. 5, dark yellow station) and has characteristics similar to the densest water found in the Victoria Channel (Fig. 4, dark yellow station), where an outflow of dense bottom water from the Barents Sea also takes place (Rudels 1987). The water found in the St. Anna Trough is likely formed by convection to and accumulation of brine enriched water at the bottom in the eastern Barents Sea (Fig. 5, red station) and then advected to the. St. Anna Trough (Schauer, Loeng et al. 2002). At the eastern flank of the trough another, colder and less saline, water mass is present (Fig. 5, blue stations). Since this water mass is not found in the Victoria Channel, it probably derives from the shallow banks in the northern Kara Sea just east of the trough. These denser waters are discussed more fully elsewhere (Schauer, Rudels et al. 2002).

The Eurasian continental slope from St. Anna Trough to the East Siberian Sea

The outflow in the St. Anna Trough is much stronger than that in the Victoria Channel, and the Barents Sea branch waters enter the Nansen Basin as a stratified water column, displacing the Fram Strait branch from the slope (Rudels, Jones et al. 1994; Schauer, Muench et al. 1997). The ΘS characteristics of the Barents Sea branch are initially located within the envelope of the Fram Strait branch ΘS curves, its Atlantic and deeper layers being colder and less saline, while its temperature minimum (Θ ~ –1.8 ºC, S ~ 34.5) and thermocline are more saline than in the Fram Strait branch (Θ ~ –1.8 ºC, S ~ 34.2) (Fig. 6, blue and yellow stations).

Mixing occurs between the two branches and is most strongly seen in the Atlantic layer. The mixing begins already in the St. Anna Trough (Fig. 5). North of Severnaya Zemlya the exchange between the two water columns is at its most intense, and parcels of cold (–1.8 ºC) Barents Sea branch halocline water are mixed into the Fram Strait branch thermocline (e.g., see the blue station in Fig. 3). The presence of water with salinity 34.4 - 34.5 and temperatures close to freezing this far to the east suggests either that the northern Kara Sea also contributes to the Barents Sea branch halocline water and/or that the Barents Sea branch, just as the Fram Strait branch, is ventilated in winter down to the pycnocline as it is advected eastward.

North of the Laptev Sea the salinity of the surface water diminishes in both branches, indicating that shelf water with lower salinity enters the Eurasian Basin at this point, and the winter convection becomes limited to the injected, less saline, surface layer. The ultimate source of the low salinity shelf water is also the inflow to the Barents Sea from the Norwegian Sea. One part,

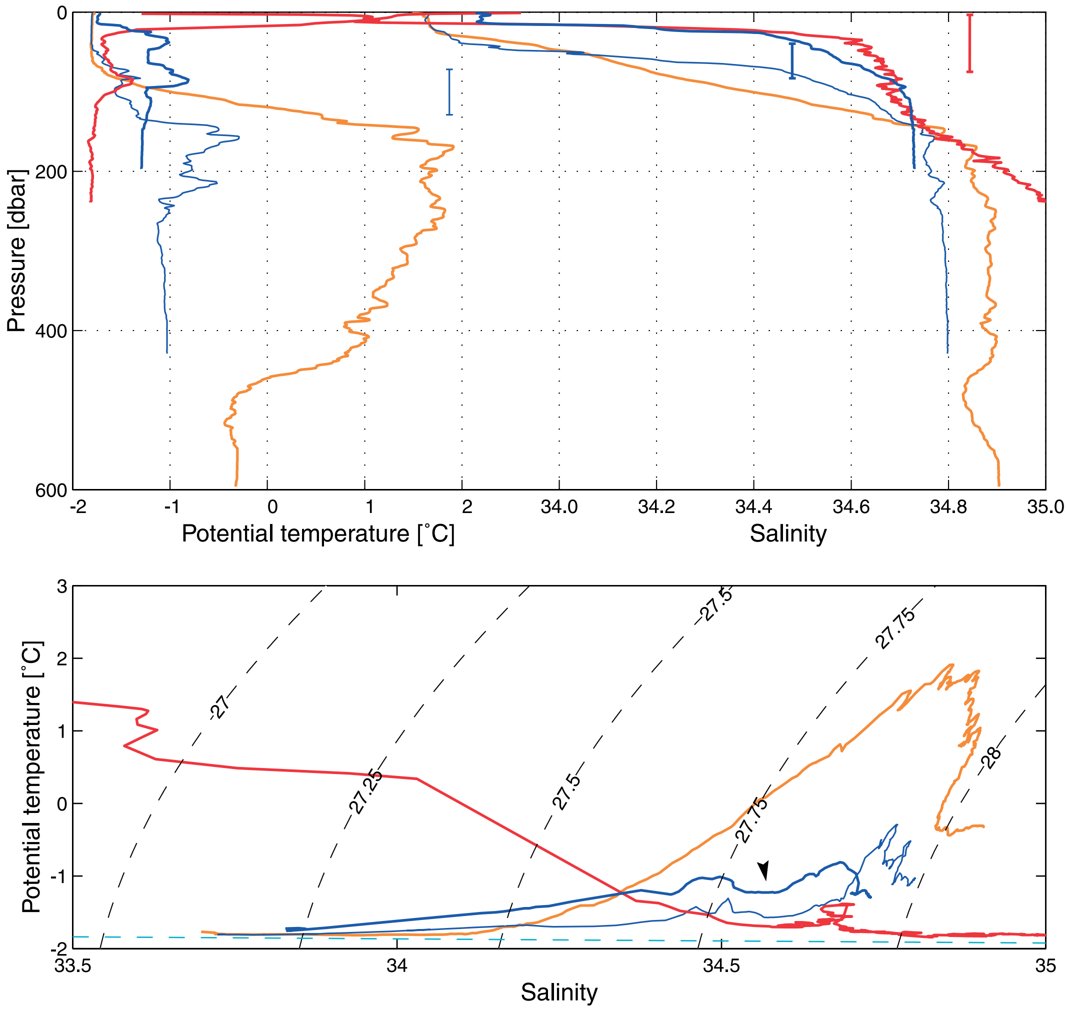
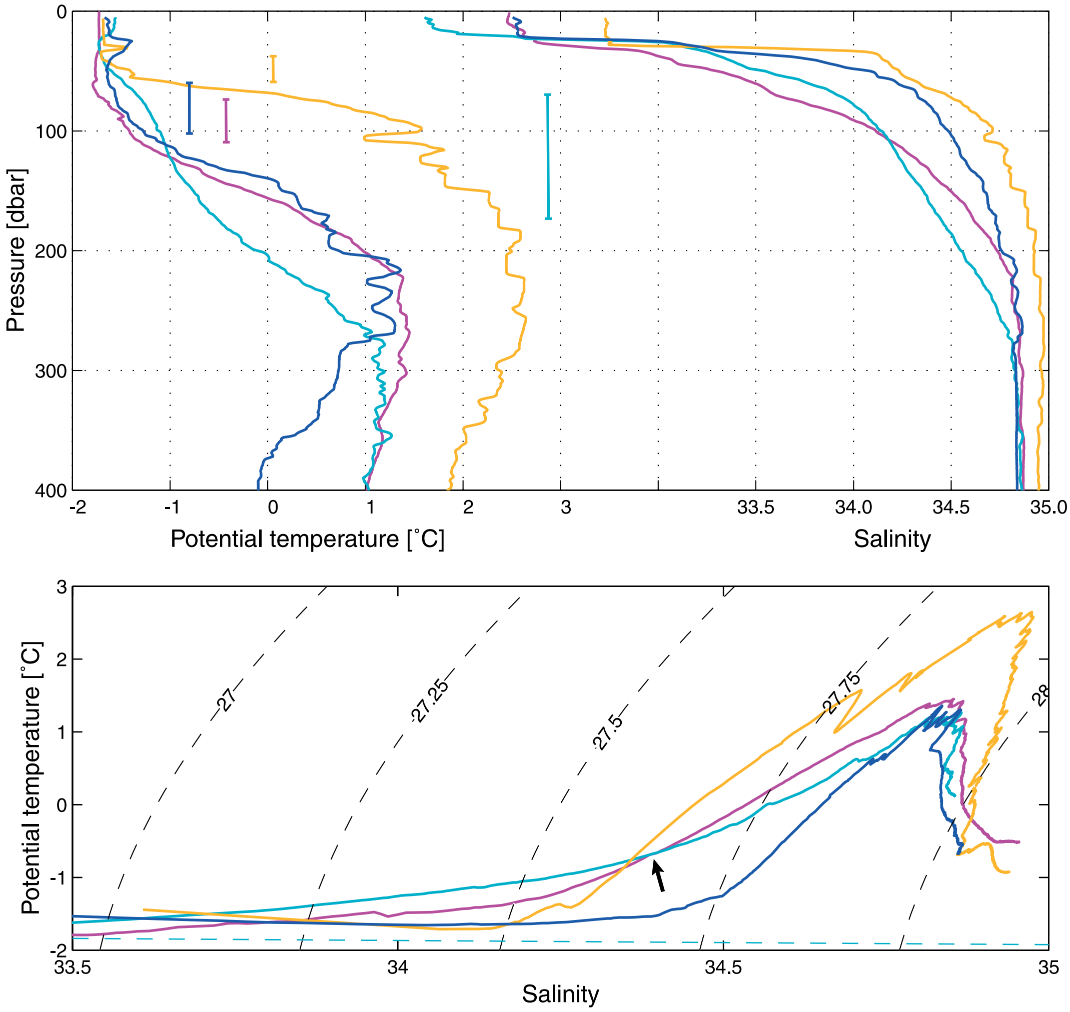


Fig. 5: The ΘS properties in the St. Anna Trough. The upper panel shows potential temperature and salinity profiles and lower panel ΘS curves from ArkXII/9 (dark yellow) and ArkXII/13 and ArkXII/16 (blue) from the St. Anna Trough and DZ/19 (red) from west of Novaya Zemlya (see insert map). The arrow indicates the temperature minimum on the ΘS curves, and the depth of the winter mixed layer west of Novaya Zemlya is shown by a red vertical bar. The blue vertical bars indicate the core of the Barents Sea branch at the eastern flank of the St. Anna Trough. A dense, cold outflow mode is seen at the bottom on the eastern flank (blue stations), different from the dense Barents Sea outflow found at the bottom in the deeper part of the trough (dark yellow station).



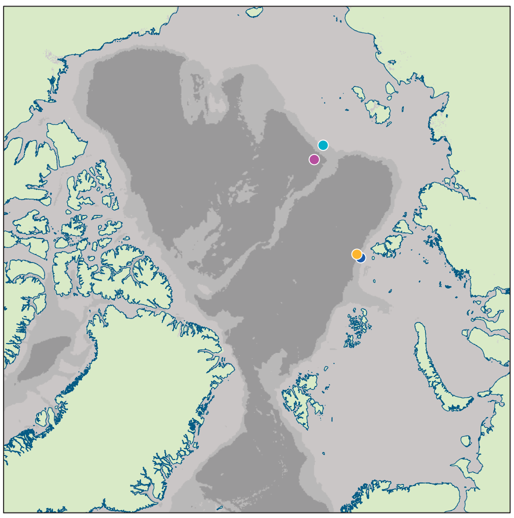


Fig. 6: The evolution of the ΘS properties in the boundary current from the Kara sea slope to the East Siberian Sea slope. The upper panel shows potential temperature profiles and salinity profiles and the lower panel ΘS curves from the Barents Sea branch (ArkXI/91, blue) and the Fram Strait branch (ArkXI/89) north of the eastern Kara Sea, and from Barents Sea branch (ArkXI/63, cyan), and the Fram strait branch (ArkXI/57, magenta) north of the western East Siberian Sea (see insert map). The vertical bars indicate the depth range of the halocline, defined as extending from the temperature minimum to the level of the maximum rate of temperature increase with depth. The Fram Strait branch halocline in Nansen Basin (yellow) is likely just seasonal. The arrow indicates the crossing of the two ΘS curves from the two branches north of the East Siberian Sea.

mainly comprising waters from the Norwegian Coastal Current, flows along the inner shelf route south of Novaya Zemlya and Severnaya Zemlya into the Kara, Laptev, and East Siberian seas. It mixes with the river runoff from Ob, Yenisey and Lena and becomes transformed into low salinity shelf water. The bulk of this Atlantic derived shelf water reaches as far east as the Laptev Sea before a substantial flow across the shelf break occurs (Rudels, Friedrich & Quadfasel 1999).

As the winter mixed layers of the two branches become isolated from the sea surface, the temperatures of the initial temperature minima increase by mixing with the underlying Atlantic water (Fig. 6). The temperature increase is greater in the Barents Sea branch than in the Fram Strait branch. As the boundary current crosses the Lomonosov Ridge, the temperature minima are no longer present but have evolved into mere breaks in the slope of the ΘS curves that mark the lower boundary of the halocline. The break in the Fram Strait curves is more distinct, occurs at lower salinities ($S \sim 34.2$) and is located close to the freezing line (Fig. 6, magenta station). The break in the ΘS curves of the inner, Barents Sea branch is more saline ($S \sim 34.5$) and warmer ($\Theta \sim -1.0\,^{\circ}C$) and the curves cross the straight thermocline of the Fram Strait branch ΘS curves (Fig. 6, cyan station).

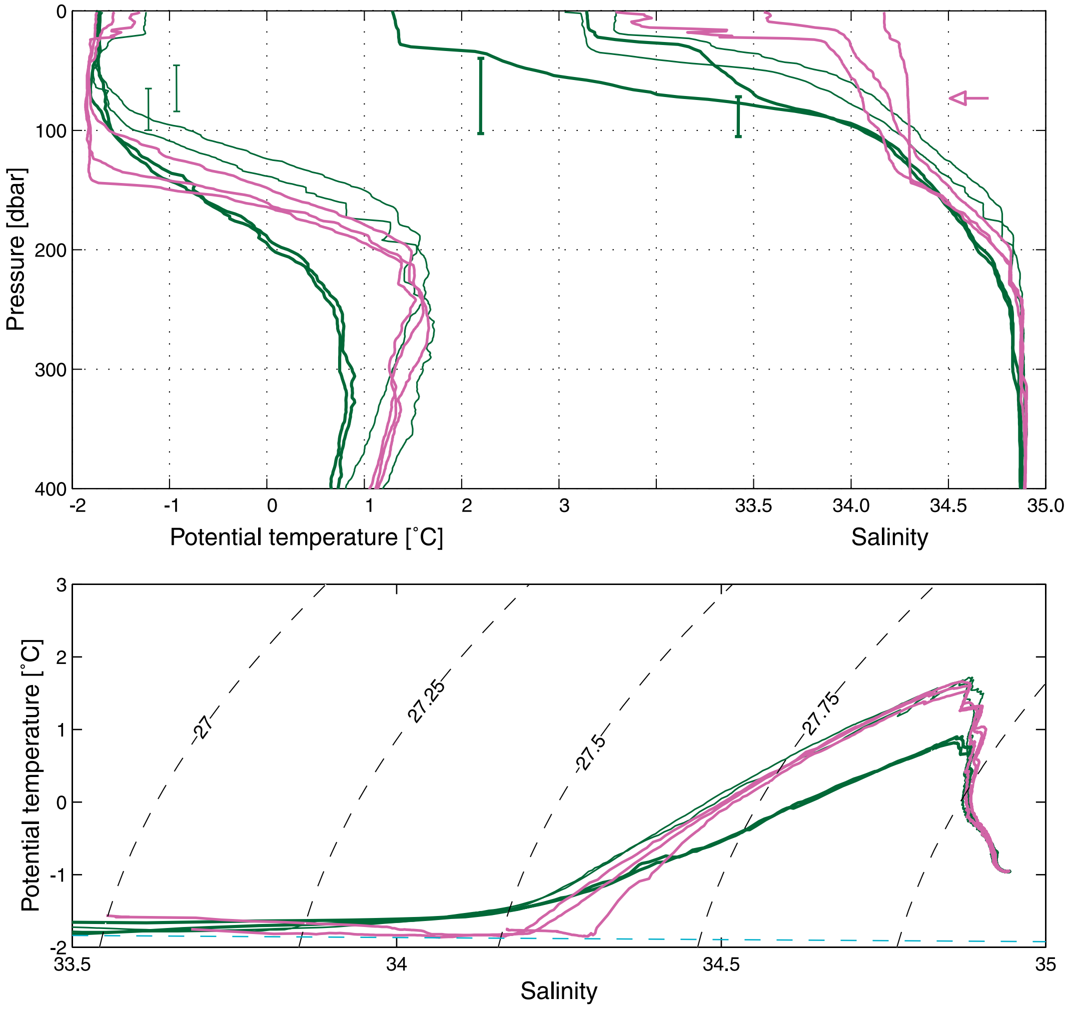
The potential temperature and salinity profiles show that these parts of the ΘS curves are distinct layers, and the breaks are at the level where the rate of temperature increase with depth is the greatest. To be specific, we identify the halocline part of the pycnocline, as distinct from the thermocline below, as the layer between the temperature minimum, marking the limit of the winter convection, and the level of most rapid change of temperature increase with depth. The halocline, so defined, is indicated on all profiles in figure 6. North of the East Siberian Sea the Barents Sea branch halocline (Fig. 6, cyan station) is much thicker than the Fram Strait halocline (Fig. 6, magenta station), extending from 50 m to close to 200 m, as compared to 70 m to 120 m. The Barents Sea branch may have developed a halocline already in the Nansen Basin (Fig. 6, blue station), while in the Fram Strait branch the halocline in the Nansen Basin is most likely only seasonal and will be removed by convection the following winter (Fig. 6, yellow station). The PML, evolving from low salinity shelf water, also pushes the halocline in both branches towards greater depth (Fig. 6).

The larger temperature increase in the Barents Sea branch halocline water implies a stronger turbulent vertical mixing at the continental slope (Woodgate et al. 2001), which not only entrains warmer Atlantic water into the halocline but also mixes colder, less saline halocline water downward into the Atlantic layer. Cooling and freshening of the upper part of the Atlantic layer transform it into halocline water, and the incorporation of Atlantic water from below is thus a possible cause for the increase in thickness of the Barents Sea branch halocline.

The Amundsen, Makarov and Canada basins

The halocline water in the interior Amundsen Basin is cold ($\Theta \sim -1.5\,^{\circ}C$, $S \sim 34.2$), indicating that it derives from the Fram Strait branch (Fig. 7, green stations). The temperature minima observed in Nansen Basin are not seen on these stations, showing that heat has been added from below, and the almost constant salinity in the Nansen Basin winter mixed layer (Fig. 3) has evolved into a salinity gradient. This is primarily caused by mixing with the overlying, less saline PML, since mixing with the Atlantic water below would have increased the temperature, as well as the salinity, in the lower part of the halocline more strongly than observed.

Both inflow branches cross the Lomonosov Ridge and contribute to the lower Atlantic derived halocline in the Canadian Basin. Observations in 1994 showed that the intermediate depth waters in the interior Makarov Basin were supplied by a splitting of the boundary current at the Mendeleyev Ridge, and lower salinities suggested that a substantial part of this input involved waters of the Barents Sea branch (Carmack, Aagaard et al. 1997; Swift et al. 1997). The Makarov Basin halocline water, however, is cold, and the salinity at the bend of the ΘS curves is ~ 34.2 (Fig. 8, violet stations). This is similar to what is observed in the Amundsen Basin (Fig. 7, green stations) and implies that the halocline waters in the interior Makarov Basin mainly derive from the Fram Strait branch. The more saline, and now warmer, Barents Sea branch halocline remains at the slope. The Barents Sea branch halocline becomes more prominent between the Mendeleyev Ridge and the Chukchi Cap (Fig. 8, cyan stations). The temperature maximum at 75 m indicates the presence



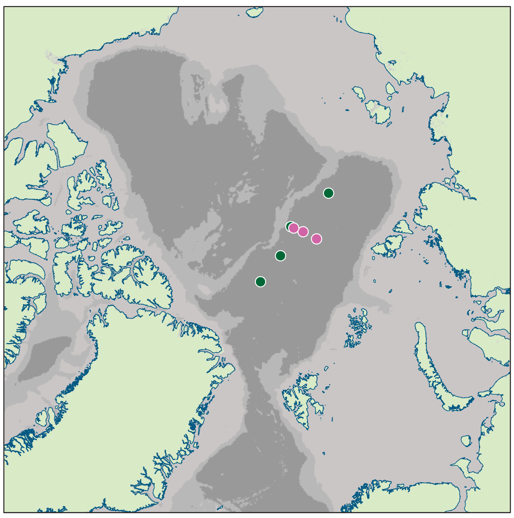


Fig. 7: The properties of the upper layers in the Amundsen Basin. The upper panel shows potential temperature and salinity profiles and the lower panel ΘS curves from Oden/18 and Oden/31 (thick green) in the central and western part and ArkXII/58 and ArkXII/82 (thin green) in the eastern part of the basin (see insert map). The temperature in the Atlantic layer is different in the eastern and western Amundsen Basin because the 1991 (Oden) stations were taken before the anomalously warm Atlantic water, first reported by Quadfasel et al. (1991), had reached this position. The vertical bars indicate the depth range of the halocline. ArkXII/50, ArkXII/55 and ArkXII/57 (magenta) from the central Amundsen Basin and the Nansen-Gakkel Ridge show a homogeneous, saline winter mixed layer of varying depth (indicated by arrow) present in 1996.

of the Pacific inflow and the Bering Sea summer water, which depresses the Barents Sea branch "lower" halocline, reducing its thickness compared to north of the East Siberian Sea.

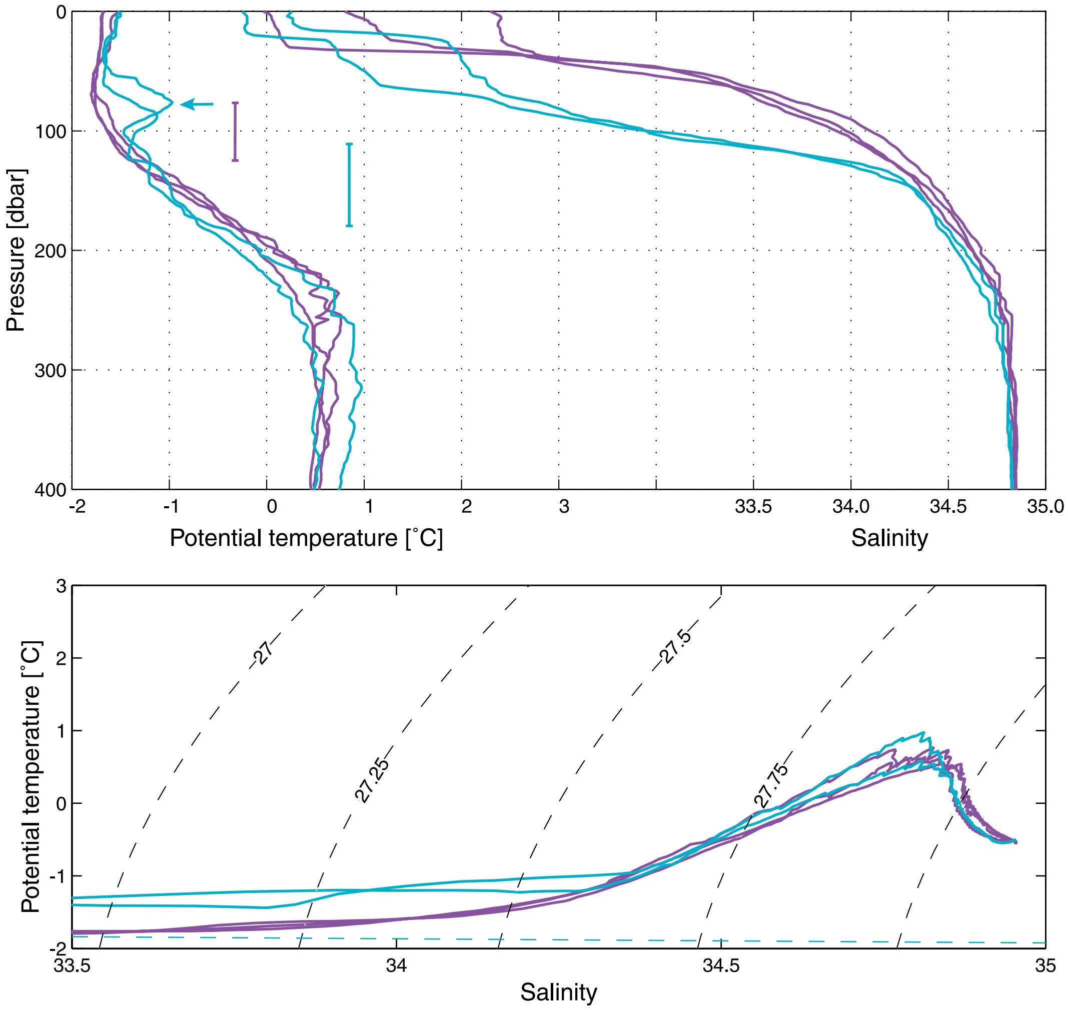
Beyond the Chukchi Cap no Fram Strait contributions can be clearly identified either at the slope or in the interior of the Canada Basin (Fig. 9). A splitting of the boundary current at the Chukchi Cap has been suggested based on the characteristics of the Atlantic and intermediate layers (Rudels, Anderson et al. 1996; Smith et al. 1999; Smethie et al. 2000) as well as by the flow of near-surface Pacific waters (Shimada et al. 2001; Jones, Anderson et al. 1998). The flow around the Chukchi Cap has also been discussed in detail by McLaughlin, Carmack, MacDonald, Melling et al. (2004). A disappearance of the Fram Strait branch halocline characteristics at the slope is consistent with a loss of the outer part of the boundary current to the interior basin. The Fram Strait branch would then mainly be confined to the part of the Canada Basin closest to the Mendeleyev and Alpha ridges, while the Barents Sea branch continues along the slope and supplies the southern Canada Basin beyond the Chukchi Cap. The ΘS curves in the Canada Basin are smooth and the bend between halocline and thermocline almost impossible to distinguish. The Barents Sea branch halocline will here be identified as lying between the Pacific derived upper water and the Atlantic layer, having the potential temperature and salinity ranges $\Theta < 0\,°C$ and $33.6 < S < 34.6$.

The Pacific water is initially less dense than the Atlantic derived lower halocline, and in the Arctic Ocean the Pacific source water is distinguished from water of Atlantic origin by nutrient concentrations. Following earlier work (Jones, Anderson et al. 1998; Jones, Swift et al. 2003), we note that north of Alaska, nitrate–phosphate relationships show Pacific source water in the Canada Basin penetrating to depths of 250 m, where salinities exceed 34.3, well into the Atlantic derived Barents Sea branch halocline (not shown). Salmon & McRoy (1994) observed an increase in silicate and temperature in the salinity range of the lower halocline in the Canada Basin as compared to the Amundsen and Makarov basins and argued that this is caused by the presence of Pacific water. The higher silicate concentrations are consistent with an input of Pacific water into the lower halocline, but the higher temperatures indicate rather the presence of the Barents Sea branch halocline water than shelf derived Pacific water, which would be colder because of freezing and brine rejection. However, the temperatures in the 34 - 34.5 salinity range are lower at some stations, which again substantiates an input of Pacific water made more saline by brine rejection, in agreement with Salmon & McRoy (1994). Large vertical motions of the isopycnals have been observed on the continental slope in Beaufort Sea (Kulikov et al. 1998) and the possibility that Pacific water is mechanically mixed into the lower halocline by increased turbulence at the slope also exists.

The predominance of the Barents Sea branch halocline in the Canada Basin is consistent with the suggestion by Melling (1998), based on the time variability of the temperature of the lower halocline, defined as the 34.5 isohaline, in the Canada Basin, that the lower halocline is responding to changes in the inflow of the Barents Sea branch. The reported variability is rather small, even in the Atlantic layer—at least it appears so to workers who are used to changes of 1 ºC between successive CTD casts at the same station, which occur in the frontal zone between the two branches (see e.g. Rudels, Muench et al. 2000). A time variation of the relative contributions from the branches could have a similar effect.

Melling sees a Barents Sea origin of the lower halocline in the Canada Basin as an argument in favour of the source and the formation of the lower halocline proposed by Steele et al. (1995) as compared to that of Rudels, Anderson et al. (1996). As to the sources, the present work shows that the halocline waters from the different sources dominate in the different basins, and the presence of Barents Sea branch halocline in the Canada Basin is in agreement with Melling's conclusions.

The difference in the proposed formation processes is mainly that Steele et al. (1995) emphasize the melting of sea ice in the marginal ice zone, in the northern Barents Sea but also north of Svalbard, while Rudels, Larsson et al. (1991) and Rudels, Andersson et al. (1996) consider the melting of sea ice merely as a preconditioning of the upper part of the water column for ice formation in winter and haline convection down to the thermocline. The winter homogenization leads to the formation of mode waters (McCartney 1977) that eventually evolve into the halocline, and the mode waters give the halocline its distinct, initial characteristics.



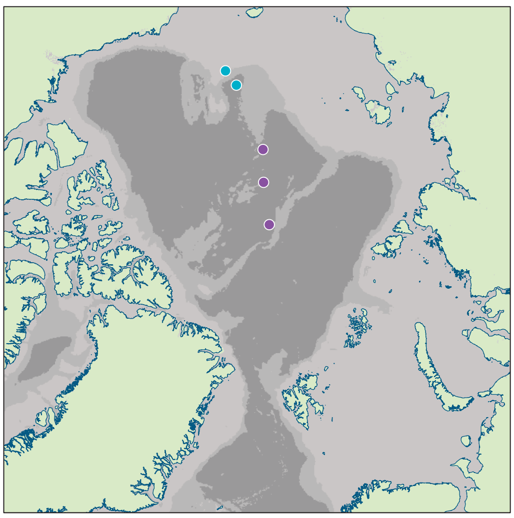
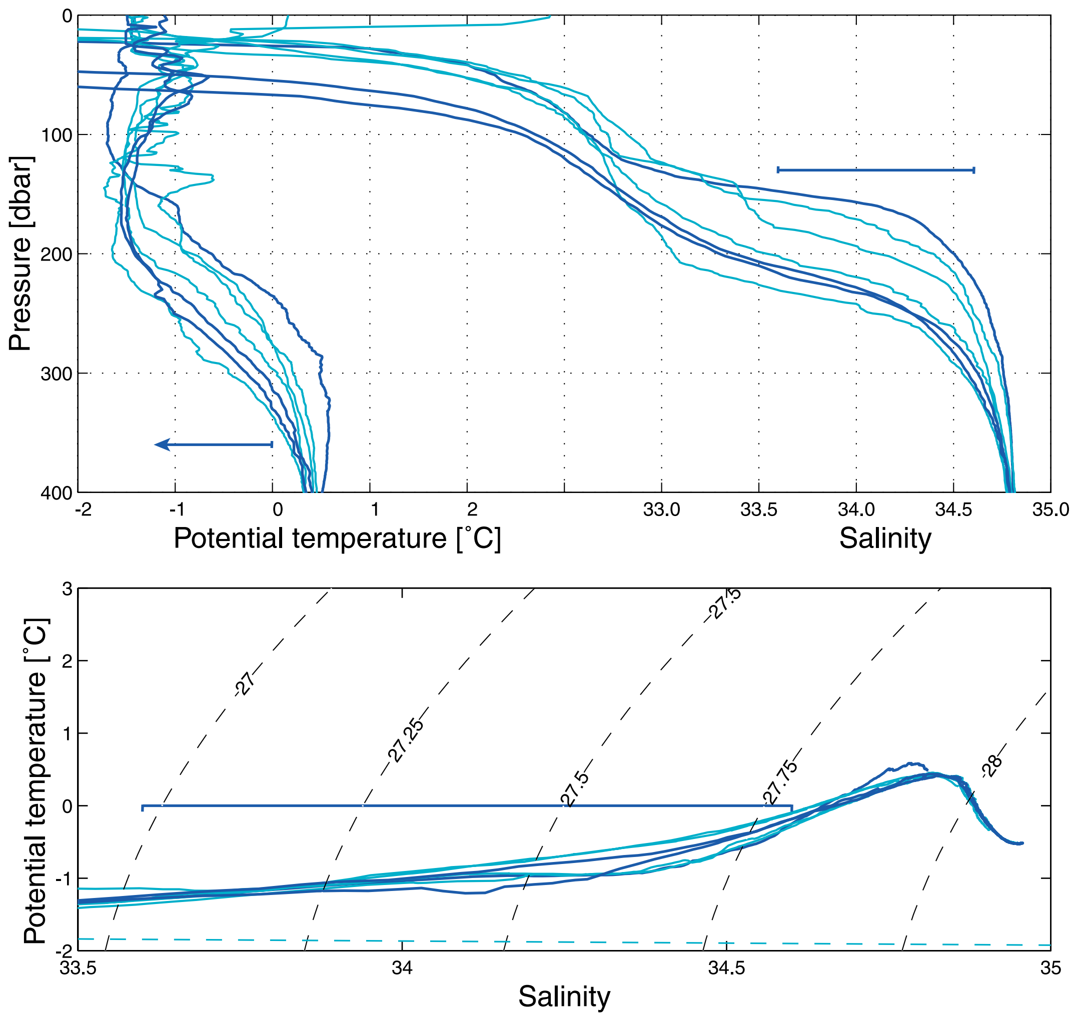


Fig. 8: The properties of the upper layers of the Makarov Basin. The upper panel shows potential temperature and salinity profiles and the lower panel ΘS curves for AOS/23, AOS/26 and AOS/29 (violet) and AOS/8 and AOS/11 (cyan) from the Makarov Basin and from the Canada Basin west of the Chukchi Cap (see insert map). The violet stations from the interior indicate Fram Strait branch halocline water, while the cyan stations from the slope show Barents Sea branch halocline water. The upper temperature maximum is due to the presence of Bering Strait summer water (arrow) and the vertical bars show the depth range of the halocline.



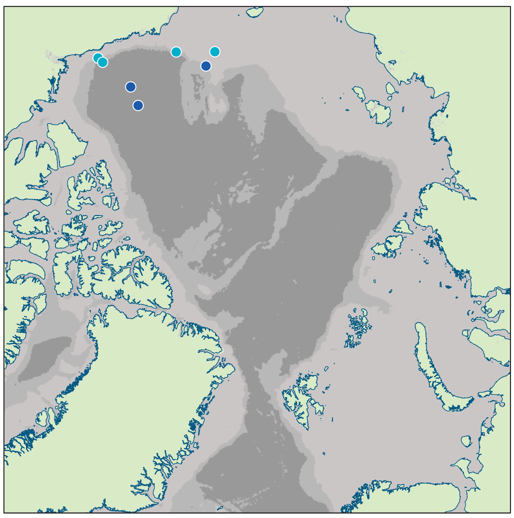


Fig. 9: The properties of the upper waters in the Canada Basin east of the Chukchi Cap (the Beaufort Sea). The upper panel shows potential temperature and salinity profiles and the lower panel ΘS curves from ArcRad/C01, ArcRad/C06, Jois/4 and Jois/7 (cyan) at the slope and ArcRad/C05, Jois/12 and Jois/14 (blue) in the deep Canada Basin (see insert map). All stations show Barents Sea branch halocline water. Because of the varying depth of the lower halocline, it is indicated not by vertical but by horizontal bars extending over the temperature and salinity ranges of the Barents Sea branch halocline.

Time variability

Observations of large changes in the Arctic Ocean water column have been reported during the last 15 years. The temperature of the Atlantic layer has increased (Quadfasel et al. 1991; Carmack, MacDonald et al. 1995) and its upper boundary has become shallower, at least in the Amundsen and Makarov Basins (Morison et al. 1998). These changes have been related to a shift in the distribution of the Pacific water in the Canadian Basin (Carmack, MacDonald et al. 1995; McLaughlin, Carmack, MacDonald & Bishop 1996). Observations from the early ice stations in the 1960s indicated that nutrient rich Pacific water was present in the Makarov Basin up to the Lomonosov Ridge (e.g. Kinney et al. 1970). By the mid 1990s Pacific water was not found on the Siberian side of the Makarov Basin (McLaughlin, Carmack, MacDonald & Bishop 1996) and a shoaling of the boundary between the Pacific and Atlantic derived haloclines has been reported also from the western part of the Canada Basin (McLaughlin, Carmack, MacDonald, Weaver et al. 2002).

These changes are likely connected with the weakening, and occasionally even the disappearance, of the PML in these basins (Steele & Boyd 1998). Close to the North Pole, the low salinity PML (or shelf water) present in 1991 was reported absent in 1995 (Steele & Boyd 1998). In 1996 the two-layer PML–halocline structure was absent on stations over the Nansen-Gakkel Ridge and in the central Amundsen Basin, and a winter mixed layer reaching down to the pycnocline was observed (Fig. 7, magenta stations). Steele & Boyd (1998) attribute these changes to the diversion of the low salinity Laptev Sea shelf water from the Amundsen Basin into the East Siberian Sea. Such change in the shelf water distribution could be the result of the strong positive AO (Arctic Oscillation) and NAO (North Atlantic Oscillation) states prevailing in the 1990s (Proshutinsky & Johnson 1997; Polyakov & Johnson 2000).

The depth of the pycnocline varied from more than 150 m over the Nansen-Gakkel Ridge to less than 60 m in the Amundsen Basin, indicating that, besides convection, internal wave and eddy motions also influence the observed thickness of the winter mixed layer. In 1995 - 1996 the region with a winter homogenized upper layer thus extended from the Nansen Basin into the Amundsen Basin and up to the Lomonosov Ridge (Schauer, Rudels et al. 2002). It did not, however, reach the eastern parts of the Eurasian Basin and the Eurasian continental slope, where a distinct PML–halocline structure was observed in 1993, 1995 and 1996. Recent observations have shown that low salinity shelf water again is present in the Amundsen Basin and at the Lomonosov Ridge (Björk et al. 2002; Boyd et al. 2002; Anderson et al. 2004).

Recently McLaughlin, Carmack, MacDonald, Melling et al. (2004) have reported observations of cold (close to freezing), saline (34.2) and oxygenated water north of the Chukchi Cap: their Type I profiles. These characteristics are close to those of the Fram Strait branch halocline (see, e.g., Fig. 6, Fig. 7 & Fig. 8), and the location suggests a spreading of the Fram Strait halocline into the Canada Basin west and north of the Chukchi Cap. By contrast the Type II profiles (McLaughlin, Carmack, MacDonald, Melling et al. 2004) found east of the Chukchi Cap had a warmer and more saline break in the ΘS curves, which corresponds to the characteristics of the Barents Sea branch halocline moving eastward south of the Chukchi Cap (see, e.g., Fig. 6, Fig. 8 & Fig. 9).

McLaughlin, Carmack, MacDonald, Melling et al. (2004) suggest the East Siberian Sea as the source for the cold Type I halocline water, which to us appears rather unlikely. The salinity of the waters on the East Siberian Sea shelf is low and more than four metres of ice must be formed before salinities above 33.5 are reached (Aagaard, Coachman et al. 1981). The recent redistribution of the low salinity shelf water from the Laptev Sea into the East Siberian Sea instead of the Amundsen Basin should lead to even lower salinities. The accumulation of brine-enriched water at the bottom is also likely to remineralize organic material, leading to high nutrient and lower oxygen concentrations.

Our interpretation of the cold and saline halocline found in the Type I profiles thus differs from that given by McLaughlin, Carmack, MacDonald, Melling et al. (2004). The cold, oxygenized halocline suggests a recently ventilated Fram Strait branch winter mixed layer. This could have been advected from the Amundsen Basin, where such layers have been observed (Fig. 7, magenta stations) or the low salinity surface water could have absent also in parts of the Makarov Basin, allowing for convection down to the thermocline at least in the southwestern Makarov Basin.

The absence of low salinity surface water implies the absence of a cold halocline, and vertical mixing and entrainment can add both salt and heat to the mixed layer. Such heat transport could reduce the ice formation in winter, since part of the heat given up to the atmosphere would be sensible heat, not just the latent heat released by freezing. However, the magnitude of the heat flux into the mixed layer depends upon the mixing mechanism. If the entrainment is driven by mechanical stirring, the heat flux might be considerable, while stirring by convecting plumes and thermals is less efficient (Solomon 1973). The plumes may overshoot into the thermocline and become heated there, causing only little heat transfer into the mixed layer. This would create a curved shape of the thermocline in the ΘS curves, as is observed at the stations in the interior Nansen and Amundsen Basins, with a deep, saline mixed layer, in contrast to the straight thermocline found in the areas of more intense mechanical mixing at the continental slope (e.g. Fig. 3 & Fig. 7; see also Figure 10 in Rudels & Friedrich 2000). Plumes require freezing and brine rejection in order to develop, and if the upper layer is stirred by convecting plumes, a stronger entrainment of warm water would reduce the ice formation and weaken the convection, which in turn would reduce the entrainment (Martinson 1990; Martinson & Steele 2001). Furthermore, a deep winter mixed layer with no underlying halocline is the norm in the Nansen Basin. Its extension into the Amundsen Basin and to the Lomonosov Ridge only makes a small contribution to the vertical heat flux normally present in the Arctic Ocean.

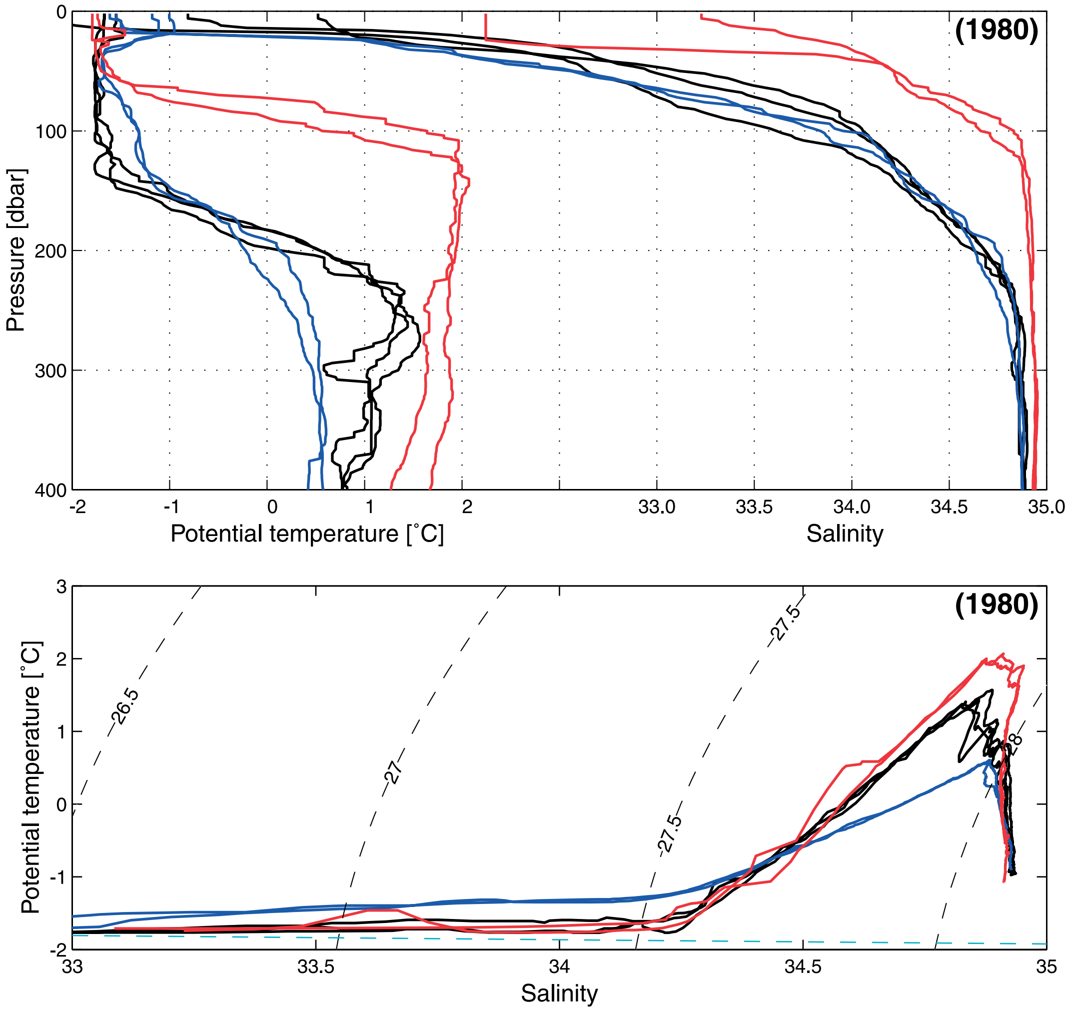
The area north of Greenland

The boundary current returns toward Fram Strait along the North American continental slope, crossing the Alpha and the Lomonosov ridges. The waters from the different loops of the different basins eventually come together and join the boundary current (Rudels, Friedrich & Quadfasel 1999). 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. The contributions from the other basins, upper waters as well as the Atlantic and intermediate layers, line up parallel to the slope from the rim into the basin. The less dense Pacific waters are mainly found close to the continent, inshore of and above the Atlantic derived waters.

Observations from 1980 and 1997 extending from the outflow region near Greenland to the inflow region north of Svalbard are shown in Fig. 10. The surface layer in the inflow area in the eastern Fram Strait had salinities similar to the Atlantic derived halocline waters, showing the first developing stage of the winter mixed layer in the Nansen Basin (Fig. 10a & 10b, red stations). The depth of the temperature maximum in the Atlantic layer was considerably less in the inflowing than in the outflowing water columns, consistent with the existence of sources other than the melting of sea ice north of Svalbard contributing to the polar surface water in the Arctic Ocean. The Barents Sea branch halocline water was more clearly distinguished in 1980 (Fig. 10a) than in 1997, when it was interacting with the colder outflow from the Nansen and Amundsen Basins (Fig. 10b, blue stations). In 1980 the Fram Strait branch formed a proper halocline between the thermocline and the PML. In 1997 the PML was absent at some stations and a deep winter mixed layer with temperatures close to freezing was observed (Fig. 10b, black stations). This suggests that part of the deep winter mixed layer encountered in the Eurasian Basin in 1995 and 1996, and discussed in the time variability section above, could be exiting the Arctic Ocean as early as 1997 (Rudels, Meyer et al. 2000). At one station the isothermal layer reached as deep as 240 m (Fig. 10b, black station). However, the salinity profile showed a three-layer structure with a low-salinity surface layer. This implies that low salinity water, perhaps including Pacific water that is known to partly exit through Fram Strait (Jones, Swift et al. 2003), has been deflected from its inshore path and moved on top of a stream of Fram Strait halocline water, forcing the denser Atlantic halocline and PML waters downward.

The outflow through the Canadian Arctic Archipelago

As the Barents Sea branch halocline water moves along the continental slope from the Canada Basin toward northern Greenland, it ascends in the water column from 250 - 300 m north of Alaska to about 150 - 200 m north of Greenland (Fig. 11). This rise is due to the gradual thinning of the, mostly Pacific, layers above. The thinning could be caused by the Pacific water either being confined to the Beaufort Sea gyre or drain-



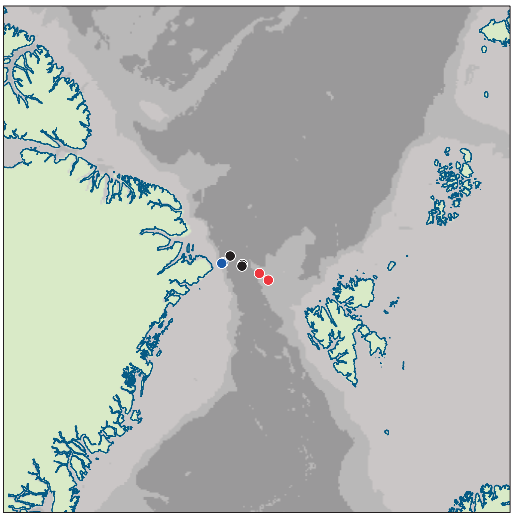


Fig. 10a: Properties of the upper layers north of Fram Strait in 1980. The upper panel shows potential temperature and salinity profiles and the lower panel shows ΘS curves from Ymer/171 and Ymer/172 (red), Ymer/166, Ymer/168 and Ymer/170 (black) and Ymer/162 and Ymer/164 (blue) (see insert map). Black stations show returning Fram Strait branch halocline water; blue station returning Barents Sea branch halocline water. The red stations are from the inflow area at the Yermak Plateau.

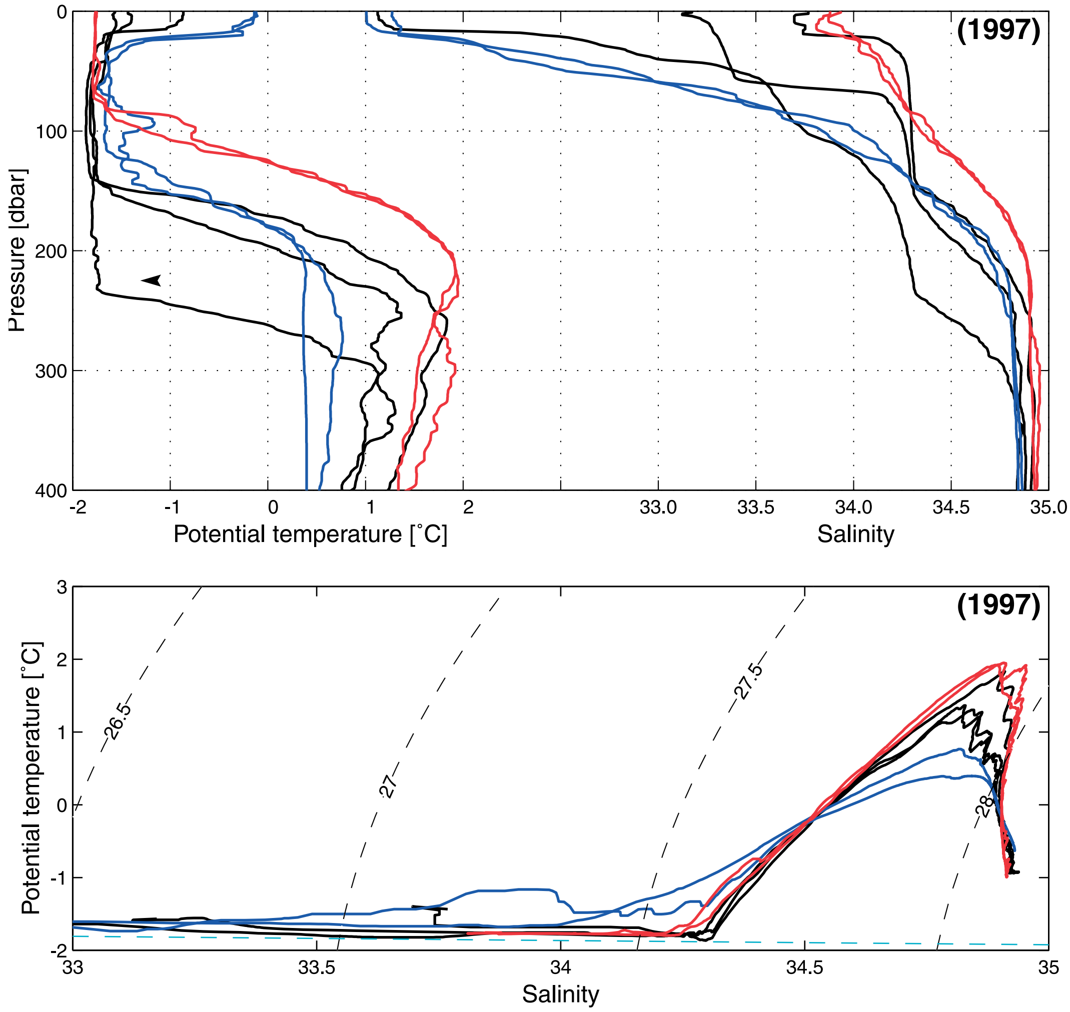
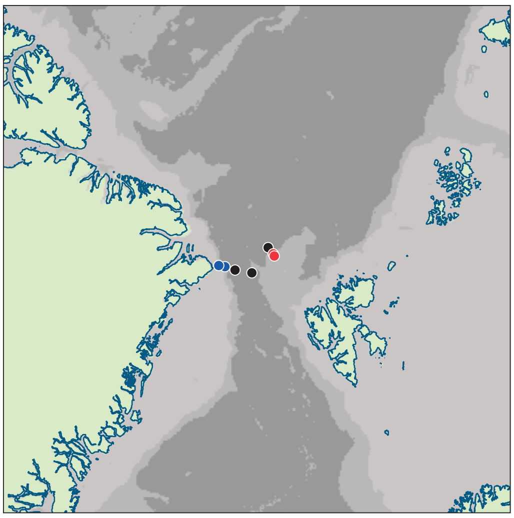


Fig. 10b: Properties of the upper layers north of Fram Strait in 1997. The upper panel shows potential temperature and salinity profiles and the lower panel ΘS curves from ArkXIII/86 and ArkXIII/87 (red), ArkXIII/68, ArkXIII/69 and ArkXIII/82 (black) and ArkXIII/72 and ArkXIII/73 (blue) (see insert map). Black stations show returning Fram Strait branch halocline water; blue stations returning Barents Sea branch halocline water. The red stations are from the inflow area at the Yermak Plateau. The almost constant salinity below the seasonal melt layer and the curved thermocline in the black salinity profiles both suggest that part of the winter mixed layer observed in the Eurasian Basin in 1995 and 1996 (Steele & Boyd 1998) now is returning to Fram Strait. The deep, cold isothermal layer on one station (indicated by arrow) in reality comprises the winter mixed layer of the Eurasian Basin and PML deriving both from Eurasian and the Canadian basins.



ing through the Canadian Arctic Archipelago and ultimately into Baffin Bay. We favour the second interpretation. Jones, Swift et al. (2003) identified Arctic water of Atlantic origin flowing through Nares Strait, and suggest that the shallowing of the Barents Sea branch halocline water allows it to pass over the sill in Nares Strait (230 m), and enter Baffin Bay. The other channels in the Canadian Arctic Archipelago are shallower and permit only the passage of Pacific water and other waters in the PML, e.g., river runoff and sea ice meltwater (Jones, Swift et al. 2003).

Baffin Bay bottom water

Bailey (1956) noted that the characteristics ($\Theta = -0.45\,^{\circ}C$, $S = 34.45$) of Baffin Bay bottom water were similar to those found at 250 m in the Beaufort Sea and suggested the Arctic Ocean as the source of the Baffin Bay deep and bottom waters. These are also the characteristics of the Barents Sea branch halocline water, and the fact that these water mass characteristics are found at comparatively shallow depths north of Greenland and close to Nares Strait supports Bailey's suggestion. However, Bourke et al. (1989) and Bourke & Paquette (1991) did not observe any water with salinity and temperature matching those of the Baffin Bay deep and bottom waters flowing through Nares Strait into Baffin Bay in 1986. Instead they proposed that dense water formation and haline convection, mainly in the North Water polynya in the southern part of Nares Strait, contribute to the Baffin Bay deep and bottom water. Such convection has, so far, not been observed.

The bottom water in Baffin Bay is old. Its oxygen saturation is just over half that of surface waters, and the CFC concentrations are near detection limit (Wallace 1985). Wallace suggested that the ventilation time could be as long as 1000 years, and thus any renewal, by inflow or by convection, would be slow and intermittent. What may be concluded from the ΘS characteristics, however, is that no water denser than the Barents Sea branch halocline enters Baffin Bay from the Arctic Ocean, and if the Barents Sea branch is an important source for the deep water in Baffin Bay, the entrainment of ambient water into the inflowing and descending plumes must be small. In Baffin Bay the intermediate waters are warmer and more saline than the bottom water. A contribution from the Arctic Ocean dense enough to allow for a stronger entrainment of intermediate water and still reach the bottom would involve water with higher temperatures and salinities. Both stronger entrainment and a denser inflow would together lead to higher temperatures and higher salinities in the Baffin Bay deep and bottom water than those observed.

Estimate of flow through the Canadian Arctic Archipelago

That the Barents Sea branch, together with the Pacific inflow through Bering Strait, supplies the outflow through the Canadian Arctic Archipelago also makes it possible to crudely assess the volume and freshwater transports through the archipelago. The existing estimates of the outflow range from less than 1 Sv to above 2 Sv, with the higher end estimates commonly chosen (e.g., Aagaard & Carmack 1989).

Some Pacific water exits through Fram Strait (e.g., Jones, Swift et al. 2003) and perhaps 0.6 Sv, 75% of the estimated 0.8 Sv (Roach et al. 1995) entering through the Bering Strait, passes through the Canadian Arctic Archipelago. We assume that the salinity of the Pacific derived water is 32.5, the same as in Bering Strait. This is likely an overestimate because runoff is added from the Mackenzie and Siberian rivers as well as other freshwater sources. Following our definition of the Barents Sea branch halocline water in the Canada Basin we take the upper salinity limit of the Pacific water to be 33.6. In the Canadian Arctic Archipelago, only Nares Strait, with a sill depth of about 230 m, is deep enough to allow the passage of waters from the Barents Sea branch halocline. The salinity cannot be higher than ~34.45 of the Baffin Bay bottom water, but it could be significantly lower, since the Atlantic inflow to the Barents Sea, together with the runoff, supplies the water on the Siberian shelves, which enters the deep Arctic Ocean above the Barents Sea branch halocline. We assume a salinity of 34.0 (mean of 33.6 and 34.45) for the Atlantic water passing through the Canadian Arctic Archipelago. This is probably high since less dense water more easily passes through the shallow straits.

We make tentative mass and freshwater budgets for outflows of 2, 1.5 and 1 Sv (Table 3). The inflow from the Norwegian Sea to the Barents Sea is around 3 Sv, of which 0.7 Sv is Norwegian Coastal Current water with salinity ~34.4.

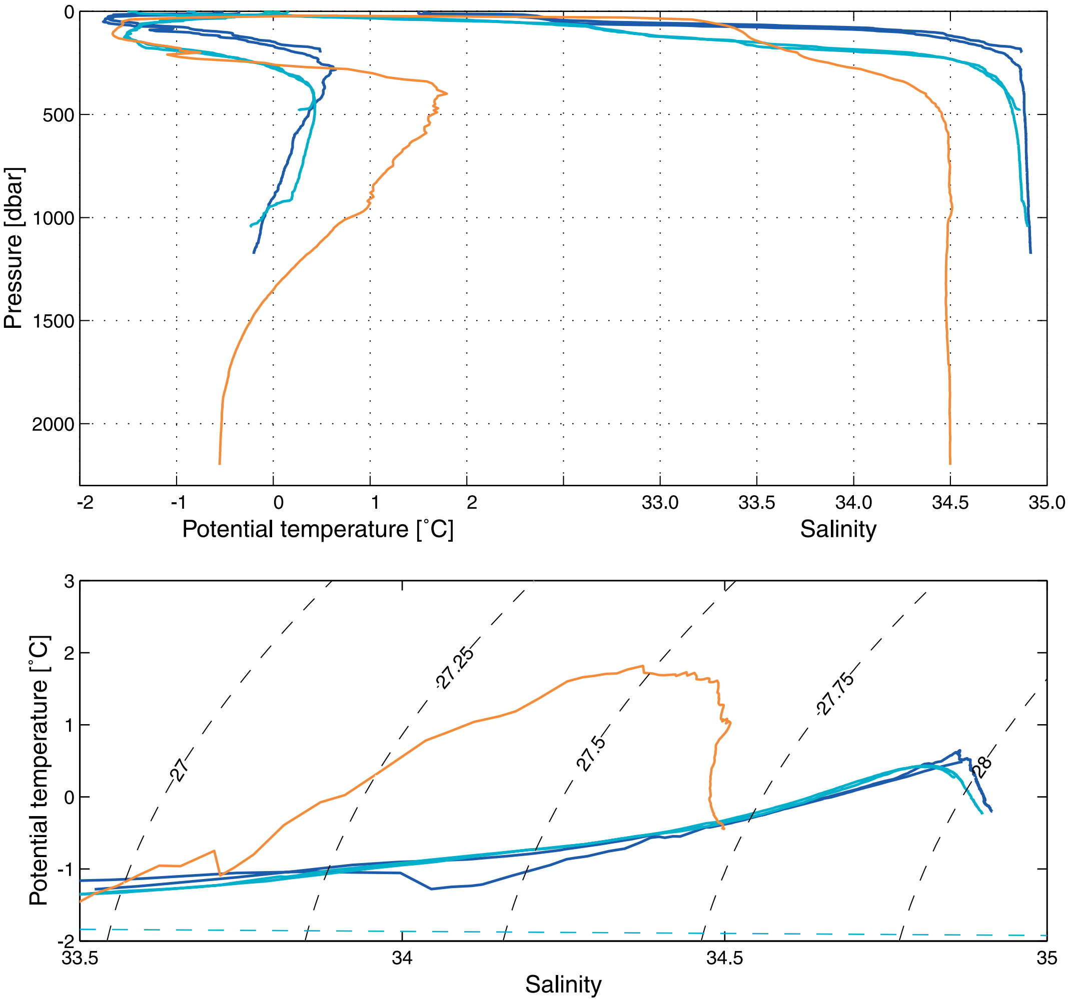
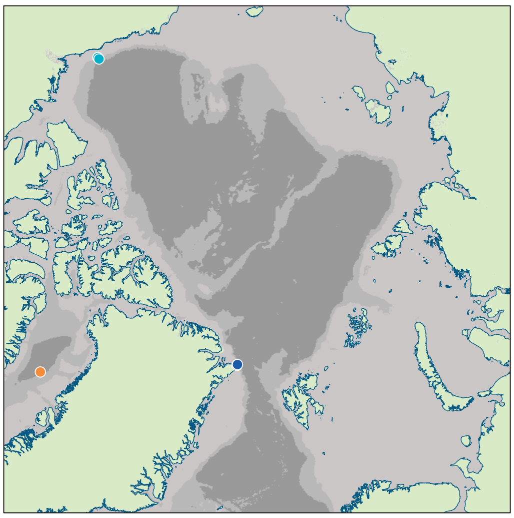


Fig. 11: A comparison between the characteristics of the upper layers in the Beaufort Sea and north of Fram Strait with the characteristics of the deep and bottom waters in Baffin Bay, showing the similarities between the Baffin Bay bottom water and the Barents Sea branch halocline. The upper panel shows potential temperature and salinity profiles and the lower panel ΘS curves from stations Jois/4 and Jois/5 (cyan) in the Beaufort Sea and ArkII/329 and ArkII/330 (blue) northeast of Greenland and BB/133 (dark yellow) in Baffin Bay (see insert map).



The rest is Atlantic water with salinity 35.05 (Blindheim 1989; Aagaard & Carmack 1989). The flow from the Barents Sea to the Norwegian Sea south of Bjørnøya is about 0.8 Sv (Blindheim 1989), and the remaining 2.2 Sv enter the Artic Ocean. With an outflow of 2 Sv through the Canadian Arctic Archipelago, more than half of this volume (i.e., 1.4 Sv) passes through the Canadian Arctic Archipelago, and all freshwater in the Norwegian Coastal Current as well as runoff corresponding almost to two of the three large Siberian rivers Ob, Yenisey and Lena are needed to close the freshwater balance. (Volumes are taken from Aagaard & Carmack 1989, and the freshwater fluxes are computed relative to 35.) An outflow of 1.5 Sv requires almost one river in addition to all of the Norwegian Coastal Current—its volume as well as its freshwater—while an outflow of 1 Sv would demand slightly more than half the volume of the Norwegian Coastal Current (0.4 Sv) and all of its freshwater (Table 3). The freshwater fluxes are large, 2 to 3 times the estimated liquid freshwater export through Fram Strait, and, in the case of 2 Sv, almost as large as the Fram Strait ice export and little less than twice the freshwater inflow through Bering Strait (Aagaard & Carmack 1989). This means that the supply of freshwater to the Arctic Ocean is not enough to balance the export of ice and liquid freshwater through Fram Strait in the case of a large outflow through the Canadian Arctic Archipelago. Considering that at least 0.8 Sv of the Barents Sea branch water enters the Arctic Ocean deeper than the halocline (Loeng et al. 1997; Schauer, Loeng et al. 2002), and that some of the low salinity shelf water leaves the Arctic Ocean through Fram Strait, also the volume balance suggests that the outflow through the Canadian Arctic Archipelago is at the lower end of the range given in Table 3.

Concluding remarks

The concept of two halocline sources derived from Atlantic water is not new. Steele & Boyd (1998) combined the shelf process (Aagaard, Coachman et al. 1981; Jones & Anderson 1986) with the interior basin formation scenario of Rudels, Anderson et al. (1996). They identified two halocline water masses: a Barents–Kara sea shelf derived "advected" halocline and a Fram Strait winter mixed layer "convected" halocline. This distinction was taken up by Woodgate et al. (2001). They combined data obtained from current meters and Seacats moored over the slopes north of the Laptev and East Siberian seas with CTD data from the same region. They concluded that the advected halocline, identified by higher temperature, dominated in the boundary current in the vicinity of the Lomonosov Ridge and that its likely origin was the inflow to the Arctic Ocean over the Barents Sea. The presence of a convective halocline was held possible but less probable.

The choice of the "advective" and "convective" terminology may be inappropriate: it gives the impression that the permanent halocline could be formed locally by convection. This is not the case. If a two layer PML–halocline structure is present, the halocline is, in both scenarios, advected from a distant source: the "advected" halocline from a shallow shelf, the "convected" halocline from another part of the basin or from a deeper shelf. Both scenarios also involve ice formation, brine rejection and haline convection. For the shelf scenario, Aagaard, Coachman et al. (1981) proposed that higher salinities are reached by accumulation of saline water at the bottom of the shelves; in the basin scenario (Rudels, Anderson et al. 1996) the combined effects of ice melt and brine rejection determine the salinity of the winter mixed layer,

Table 3: The salinity of the outflow and the freshwater flux through the Canadian Arctic Archipelago for different total outflows. Freshwater fluxes are computed relative to 35, roughly the salinity of the entering Atlantic water. The freshwater supplied by the 0.6 Sv Pacific water is 0.043 Sv (1353 km^3y^{-1}). PW = Pacific water, BSBW = Barents Sea branch water, FW = freshwater

Outflow vSv	Outflow PW Sv	Salinity PW	Outflow BSBW Sv	Salinity BSBW	Salinity Outflow	Outflow FW Sv	Outflow FW km^3y^{-1}
2	0.6	32.5	1.4	34.0	33.55	0.083	2617
1.5	0.6	32.5	0.9	34.0	33.4	0.069	2176
1	0.6	32.5	0.4	34.0	33.1	0.054	1703

and the convection does not extend to the bottom, only to the pycnocline above the Atlantic layer.

Aagaard & Woodgate (2001) and Woodgate et al. (2001) suggested that ice, advected from the Arctic Ocean and coming into contact with the Atlantic water entering the Barents Sea from the Norwegian Sea, would melt and create a less saline water mass, denser than the PML, which eventually enters the Arctic Ocean and supplies the halocline. Such a process is more akin to the one suggested by Steele et al. (1995) for the Barents Sea polar front, and to the one discussed by Rudels, Anderson et al. (1996) for the Fram Strait inflow, than it is to the shelf scenario proposed by Aagaard, Coachman et al. (1981).

In summary, the contributions from the North Atlantic to the upper layers of the Arctic Ocean have been examined. We conclude that the lower halocline waters in the Amundsen, Makarov and part of the Canada Basin derive from the Fram Strait inflow branch and the winter mixed layer of the Nansen Basin, while the lower halocline water at the Eurasian continental slope, in the southern Canada Basin and along the North American continental slope, originates from the upper part of the Barents Sea branch entering the Arctic Ocean through St. Anna Trough (Fig. 12).

A substantial part of the PML observed in the Amundsen and Makarov Basins is supplied from the Laptev and East Siberian seas, where Atlantic derived low salinity shelf water enters the Arctic Ocean after passing over the shelves of the Barents, Kara and Laptev seas, incorporating the runoff of the Siberian rivers. In spite of the substantial changes in the upper layers of the interior Arctic Ocean reported in recent years, the characteristics of the Atlantic derived halocline waters have been surprisingly stable, although their distribution in the different basins has varied considerably.

The formation of the Barents Sea branch halocline water does not appear to involve either of the two earlier proposed shelf processes, cooling and freshening of Atlantic water brought along deeper canyons onto the shelves from the interior of the Arctic Ocean (Coachman & Barnes 1963), or freezing and brine rejection leading to accumulation of saline water at the bottom of the shelves during winter (Aagaard, Coachman et al. 1981). Both the Fram Strait branch and the Barents Sea branch contributions are created by winter homogenization of less saline, cold waters lying above warm Atlantic water. The Barents Sea branch halocline is distinguished from the Fram Strait branch halocline by higher salinities and higher temperatures that result from the stronger vertical mixing with warmer Atlantic water occurring at the continental slope. The Barents Sea branch halocline thus becomes more like a salinity-stratified thermocline, while the Fram Strait branch remains a halocline. The Barents Sea branch supplies the Atlantic derived water, contributing to the outflow through the Canadian Arctic Archipelago, where it passes through Nares Strait, and it could be a possible source for the deep and bottom waters in Baffin Bay. Apart from this outflow to Baffin Bay, the Barents Sea branch halocline water exits together with the Fram Strait branch halocline water through Fram Strait as a part of the polar surface water of the East Greenland Current.

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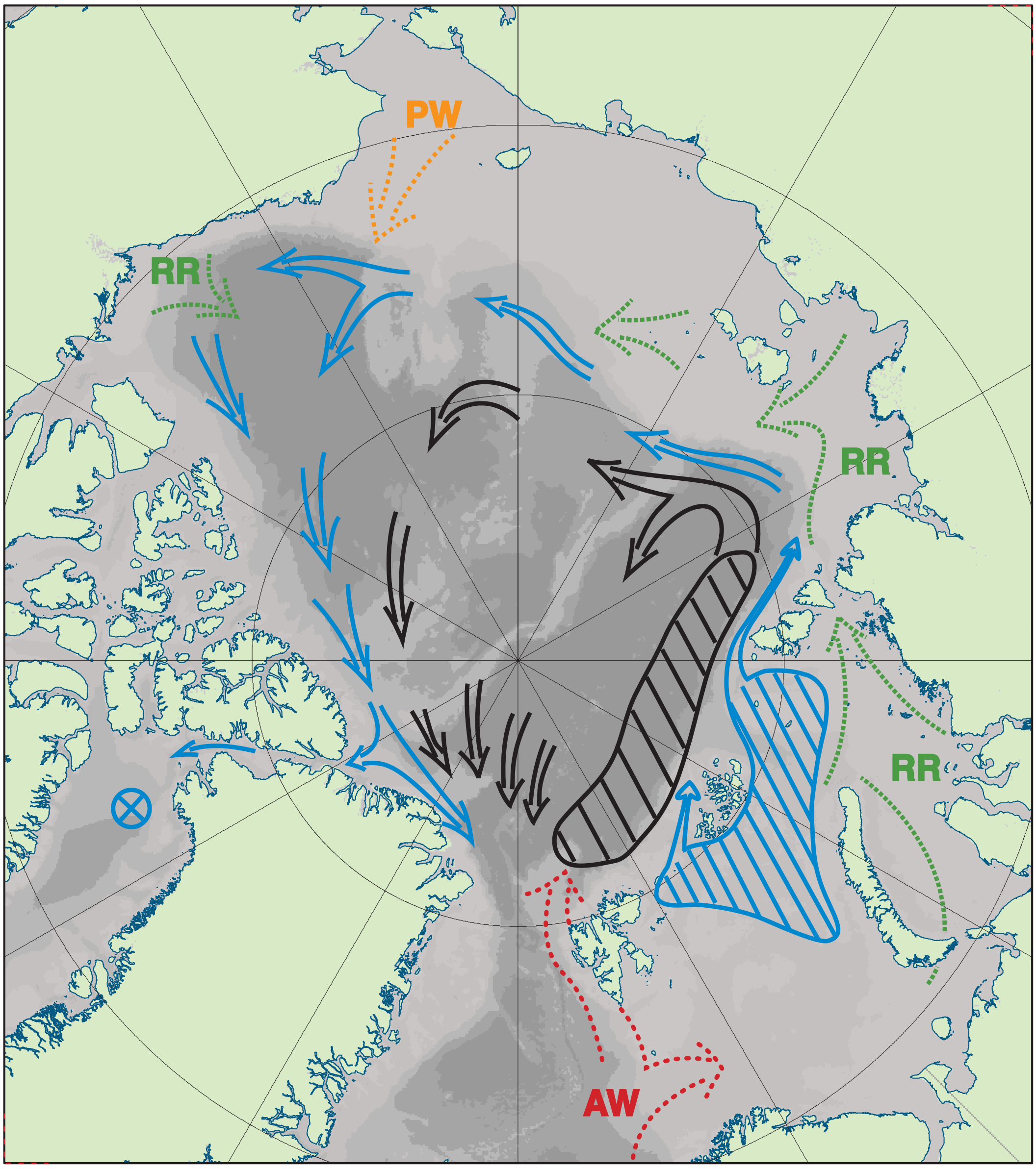

Fig. 12: Schematics showing the inflow of Atlantic Water (AW, red), river runoff (RR green), Pacific inflow (PW, orange), the formation area (black diagonals) and the circulation (black arrows) of the Fram Strait branch halocline, and the formation area (blue diagonals) and circulation (blue arrows) of the Barents Sea branch halocline.

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